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Earthquakes and Seismicity of the Grand Canyon Region

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INTRODUCTION

The Grand Canyon is located in a seismically active part of the earth's crust. This region, shown in Fig. 18.1, also contains a highly complex structure. The only faults represented in Fig. 18.1 are those that have been active in the last four million years (m.y.). The seismicity of this area suggests that some of these faults may be active at present.

Physiographically, the Grand Canyon lies within the Colorado Plateau, which really is a series of plateaus rising eastward across the major boundary faults: Grand Wash, Hurricane, Toroweap, and West Kaibab (Fig. 18.1). The Grand Canyon lies at the southern end of the Intermountain Seismic Belt, which stretches north–south across central Utah. The continuation of this belt of seismicity into Arizona has been the topic of debate; nevertheless, there is no argument that the Grand Canyon region is seismically active. This seismicity may represent a tectonic boundary for the Colorado Plateau, which therefore lies inside of the accepted physiographic boundary at the Grand Wash Cliffs (Brumbaugh 1987).

Little previous work of any detail exists on the seismicity of the Grand Canyon region. Earlier workers surveyed the seismicity of the entire state of Arizona but placed little or no emphasis on the Grand Canyon region (Sturgul and Irwin 1971; DuBois et al. 1982). While studies of the northern part of the state, from the Mogollon Rim to the Utah border, do exist, the Grand Canyon region is not the focus of such works.

Seismologists know that in this century many earthquakes that range in size from Mb6.2 down to Richter Magnitude 3.0 have occurred in the Grand Canyon region (Table 18.1, Fig. 18.1). Events smaller than 3.0 have occurred, but their numbers cannot be estimated reliably because seismologists do not have instrumental coverage of sufficient density to detect these small events unless they occur close to established seismograph stations.

Five events of magnitude 5.0 or greater on the Richter scale have occurred in the Grand Canyon region since 1900. On January 25, 1906 a magnitude 6.2 tremor located 40 km northwest of Flagstaff and 55 km southwest of the rim of the Grand Canyon shook northern Arizona, Utah, and New Mexico. Minor damage occurred at Flagstaff.

On September 24, 1910, an event with a magnitude of 6.0 occurred 50 km southeast of the Grand Canyon along the Mesa Butte fault system. Minor damage was reported at Cedar Wash and at Flagstaff, Arizona.

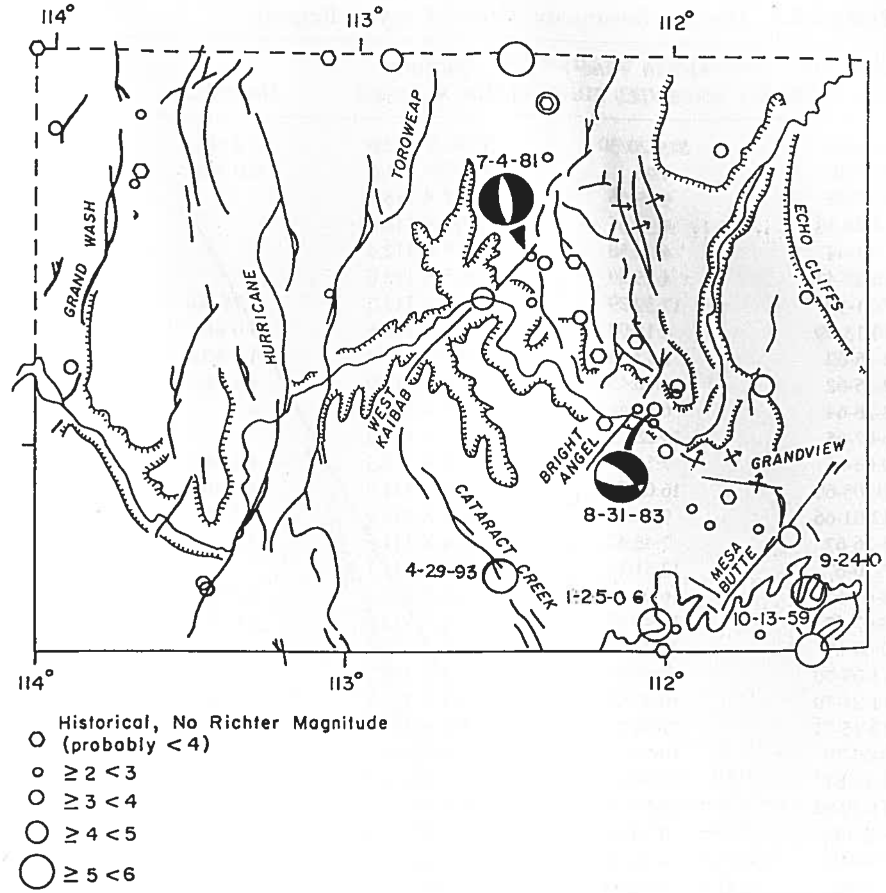

Figure 18.1. Seismicity of the Grand Canyon region. Epicenters are indicated by circles and hexagons. Faults active in the last 4 m.y. are shown by solid lines; monoclines are shown by →. Grand Canyon escarpment = ||||, except where noted (Echo Cliffs). Fault plane solutions (focal mechanisms) are shown for two events: 7/4/81 and 8/31/83.

During 1959, two events of 5.0 or greater on the Richter scale occurred in the Grand Canyon region. On July 21, 1959, Fredonia, Arizona, located on the Arizona–Utah border, was shaken by the largest earthquake since 1912. This event had a calculated Richter magnitude of 5.75 and caused minor damage at Fredonia. It was felt over a 24,400-square-mile (62,500-square-kilometer) area of Arizona and Utah. This event was followed on October 13 by a smaller shock of magnitude 5.0 that was located to the southeast of the Canyon (Fig. 18.1). It was felt strongly in Flagstaff, but no damage was associated with it.

A 1993 tremor caused minor damage over a broad area. On April 29 near Valle, Arizona, a magnitude 5.4 tremor resulted in damage at Valle, Flagstaff, and the Grand Canyon. Power outages were also caused by this event at the Grand Canyon.

The rate of occurence of earthquakes in this century in the Grand Canyon region can be compared to that in the Intermountain Seismic Belt of Utah or,

TABLE 18.1. Historic Seismicity, Grand Canyon Region

Date (Mo-D-Yr)	*Origin Time (UTC)*	*Epicenter (Lat. × Long.)*	*Magnitude*	*Intensity*
1-25-06	203220/30	35.60 × 112.0	6.2 Mb	VII
9-24-10	4:05	35.65 × 111.6	6.0 Mb	VIII
8-17-38	9:08:06	36.7 × 113.7		
12-28-38	4:37:36	37.0 × 114.0		
1-31-44	4:24:58	36.9 × 112.4		
10-27-47	4:15:40	35.5 × 112.0		
7-21-59	17:39:29	37.0 × 112.5	5.75 ML	VII
10-13-59	8:15:00	35.5 × 111.5	5.0 ML	VI
2-15-62	7:12:43	36.9 × 112.4	4.5 ML	V
2-15-62	9:06:45	37.0 × 112.9	4.4 ML	V
8-28-64	6:50:46	37.0 × 113.1		
6-07-65	14:28:01	36.1 × 112.2		
9-03-66	7:53:20	36.5 × 112.3	4.4 Mb	V
10-03-66	16:03:50	35.8 × 111.6	4.4 Mb	V
12-01-66	9:20:40	36.2 × 113.9	3.7 Mb	IV
5-26-67	7:48:42	36.4 × 111.6	3.1	II
7-20-67	13:51:10	36.3 × 112.1		
8-07-67	16:24:44	36.5 × 112.4	3.9 ML	IV
8-07-67	16:40:32	36.4 × 112.6	4.0 ML	IV
9-04-67	23:27:44	36.2 × 111.7	4.6 ML	V
11-05-70	9:45:57	36.3 × 112.2		
11-24-70	16:47:56	36.4 × 112.3	3.0 ML	II
12-15-71	12:58:14	36.8 × 111.8	3.0 ML	II
8-05-79	19:10:15	36.8 × 114.0	3.7 ML	IV
1-12-81	8:59:13	35.7 × 113.5	3.5 ML	III
11-19-82	20:57:34	36.0 × 112.0	3.0 ML	II
8-31-83	8:10:09	36.1 × 112.0	3.3	III
7-18-84	14:29:33	36.2 × 112.0	3.0 ML	II
9-6-88	9:44:00	36.0 × 112.2	3.0 ML	
9-7-88	1:17:40	36.0 × 112.1	3.1 ML	
9-7-88	3:22:07	36.0 × 112.2	3.0 ML	
3-5-89	4:40:32	36.0 × 112.1	4.0 ML	
3-5-89	7:12:57	36.0 × 112.1	4.0 ML	
11-28-89	18:37:32	36.1 × 112.2	3.0 ML	
4-26-91	13:08:30	36.6 × 112.6	4.0 ML	
8-22-91	16:41:01	36.0 × 112.1	3.0 ML	
3-13-92	11:28:36	36.0 × 112.2	3.9 ML	
3-14-92	5:12:08	36.0 × 112.2	4.0 ML	
3-14-92	5:13:36	35.9 × 112.2	4.5 ML	
5-6-92	1:40:58	36.0 × 112.2	3.0 ML	
5-20-92	21:46:05	36.0 × 112.2	3.1 ML	
7-5-92	18:18:33	35.9 × 112.3	3.9 ML	
4-25-93	9:29:45	35.7 × 112.2	4.9 Mb	
4-29-93	8:21:00	35.7 × 112.2	5.4 Mb	

on a broader scale, that of the Western Mountain Region (Fig. 18.2). It should be understood that the time span for which data are available is rather small for the moderate-to-large events (MM VI–IX). Therefore, our information may not be as reflective of true rates of recurrence as we would like. Nevertheless, despite effects caused by periodicity, it is clear that recurrence rates in northern

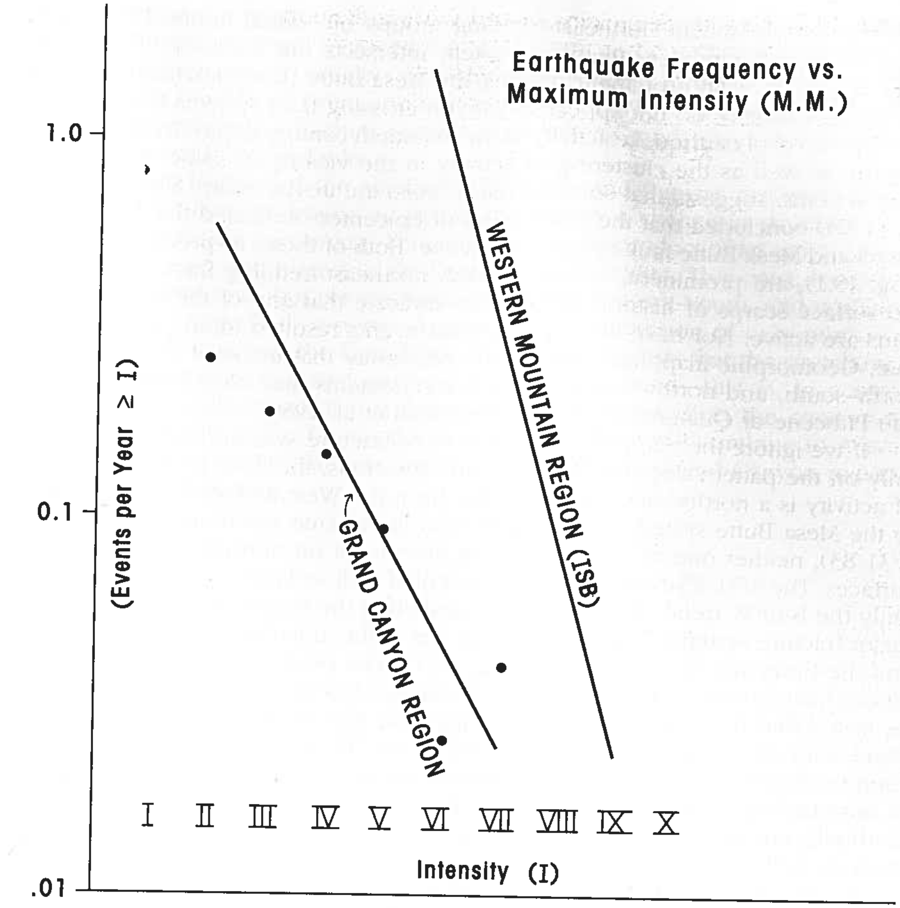

FIGURE 18.2. Earthquake frequency for the Grand Canyon region compared to the Intermountain Seismic Belt (ISB). Recurrence curves are represented by a solid line for each region. Western Mountain Region curve is from Podolny and Cooper (1974).

Arizona are lower, on the average, than those for the Intermountain Seismic Belt. The rate varies from a maximum north of the Grand Canyon, which is comparable to that of the Intermountain Seismic Belt in Utah (Kruger-Kneupfer et al. 1985), to much lower rates south of the Canyon. The slope of the recurrence curve for the Grand Canyon data (Fig. 18.2) suggests the occurrence of an MM VII (5.5–6.0) event twice every 50 years or so.

TECTONICS

The Grand Canyon region of the southern Colorado Plateau is dominated structurally by high-angle normal faults that cluster into large, northeast, north–south, and northwest-trending fracture systems. Several of these fracture systems are well-exposed where they cross the Grand Canyon (Fig. 18.1). Crossing the canyon from west to east are Grand Wash, Hurricane, and Toroweap faults, all predominantly north–south trending. The West Kaibab and Bright Angel fracture systems are prominent northeast-trending groups of normal faults. The northwest-trending Grandview–Phantom system intersects the walls of the eastern Grand Canyon. Two prominent systems, the Mesa Butte (northeast) and Cataract Creek (northwest), do not appear as canyon-crossing fault systems (Fig. 18.1).

The level of earthquake activity in the twentieth century in the Grand Canyon region, as well as the clustering of activity in the vicinity of some of the fracture systems, suggests that some of these faults are active today. Shoemaker et al. (1974) concluded that the distribution of epicenters indicated that the Bright Angel and Mesa Butte fault systems are active. Both of these, as previously noted (Fig. 18.1), are prominent, well-expressed, northeast-trending fracture systems. No surface scarps of historic age exist to indicate that any of the fracture systems are active. Nor have the 5.0 or greater events resulted in any surface rupture. Geomorphic mapping suggests the possibility that many of the northeast, north–south, and northwest-trending fracture systems may have been active in late Pliocene or Quaternary time (Scarborough et al. 1986).

If we ignore the mapped surface traces of fracture systems and concentrate only on the pattern displayed by epicenter locations, the most prominent trend of activity is a northwest trend stretching from the West Kaibab fracture system to the Mesa Butte system. Furthermore, two fault plane solutions exist (7/4/81, 8/31/83), neither one of which indicates movement on northeast-trending fault surfaces. The 8/31/83 event shows two potential fault surfaces, N40°W and N86°E. Only the N40°W trend comes close to paralleling the trends of any of the three major fracture systems. The 7/4/81 event lies at the junction of the West Kaibab and the Crazy Jug fracture systems (Fig. 18.1). The potential fault surface orientations have trends of N24°W and N30°W (Kruger-Kneupfer et al. 1985). It should be noted that the solution is a composite one; that is, it combines the 7/4/81 event with others close to it in time and space. There are those who feel that composite solutions are less reliable than single event ones. Finally, examination of the distribution of epicenters of just the largest events ($M > 4.0$), which undoubtedly are controlled by tectonic stresses, clearly shows a northwestern alignment as well.

A tectonic boundary may be defined by changes in tectonic elements. This includes parameters like structural style, crustal thickness, and seismicity. The concentrated band of seismicity in the Grand Canyon region marks the locus of a change in crustal thickness and a change in structural style. The West Kaibab fault zone marks the eastern limit of Basin- and Range-type faults on the southern Colorado Plateau. Crustal thicknesses in the western Grand Canyon region are more like those found beneath the Basin and Range terrane.

Brumbaugh (1987) has suggested that the belt of concentrated seismicity in the Grand Canyon region represents the present-day tectonic boundary of the Colorado Plateau—a boundary, furthermore, that has migrated with time, as suggested by earlier workers (Best and Hamblin 1978). This migration has expanded the area of the Basin and Range from the southwest and encroached into the Colorado Plateau.

SUMMARY

The seismicity of the Grand Canyon region of northern Arizona is contiguous with the Intermountain Seismic Belt of Utah. This, plus a similar recurrence rate north of the canyon (and a common style of faulting), suggests that the seismicity of the Grand Canyon region represents an extension of the Intermountain Seismic Belt of Utah. An analysis of the distribution of epicenters (Fig. 18.1)

indicates that the majority of the events are concentrated in a rather narrow, northwest-trending band between the Mesa Butte and the West Kaibab fracture systems. This trend of activity, plus the northwestern trend of potential fault planes from the 7/4/81 event and one from the 8/31/83 event, suggest that it is the northwest-trending fracture systems, such as the Grandview–Phantom, which contain presently active faults.

It seems likely that the seismicity of the Grand Canyon region is an expression of a tectonic boundary that is marked by a change in structural style as well as in crustal thickening. This seismicity is the product of an active stretching or extension of the crust in the Grand Canyon region. This crust, moreover, is not homogeneous, but pre-fractured in northeast, north–south, and northwest directions by large fault systems. A preferential reactivation of only northwest-trending faults, such as the Grandview–Phantom, suggests that the stretching of the crust must be occurring perpendicular to this trend—that is, in a north-east–southwest direction. If this extension continues to affect the crust of the Colorado Plateau in the future, the tectonic boundary will continue to migrate into the interior of the plateau—eventually leaving the Grand Canyon region a seismically quiet area.

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Holocene Terraces, Sand Dunes, and Debris Fans along the Colorado River in Grand Canyon

Kelly J. Burke, Helen C. Fairley, Richard Hereford, and Kathryn S. Thompson

INTRODUCTION

River terraces, sand dunes, and debris fans at the mouths of tributaries are the main elements of Holocene surficial geology of the Colorado River in Grand Canyon (Fig. 19.1), as shown by large-scale geologic maps. These deposits partly occupy the river corridor, the ribbon of water and land between steep bedrock walls of Grand Canyon. Because the deposits formed only during the past several thousand years, they contain little information about the origin of Grand Canyon. However, these deposits record millennia of interactions between river and tributaries in the natural regimen, before Glen Canyon Dam was built.

The pre-dam flood regimen resulted from snowmelt in the Rocky Mountains. The river flooded in late spring and early summer; flood waters, laden with large quantities of suspended sediment, moved boulders along the channel and deposited silt and sand forming high-level terraces. Over time, perhaps several millennia, the largest floods attained a relatively open channel without waterfall-like obstructions, although rapids remain as evidence that obstructions were not removed completely. Sediment deposited in the main channel by tributaries is capable of producing unnavigable obstructions (for example, see Webb et al. 1996, Fig. 35), but such debris was in time removed by the unregulated Colorado River.

With closure of Glen Canyon Dam in 1963, major floods were eliminated and the sediment load was greatly diminished (Andrews 1991). The ability of floods to remove unnavigable obstructions is limited, and sand is no longer deposited in high-level sites. The Holocene surficial geology discussed here shows how the channel system evolved under natural conditions, which in turn permits speculation about how regulated streamflow influences evolution of the system.

The high-level terraces are now eroding through incision by small ephemeral streams, revealing archeologic remains (Hereford et al. 1993). This cultural material is locally abundant along the Colorado River in association with terraces and debris fans that prehistoric people used for camping, agriculture, and construction of masonry structures (Fairley et al. 1994). Most of the remains are affiliated with the Pueblo II Anasazi, dating between 1000 A.D. and 1150 A.D., al-

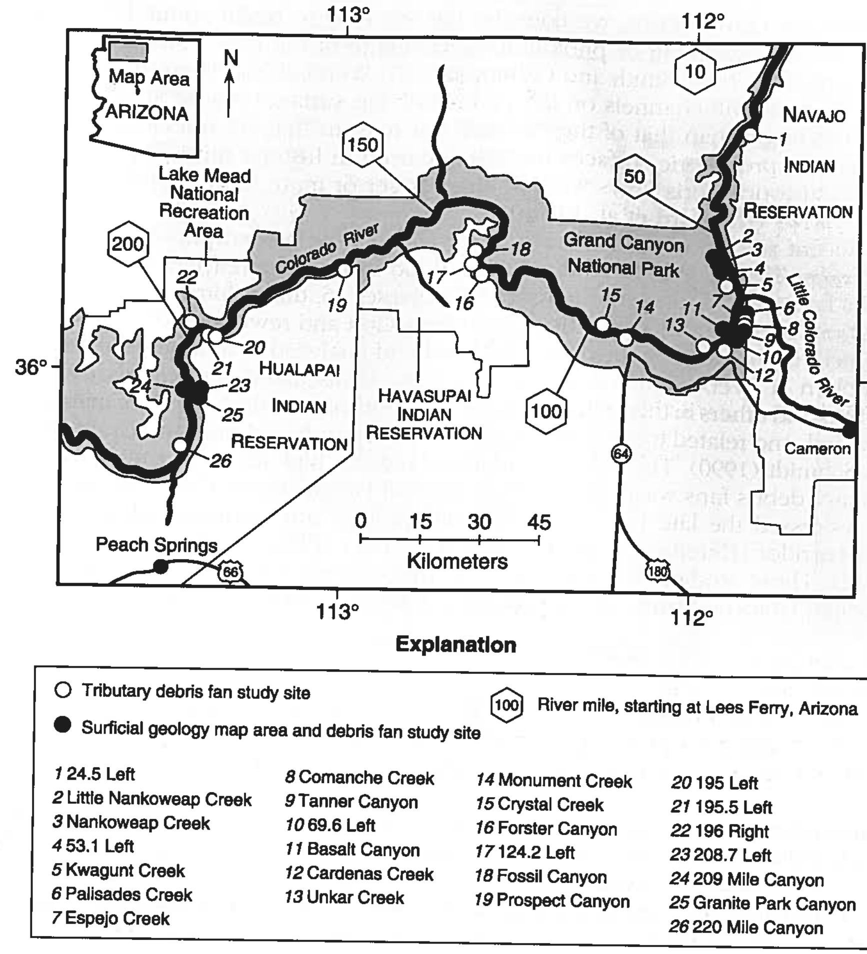


Figure 19.1. Study sites, Colorado River, Grand Canyon National Park, Arizona.

though cultural artifacts range in age from about 800 B.C. to the early twentieth century (Jones 1986; Altschul and Fairley 1989). These same deposits are now used for recreational purposes, mainly hiking or rafting through Grand Canyon. The spectacular scenery of Grand Canyon and whitewater rapids near the eroded banks of many debris fans attract 22,000 rafters annually (Stevens 1990) who are exhilarated by the powerful waves and swift currents. In addition, most of the sand deposited by the Colorado River accumulates near debris fans (Schmidt and Graf 1990), and these deposits are popular camp sites. The sand also forms the substrate for riparian vegetation, which in turn supports the diverse ecosystem of the Colorado River (Carothers and Brown 1991).

Pre-dam terraces, debris fans, and sand dunes discussed in this chapter occupy most of the river corridor in terms of surface area and sediment volume. Post-dam alluvial deposits are substantially reduced in area and height above the river compared with pre-dam terraces. The disparity in volume between the post- and pre-dam alluvial deposits results directly from reduced streamflow. Likewise, the historic-age debris fans are substantially smaller than their prehistoric counterparts. For convenience, we consider historic time to begin about 1870–1890 A.D. with the beginning of photographic coverage of the river corridor (Turner and Karpiscak, 1980; Smith and Crampton 1987; Webb 1996). These historic-age deposits partly fill channels on the prehistoric fan surfaces whose area is about six times larger than that of the channel. For reasons that are not clear, deposition on the prehistoric surfaces has not occurred in historic time. This suggests that prehistoric debris flows were at times larger or more frequent than historic debris flows (Hereford et al. 1996a).

Recent studies of the Colorado River address the geomorphology of active processes. The hydraulics of rapids and flood-related geomorphic evolution of debris fans were studied by Kieffer (1985; Chapter 16, this volume). The process, initiation, frequency of debris flow, and deposition and reworking of debris-flow sediment in the main channel are described and analyzed in a number of papers by Robert H. Webb and colleagues. This work is discussed and summarized by Griffiths and others in this volume. Alluvial deposits of post-dam age were mapped, classified, and related to hydraulic conditions at rapids by Schmidt and Graf (1990) and Schmidt (1990). The close association between high-level river terraces and tributary debris fans was noted by Howard and Dolan (1981). Other recent studies addressed the late Quaternary surficial geology and geomorphology of the river corridor (Hereford 1996; Hereford et al. 1993, 1996a, 1997, 1997a, b, 2001, 2001a). These studies, the main topic of this chapter, identified and dated the principal Holocene surficial geologic elements of the river corridor.

GENERALIZED GEOLOGIC SETTING, CLASSIFICATION, AND AGE OF LATE QUATERNARY DEPOSITS

Holocene deposits along the Colorado River developed in a geologic framework of late Pleistocene gravel, talus, and bedrock. These deposits are best developed where the canyon at river level is wide. A minimum width of 200–400 m evidently provides the conditions necessary for deposition and preservation of sediment. For the most part, deposits are not present or are poorly developed in the upper and lower Granite Gorges where the river corridor is narrow. Classification and correlation of the late Quaternary deposits along the Colorado River in Grand Canyon are shown in Figure 19.2.

In wide reaches of the river corridor, the Holocene deposits are usually bounded by bedrock. Locally, however, the Holocene deposits are bounded at the channel margin by Pleistocene terrace-forming gravel. Two terrace levels (gvy and gvo of Fig. 19.2) form the margins of the river corridor at or near river level. The terraces are well developed near Lees Ferry, Nankoweap Rapids, Furnace Flats, and Granite Park (Fig. 19.1, river mile 0, 52, 65–74, and 208–209, respectively). The base of the younger gravel is below or slightly above river level, and the base of the older gravel is typically near river level to 20–30 m above the river. The deposits are weakly to moderately consolidated and up to 30 m thick. They consist of moderately well-rounded boulder-size clasts of Paleozoic limestone and sandstone in Grand Canyon and Mesozoic sandstone in the Lees Ferry area. Distinctive well-rounded pebbles of porphyritic rock occur sparingly in the gravels. These pebbles were derived from distant igneous sources in the laccolithic mountains of the Colorado Plateau and San Juan Mountains. Medium- to very coarse-grained sand is present in the gravel matrix or interbedded with the gravel.

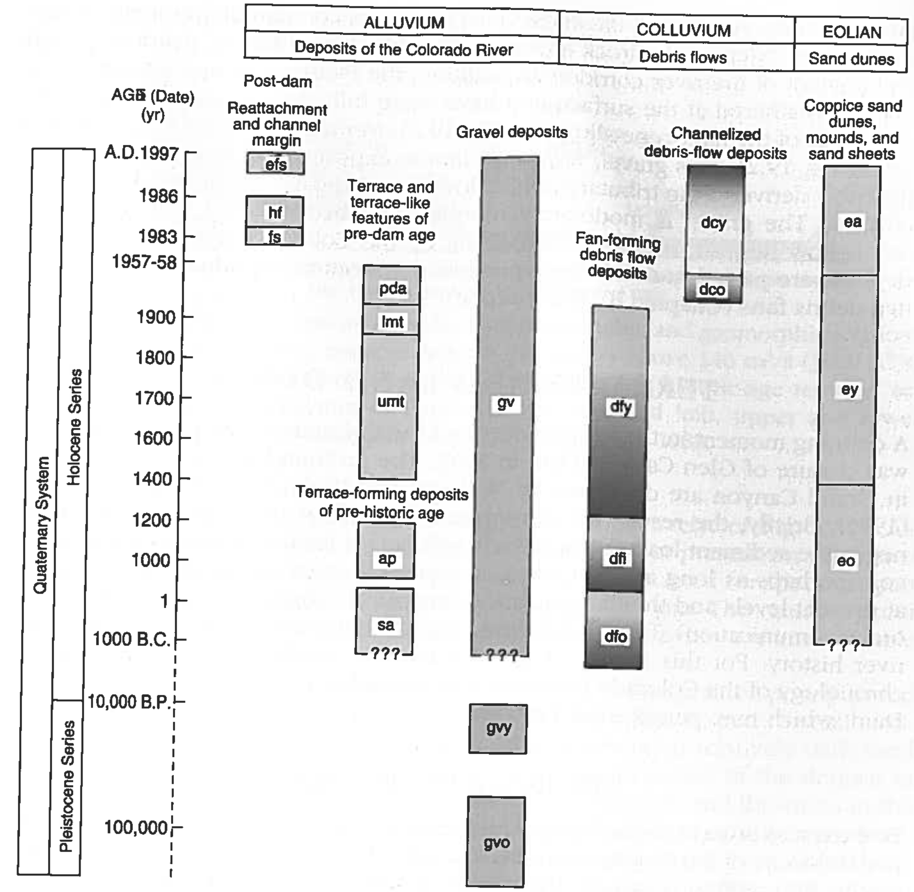


FIGURE 19.2. Correlation chart and classification of late Quaternary deposits along Colorado River. Symbols explained in text.

The gravels are two ancient channel-fill deposits of the Colorado River dating from the Pleistocene (Hereford et al. 1993; Hereford 1996). The contact between gravel and bedrock is strongly concave, and the contact slopes steeply toward the river. This contact is the margin of an ancient river channel. The Pleistocene age of the gravels is based on their elevated position above Holocene deposits and by correlation with dated late Pleistocene deposits in eastern Grand Canyon (Machette and Rosholt 1991). The younger gravel (gvy of Fig. 19.2) probably corresponds with river levels one to three, and the older gravel (gvo of Fig. 19.2) correlates with river levels four and five of Machette and Rosholt (1991). Levels one to three range from 5 ± 5 to 40 ± 24 ka; levels four and five range from 75 ± 15 to 150 ± 30 ka. In the Nankoweap Rapids area, the older gravel is substantially older than the latter age. These ages are roughly comparable to those obtained by Lucchitta et al. (1995) for gravel deposits in similar topographic positions in eastern Grand Canyon.

The younger gravel is considered late Pleistocene rather than Holocene. The Pleistocene age is inferred from the coarse grain size (gravel compared with fine-grained sand), substantial thickness (tens of meters compared with meters), and evidence of extensive bedrock erosion that resulted in widening, deepening, and realignment of the river corridor. In addition, the Pleistocene deposits are more deeply weathered at the surface and have more fully developed soil horizons.

Much of the Holocene alluvium (Fig. 19.2) overlies an unconsolidated gravel (gv of Fig. 19.2). This gravel, with a maximum exposed thickness of 4–5 m, was probably derived from tributary debris flows that deposited sediment in the main channel. The gravel is moderately rounded and better sorted than a primary debris-flow deposit, indicating reworking by the Colorado River. These gravel deposits are probably debris bars, material that accumulates downstream of active debris fans (Chapter 20, this volume).

TERRACES AND SAND DUNES

A defining moment in the late Holocene of the Colorado River in Grand Canyon was closure of Glen Canyon Dam in 1963. The profound affects on streamflow in Grand Canyon are discussed by Williams and Wolman (1984) and Andrews (1991). Briefly, the results are diminished streamflow and reduced, essentially negligible sediment load. These effects will persist for the life span of the reservoir, perhaps as long as 1200 years, assuming average sediment input remains at present levels and the dam retains its structural integrity (E.D. Andrews 1997, oral communication). Twelve hundred years is a substantial portion of Holocene-river history. For this reason, it is convenient to divide the Holocene alluvial chronology of the Colorado River into two episodes: pre- and post-Glen Canyon Dam, which may persist until 3200 A.D.

Alluvium of Pre-dam Age

Five terraces or terrace-like features and related alluvial deposits constitute the upper Holocene of the Colorado River in Grand Canyon (Hereford et al. 1993, 1996a, 1997a, 2001, 2001a; Hereford 1996). From oldest to youngest, the deposits are the striped alluvium, alluvium of Pueblo II age, alluvium of the upper and lower mesquite terraces, and the pre-dam alluvium (sa, ap, umt, lmt, and pda, respectively of Fig. 19.2). These deposits form discontinuous terraces and terrace-like features upstream, downstream, and encircling Holocene debris fans. Figure 19.3 illustrates schematically the geomorphology and geology of the terraces, the height of the terraces above the river, and the position of the terraces relative to historic streamflow levels. Each terrace occupies a distinctive topographic position with the oldest terrace highest and farthest from the river. The full suite of terraces is not present at any one location, nor are the terraces paired across the river. Terraces are unpaired because the river usually flows between bedrock or talus on one bank while the other bank is alluvium, tributary debris fan, or both. In addition, the position of the active channel relative to the older terraces varies considerably. Locally, the channel can be adjacent to any of the older terraces.

The terraces have inset geomorphic relations such that younger terraces are topographically lower than older terraces; however, the units partly overlap in the subsurface (Fig. 19.3) as indicated by the stratigraphic relations among the deposits. The area of overlap between units is an erosional hiatus that is difficult to detect (see Figs. 3 and 4 in Hereford 1996) and is easily overlooked (Lucchitta et al. 1995) without excavation, analysis, and dating.

Striped Alluvium and Alluvium of Pueblo II Age These deposits form the highest terraces of the river corridor, and they are the oldest Holocene alluvial

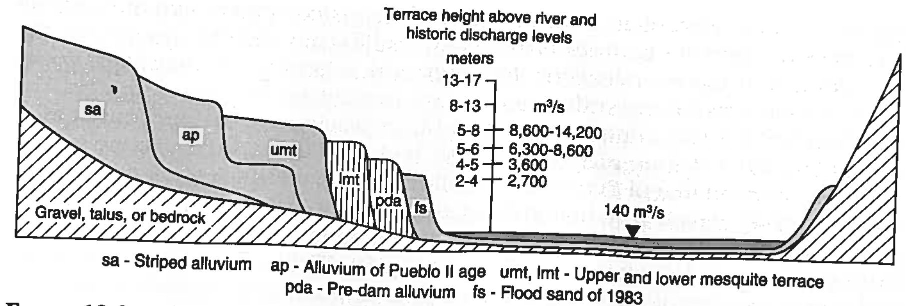


Figure 19.3. Schematic cross section showing stratigraphic and geomorphic relations of late Holocene alluvium, range of heights of terraces above 140 m^3/s (5000 ft^3/s) discharge level, and maximum discharges levels affecting historic-age terraces. sa, striped alluvium; ap, alluvium of Pueblo II age; umt and lmb, upper and lower mesquite terrace; pda, pre-dam alluvium; fs, flood sand of 1983.

deposits adjacent to the river (units sa and ap, respectively, of Fig. 19.2). Although driftwood is present on younger terraces, it is not present on these two oldest terraces. Prehistoric archeologic remains are associated for the most part with the alluvium of Pueblo II age and the striped alluvium. The remains occur on or near the surface or buried within the deposits, where they are undetectable unless exposed by erosion.

The striped alluvium consists of light-colored very fine-grained poorly sorted sand interbedded with thin beds (5–15 cm) of dark-colored sand to small-cobble gravel derived from nearby hillslopes. This interbedded relatively dark sand and gravel imparts the distinct "stripes" that are characteristic of the deposit in eastern Grand Canyon. The gravel beds increase in number and thickness in the direction of the nearby bedrock hillslopes; the gravel resulted mainly from sheetwash. The sand beds increase in number in the direction of the river. These beds are largely of fluvial origin; they contain relatively high quantities of silt and sand, and fluvial sedimentary structures are present locally, although most of the deposits lack sedimentary structures.

The alluvium of Pueblo II age derives its name from the locally abundant archeologic material of Pueblo II affinity, although diagnostic Pueblo I material is present near the base and early Pueblo III ceramic material is present on the terrace. The archeologic material consists of potsherds, flakes, rock alignments, upright slabs, walls, and other artifacts or features. This deposit consists mainly of light-colored poorly sorted very fine-grained sand of fluvial origin interbedded locally with moderately well-sorted fine-grained sand of possible eolian origin. Interbedded dark-colored sand and gravel beds forming stripes are also present, but they are not as conspicuous as those in the striped alluvium. In places, the alluvium of Pueblo II age disconformably overlies the striped alluvium. The contact between the alluviums is an eroded surface with up to one meter of relief locally, and stratification in the striped alluvium is truncated at this surface. Thus, the alluvium of Pueblo II age is stratigraphically distinguishable from the striped alluvium, although in places there is little or no topographic separation between the alluviums at the surface.

Alluvium of the Upper and Lower Mesquite Terraces The mesquite terraces (units umt and lmt of Fig. 19.2) range from narrow, discontinuous surfaces or scoured zones to well-developed readily identifiable terraces; they are topographically below the terraces of the striped alluvium and alluvium of Pueblo II age (Fig. 19.3). The name stems from the abundant western honey mesquite (*Prosopis glandulosa* var. *torreyana*; Turnet and Karpiscak 1980) present on the terraces. By and large, relatively large mesquite is present on the older, upper mesquite terrace, whereas smaller trees are present on the lower terrace. The mesquite terraces are deposits in the "old high-water zone" of previous studies (Johnson 1991; Carothers and Brown 1991; Webb 1996, Fig. 6.1). Like the striped alluvium and alluvium of Pueblo II age, alluvia of the mesquite terraces are light-colored poorly sorted very fine-grained silty sand of Colorado River origin.

Photographs of the Palisades Creek area taken in January 1890 by Robert B. Stanton (Smith and Crampton 1987, pp. 149–159; Webb 1996) show the upper and lower mesquite terraces. At that time, the upper mesquite terrace was vegetated and appeared to not have been flooded recently. In contrast, the lower mesquite terrace was sparsely vegetated and consisted of high-albedo sand with dark elongated objects interpreted as driftwood. These differences suggest that in 1890 the lower terrace had been recently flooded whereas the upper terrace was inactive.

Driftwood in this area associated with the lower mesquite terrace contains less than 5 percent of milled and cut wood. The paucity of milled and cut wood implies that the driftwood is of early historic age, therefore, it was probably deposited by the flood of July 1884, the largest flood of the systematic record with estimated peak discharge of 8500 m^3/s (300,000 ft^3/s). In contrast, although rare, older and higher driftwood associated with the upper mesquite terrace consists mainly of beaver-cut cottonwood without milled and cut wood (Hereford et al. 1997).

The two terraces formed during the larger floods of the late prehistoric to early historic period. Judging from its relatively high topographic position above the lower mesquite terrace, sand of the upper mesquite terrace probably accumulated during the largest floods of the post-Anasazi era. The lower mesquite terrace, which accommodated floods with lower stage than those affecting the upper terrace, quite likely began to develop with the flood of July 1884.

Age of the Striped Alluvium, Alluvium of Pueblo II Age, and Alluvium of the Upper Mesquite Terrace The ages of the striped alluvium, alluvium of Pueblo II age, and alluvium of the upper mesquite terrace in eastern Grand Canyon are constrained by radiocarbon dates and archeologic material (Fig. 19.4). Organic material collected from the three units was dated using the radiocarbon method. The sampling procedure, limitations of the radiocarbon dates, and stratigraphic context of the carbon are in Hereford et al. (1993, 1996a) and Hereford (1996). Deposition of the striped alluvium in eastern Grand Canyon began before 800 B.C. and could have lasted until about 300 A.D. Older dates have been obtained from the unit near Granite Park where deposition was ongoing by about 1300 B.C. (Hereford et al. 2001). In upper Marble Canyon, O'Conner et al. (1994) dated the base of a sequence that may be equivalent to the striped alluvium at 2500 B.C. This age range of 2500–1300 B.C. to 300 A.D. is consistent with the aceramic character of the archeologic remains in the alluvium. This indicates that the deposit predates the Pueblo era and is most likely of late Archaic to Basketmaker III age.

Deposition of the alluvium of Pueblo II age probably began by about 700 A.D. This is supported by the presence of Pueblo I ceramics near the base of the alluvium in the Upper Unkar area (Fig. 19.1) that date from 800 to 900 A.D. Late Pueblo II to early Pueblo III ceramics are present on the surface of the alluvium. The age and stratigraphic context of these younger ceramics suggest that deposition of the alluvium ended between 1150 A.D. and 1200 A.D., which was probably coincident with abandonment of the area by the Anasazi. The period of erosion and nondeposition (Fig. 19.4) between the two alluvia lasted about 400 years, from 300 A.D. to 700 A.D.

The alluvium of the upper mesquite terrace is younger than about 1200 A.D. and was deposited after Anasazi occupation of Grand Canyon, based on the lack of Anasazi ceramics and other material in the alluvium. Deposition of the alluvium forming the upper mesquite terrace must have begun after 1150–1200 A.D., and radiocarbon ages (Fig. 19.4) suggest that deposition could have begun as late as 1400 A.D. Thus, the erosional hiatus probably lasted 200 years, between 1200 A.D. and 1400 A.D.

The causes of this alternating fluvial deposition and erosion in Grand Canyon during the late Holocene are not well understood. Water in the Colorado River is derived from snowmelt in the headwaters of the Rocky Mountains, whereas the sediment load is mainly from tributaries of the Colorado Plateau (Andrews 1991). Thus, climate of either region could influence alluviation in Grand Canyon in ways that are not immediately apparent. Nevertheless, the alluvial chronology of the Colorado River in Grand Canyon correlates broadly with a late Holocene

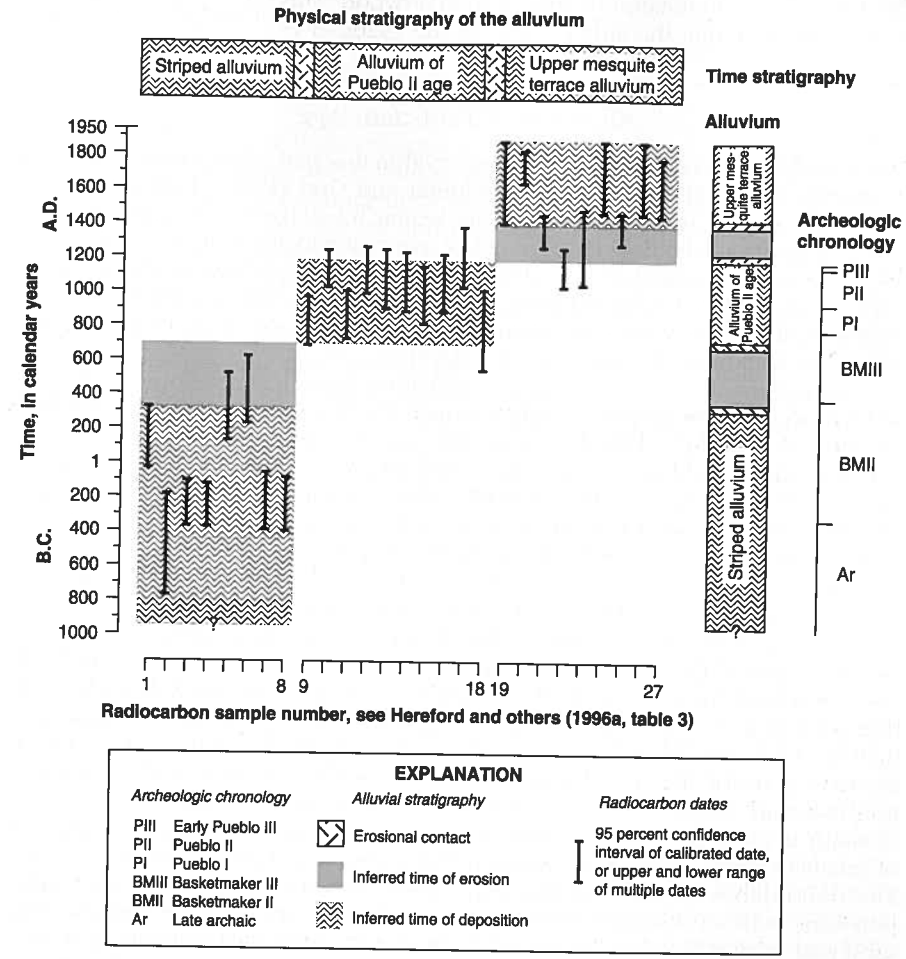


FIGURE 19.4. Radiocarbon dates calibrated to calendar years arranged by stratigraphic unit showing physical and time stratigraphy of late Holocene alluvium, along with archeologic chronology. [Modified from Hereford et al. (1996a, Fig. 19.9). Archeologic chronology of Altschul and Fairley (1989).]

chronology of the southern Colorado Plateau developed by Karlstrom (1988) and elaborated by Dean (1988, p. 129). The alluvium of Pueblo II age probably correlates with the upper Tsegi Formation, and the alluvium of the upper mesquite terrace correlates roughly with the Naha Formation, both of northeast Arizona and southern Utah (Hack 1942; Cooley 1962). Erosion in Grand Canyon around 1200–1400 A.D. coincides with widespread stream entrenchment on the southern Colorado Plateau at about this time (Hereford et al. 1996a).

Pre-dam Alluvium The pre-dam alluvium and flood debris (pda of Fig. 19.2) form a terrace, or in places a scoured surface, that is topographically beneath the lower mesquite terrace (Fig. 19.3). The dominant vegetation at this level is saltcedar (*Tamarix chinensis* Lour.; Turnet and Karpiscak 1980). These trees are normally large, mature, and partially buried in the alluvium. Tree-ring dates obtained from two trees at the Palisades Creek area (Fig. 19.1) indicate germination in 1937 and 1951. Flood debris contains artifacts dating from the mid-1930s to mid-1950s, and the driftwood is dominated by milled and cut wood. This terrace and related deposits formed during the larger floods of the 1920s to 1957–1958.

Alluvium of Post-dam Age

The post-dam alluvial deposits lie entirely within that part of the channel affected by operation of Glen Canyon Dam. Schmidt and Graf (1990) classify the post-dam alluvium as channel-margin deposits, reattachment bars, and separation bars. These deposits accumulate in areas of low current velocity related to changes in the width of the channel or in hydraulic roughness of the channel margin. They have formed at several topographic levels controlled by the discharge rate prevailing during deposition. River runners refer to all of these deposits as beaches, which are well-liked as camp sites for raft trips.

Reattachment bars result from recirculating flow that develops in the distinctive, arcuate topographic setting downstream of a large channel constriction, usually a debris fan. Downstream of the constriction, the flow separates from the main current and moves upstream, rejoining the main current at the head of the recirculation zone or eddy (Schmidt 1990). Several studies have addressed the sedimentology of reattachment bars and their relation to regulated streamflow (Schmidt 1990; Schmidt and Graf 1990; Rubin et al. 1990).

Aside from the deposits of the active channel, the most conspicuous post-dam deposits are those of the 1983 flood, a largely unplanned flood release, and those dating from 1984 to 1986. Earlier deposits dating from 1963 to 1983 were probably eroded during the floods and prolonged high releases of 1983 to 1986. The flood sand (unit fs of Fig. 19.2) is distinctively light-colored well-sorted very fine-grained sand with clay and silt content less than 5 percent. Because it is light-colored and relatively well sorted, the flood sand resembles eolian sand in the river corridor. The flood sand is usually about 1–2 m thick, and its distribution is locally spotty; where the sand is absent, a flood line showing evidence of scour is present. The scour line is marked by flood debris and an alignment of relatively small, submature saltcedar that germinated between 1970 and 1983. The flood debris contains vintage early 1980s material such as beer and soda-pop cans without disposable tabs and partly decomposed plastic artifacts. This sand and related floodmarks are below the level of the pre-dam alluvium.

The high-flow sand (unit hf in Fig. 19.2) is a very fine- to fine-grained sand with silt and clay content greater than 5 percent. Geomorphically, the high-flow sand is similar to the flood sand except that it is inset beneath the 1983 deposit. The high-flow sand was deposited during high flows from 1984 to 1986.

Sand deposited in the active channel below the 1986 high-flow level was partly reworked and covered during a controlled, experimental flood between March 22 and April 8, 1996, when flow rates were increased to 1270 m^3/s (45,000 ft^3/s). The stage of the experimental flood was near the upper limit of the high-flow sand. This controlled flood release was designed to reposition sand submerged in the channel onto subaerial river banks. Although we have not mapped these deposits, we anticipate that the experimental flood sand (unif efs of Fig. 19.2) was deposited just below the level of the high-flow sand.

Generally, the post-dam alluvial deposits record the depositional activity of the Colorado River since 1983. The largest flows of the post-dam era (Hyatt 1990) produced the 1983 flood sand in June–August of 1983, when peak discharge was 2700 m^3/s (96,000 ft^3/s) and sustained flows were above 1400 m^3/s (50,000 ft^3/s). During May–June of 1984–1986, sustained daily releases were the second highest of the post-dam era, ranging from about 900 to 1400 m^3/s (32,000 to 50,000 ft^3/s); this flow regimen resulted in the high-flow sand. Thus, a broad pattern of erosion followed by deposition at progressively lower levels produced the present configuration of the post-dam channel.

Sand Dunes

Eolian deposits forming dunes ranging in height from a few meters up to 10–20 m are widespread along the Colorado River in Grand Canyon. Grain-size analyzes show that the sand is moderately well-sorted to moderately sorted very fine- to fine-grained sand. Average silt and clay content is 5 percent with a range of 2–9 percent; this contrasts with pre-dam alluvium, which usually has average silt and clay content of 7 percent with a range of 2–16 percent. The eolian sand forms dunes, sand sheets, and other dune-like features that blanket older deposits. For the most part, these are coppice dunes or nabkhas, the terms applied to sand hummocks or mounds that develop around plants, which partially anchor the wind-blown sand (McKee 1982b, pp. 48–49; Cooke et al. 1993, pp. 356–358). In the river corridor, mesquite is usually associated with the coppice dunes.

The dunes are classified as active, moderately active to inactive, and largely inactive (units ea, ey, and eo of Fig. 19.2). The degree of eolian activity coincides with vegetation cover and topographic relief of the dune. Active dunes are sparsely vegetated and have relatively high relief. Vegetation is more abundant and relief is relatively subdued on the less active dunes (ey and eo).

The alluvial deposits of the river corridor are the immediate source of wind-blown sand. Eolian deposits typically occur downwind of sand flats formed on gravel bars, downwind of a terrace rise, or parallel with a terrace rise. Eolian erosion is enhanced where alluvial deposits are directly exposed to the wind, either by having large surface area or by exposure of sand in steep banks. The sand flats provide the fetch necessary for wind erosion, and a terrace rise exposes sand along the steep bank.

Holocene Tributary Debris Fans

Large boulders of locally derived bedrock are deposited in the river corridor by debris flows (Webb et al. 1988, 1989). We classify debris flows with colluvium (Fig. 19.2) to emphasize their close association with mass movement (Costa 1984; Hooke 1987). These impressive, potentially dangerous events result from mass wasting of the steep, talus-covered slopes of Grand Canyon following intense rainfall. For the most part, a mixture of water and sediment is driven downslope by the weight of the sediment, without water as the transporting medium. At the base of the slope, the debris enters a tributary channel and flows to the river. There, the deposits accumulate on debris fans, roughly fan-shaped landforms at the mouths of most tributaries (Hamblin and Rigby 1968; Dolan et al. 1978). The river is unable to move the debris except during relatively large floods (Graf 1980; Kieffer 1985; Webb et al. 1996). As a result, these deposits determine the course of the river between bedrock walls, the location and severity of rapids, and sites of alluvial deposition (Graf 1979; Howard and Dolan 1981; Kieffer 1985). In map view, these alluvial deposits encircle the debris fans outlining the fan-like shape (Hereford 1996; Hereford et al. 1997, 2001, 2001a).

Surficial Geology and Geomorphology of Large Tributary Debris Fans

The geomorphology of the large debris fans discussed here (Fig. 19.1, nos. 1–26) includes two elements (Hereford et al. 1996a). The larger element is the primary fan surface, which is segmented into several surfaces of different ages. Segmentation of fan surfaces is discussed by Cooke et al. (1993, pp. 19–184). The larger debris fans are evidently similar to Type IB alluvial fans of Blair and McPherson (1994, pp. 394–395), except that fan margins are extensively eroded by the Colorado River. These fans have a stratigraphic record of multiple, clast-rich debris flows and minor interbedded fluvial gravels. In the Grand Canyon, the deposits underlying the main fan surfaces are referred to as the fan-forming debris-flow deposits, because the primary surfaces are developed on these deposits.

The smaller element is an active debris-flow channel with a small debris fan at the river. The active channel is usually entrenched 2–5 m below the segmented surface, a notable exception is the deeply entrenched channel of the Prospect Canyon fan (Webb et al. 1996). Deposits transported down the active channel spread into the channel of the Colorado River, forming a small debris fan at the margin of the primary fan. Deposits of the entrenched channel and related fan are referred to as the channelized debris-flow deposits. The channelized deposits are identical to the historic-age debris flows or inset debris-flow surfaces discussed by Griffiths and others (Chapter 20, this volume). In contrast, the fan-forming deposits are mainly of prehistoric age, and they do not have a historic-age counterpart at the debris fans we studied (Fig. 19.1, nos. 1–26).

Both the fan-forming and channelized debris-flow deposits consist primarily of clast-supported angular to subangular boulders of local bedrock ranging in size from granules to large boulders as large as 3–5 m on an edge. The boulders are a mixture of resistant Paleozoic sandstone, limestone, and dolomite, which are present in the walls of the canyon and steep talus slopes beneath the cliffs (Huntoon et al. 1986). The debris-flow matrix is a poorly sorted mixture of clay to coarse sand and granule gravel. Large, angular boulders up to 1–3 m on an edge occur near the apex of many fans, although boulders this large are also present at the margin of the fans. Broadly speaking, the degree of rounding and maximum size of boulders are related to the size of the drainage basin (Hereford et al. 1997). Small basins with relatively short, steep channels produce large, angular boulders, whereas larger basins with relatively long channels and low gradients produce smaller subrounded to subangular boulders. Fluvial gravel of tributary origin is locally interbedded with the debris-flow sediment. The fluvial gravel typically consists of subrounded weakly imbricated pebble to small boulder-size clasts with minor coarse sand matrix; these gravels are distinctly finer grained and relatively well sorted compared with debris-flow gravel.

Large-scale topographic maps (Lucchitta 1991; Hereford et al. 1993, 1997, 2001; Hereford 1996; Webb et al. 1996) show that the area of the fan-forming deposits is substantially larger than the area of the entrenched channel and its associated fan. Typically, the area of the primary fan surface is about six times larger than the area of the active debris-flow channel and related fan, based on the median ratio of fan to channel area (Hereford et al. 1996a). In eastern Grand Canyon, the total area of the debris fans we studied ranges from 1 ha (2.47 ac) to 22 ha (54.3 ac) at upper Palisades Creek and Unkar Creek, respectively (Fig. 19.1, nos. 6 and 13).

The channelized and fan-forming debris-flow deposits are further subdivided according to age and topographic expression. Where we have mapped them (Fig. 19.1), the channelized debris-flow deposits consist of a younger unit forming the active debris-flow channel and an older unit forming one or more poorly to well-developed surfaces in the entrenched channel (units dcy and dco of Fig. 19.2, respectively). Both units were deposited by at least two debris flows in the past 100 years, based on the relation of the deposits to Colorado River alluvium of known age (Hereford et al. 1997, 2001; Hereford, 1996).

Between one and six fan-forming debris-flow deposits form the primary surface of the typical segmented fan in Grand Canyon (Fig. 19.1, nos. 1–26). These surfaces parallel the underlying deposits and are contemporaneous with deposition—as opposed to cross-cutting surfaces, which are younger than the underlying deposit. The deposits and surfaces are grouped into three main units: the older, intermediate, and younger fan-forming debris flows (units dfo, dfi, and dfy of Fig. 19.2), which are further subdivided where necessary. The surfaces are distinguished from each other by relative topographic position and surface-weathering characteristics (Table 19.1). The older surfaces have the highest elevation and are farthest from the river. In addition, the degree of surface weathering increases with distance from an elevation above the river.

Erosion of Prehistoric Debris Fans by the Colorado River

Geologic and large-scale topographic maps of several prehistoric debris fans reveal the extent of erosion by the Colorado River, which in turn affects how the fans aggraded. Most of the prehistoric debris fans are eroded and truncated at their margins. The rate and processes of debris-fan erosion by floods during historic time were discussed by Webb et al. (1996, pp. 84–87). This study showed that relatively small fans from channelized debris flows were largely removed from the active channel in only 1–2 years before regulation of streamflow and in 3–7 years following regulated streamflow. However, if fan-forming debris flows had substantially larger volume than channelized debris flows, as suggested by Hereford et al. (1993, 1996a), erosion of the larger deposit should require more than a few years and exposure to numerous large floods, given similar geomorphic setting.

Here we discuss several examples of prehistoric, partly eroded debris fans in eastern Grand Canyon (Hereford et al. 1996a). The distribution of the fan-forming deposits and the shape of the upper part of the Palisades Creek fan (Fig. 19.1, no. 6) resemble the radiating pattern normally associated with alluvial fans (Hooke 1987). An erosional scarp at the toe of the fan (Fig. 19.5a), however, is evidence that the Colorado River eroded and partly truncated the fan. The minimum height of the scarp ranges from 2 to 3 m. Downslope projection of the intermediate-age surface suggests that the fan probably extended across the river at Lava Canyon Rapids (Fig. 19.5a). The scarp truncates the older and intermediate-age debris-flow deposits, and it is overlapped by alluvium that is younger than 1200 A.D. (Hereford 1996). Therefore, erosion of the scarp occurred between 550 A.D. and 1200 A.D. Much of the fan remains unaffected by erosion because most of it lies well above the level of prehistoric and historic floods. Prehistoric alluvial deposits range in elevation from 823 m to 825 m around the margin of the fan, which is well below the fan apex at 845 m. The fan lacks well-developed inset relations among the surfaces, and it appears to have aggraded vertically (Fig. 19.5a).

Unlike the Palisades Creek fan, the lower Tanner Canyon fan (Fig. 19.1, no. 10) retains little of the original fan shape (Fig. 19.5b). Most of the lower Tanner Canyon fan lies within the range of prehistoric floods. Prehistoric flood-related alluvial deposits are downstream of the fan at elevations of 808–812 m; the elevation of the two fan segments range from 810 m to 815 m, which is well within the range of prehistoric floods. The intermediate and older debris-flow deposits are both truncated along two distinct west-trending scarps with relief of 3 m on the south scarp and 5 m on the north scarp (Fig. 19.5b). The erosional scarps were produced by at least two mainstem floods. The first flood or floods truncated the older debris flow. Truncation of the fan margin steepened the fan gradient, causing entrenchment of a channel. The intermediate-age flow was subsequently deposited in this channel, 2–3 m below the surface of the older deposit, and extended beyond the scarp in the older deposits. The second flood truncated the fan margin formed by the intermediate-age deposits (Fig. 19.5b). Finally, the fan has not aggraded vertically in the usual meaning of the term; instead, it developed mainly by aggradation of inset segments. The inset segments result from one or more floods that cut into the medial portion of the fan. This lowered the base level of the fan leading to entrenchment of the surface and emplacement of subsequent debris flows at a lower level.

Detailed topographic maps of Holocene fans in eastern Grand Canyon (Lucchitta 1991; Hereford et al. 1993) show that the Unkar Creek, Cardenas Creek, Comanche Creek, Espejo Creek, and the upper Palisades Creek fans (Fig. 19.1, nos. 13, 12, 8, 7, and 6, respectively) are moderately truncated, because they retain an almost complete fan shape. In contrast, the Basalt Creek and Tanner Canyon fans (Fig. 19.1, nos. 11 and 9) are severely truncated, as indicated by steep scarps, with up to 5 m of relief, that surround the margin of each fan. These different morphologies are probably related to the course of the river around the fan, the elevation of the fan relative to flood levels (which is related to particle size of the deposits), and the age of the fan.

The course of the channel relative to the debris fan differs between the moderately and severely truncated fans. At moderately truncated fans, the river is relatively straight and essentially parallel with the distal-fan margin, whereas at the severely truncated fans the river flows into and around the fan margin (Fig. 19.1, nos. 9 and 10, for example) or it flows entirely around the fan margin (Basalt Canyon; Fig. 19.1, no. 11). In these situations, the river is particularly effective at eroding the fan because it flows directly into the upstream margin of the fan, or the river completely surrounds the fan margin. A severely truncated fan could be older than a moderately truncated fan, because the older fan has been subjected to a larger number of mainstem floods. Large debris fans in eastern Grand Canyon and elsewhere in Grand Canyon (Fig. 19.1), however, differ in age by only a few thousand years.

Finally, the elevation of the fan relative to the river also controls the extent of erosional modification. The apex of the moderately truncated Palisades Creek fan (Fig. 19.1, no. 6) lies well above flood levels, suggesting that the volume and particle size of sediment was larger than the river was able to remove during recent millennia. On the other hand, most of the severely truncated lower Tanner Canyon fan (Fig. 19.1, no. 10) lies within the level of prehistoric floods, and the fan was relatively low and easily eroded. This difference between high and low elevation fans is evidently related to the slope or gradient of the fan, which in turn is a function of basin relief, lithology, and area (Cooke et al. 1993, pp. 177–178; Webb et al. 1996).

Table 19.1. Surface Weathering Characteristics of Fan-Forming Debris Flows, Grand Canyon, Arizona

Characteristic	*Debris-Flow Age Category* Younger	Intermediate	Older
Carbonate coatings underside of boulders	None	Incipient Stage I, discontinuous, thin, <0.1 mm	Stage I, continuous, thin <0.5 mm
Splitting, spalling, and granular disintegration of sandstone boulders	None	Slight to common	Common
Average pit depth carbonate boulders (mm)	0 to ~1.5	>1.5 to ~4.7	>4.7 to ~17.5
Rilling of carbonate boulders	None	None	Present on 5 percent of boulders
Rock varnish, sandstone boulders dark	Absent to incipient on 50 percent of boulders	Present on all boulders, brown to dark brown	Well-developed on all boulders, brown to black

Ages and Correlation of Fan Surfaces

The relative ages of fan surfaces or segments is established by the degree of surface weathering (Table 19.1). The younger surface has weakly developed or no discernible rock varnish and the carbonate boulders appear fresh with little surface roughness. On the intermediate-age surface, rock varnish coats 50–100 percent of the sandstone boulders and carbonate boulders are distinctly roughened with dissolution pits. The undersides of boulders on the surface have a thin, very light-gray to white discontinuous coating of calcium carbonate, which is weakly developed Stage I soil carbonate morphology (Machette 1985). On the older surface, up to 100 percent of the sandstone boulders have well-developed rock varnish; carbonate boulders are distinctly and deeply pitted; and the undersides of boulders have a thin, mostly continuous coating of calcium carbonate. On the oldest surfaces, rillenkarren (Table 19.1), a distinctive pattern of shallow grooves or rills, is on up to 5 percent of the carbonate boulders.

Although relative ages are readily determined using weathering criteria (Table 19.1), finding the numerical age of fan segments is difficult. The relation of debris fans to dated Holocene alluvial deposits places the age of the larger fans in the late Holocene (Hereford et al. 1996a). Dated archeologic sites are present on fan surfaces locally, but sites provide only minimum ages. Briefly, fan surfaces are difficult to date directly because organic material is extremely rare in the coarse-grained deposits, which severely limits the use of radiocarbon dating. However, the depth of dissolution pits on carbonate boulders increases systematically with the relative age of the surface as shown independently by surface weathering such as degree of patination and soil development (Hereford et al. 1996a). Dissolution pits result from weakly acidified rainfall when atmospheric or metabolic CO_2 combines with water to form carbonic acid (H_2CO_3), which slowly

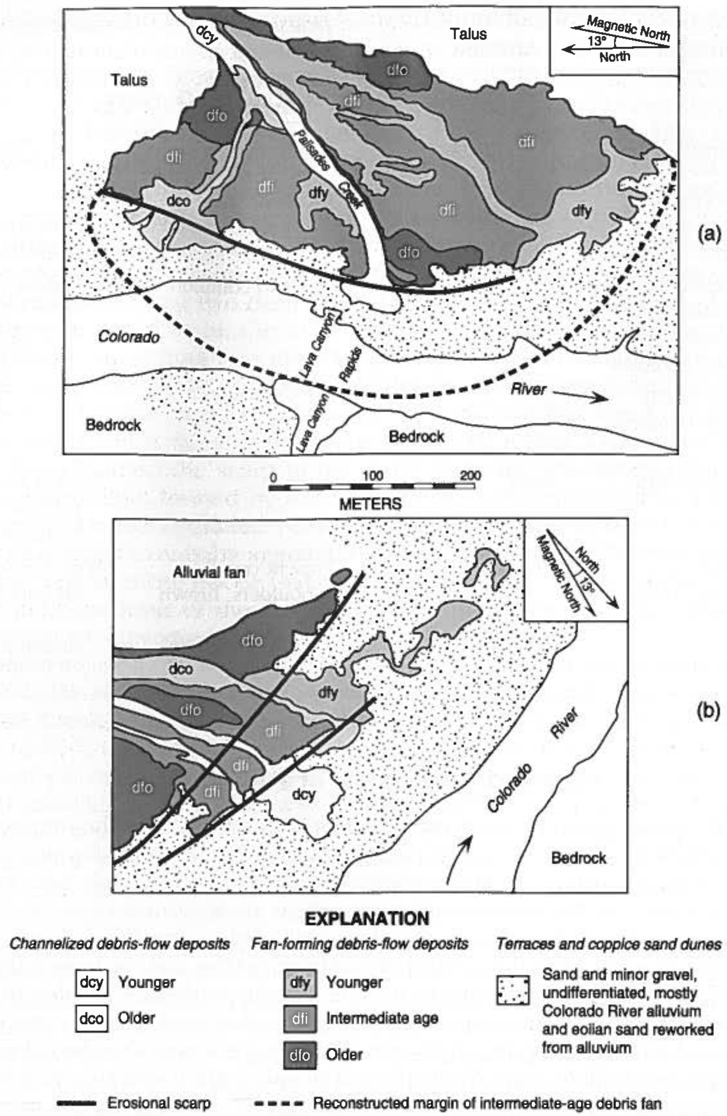


Figure 19.5. Generalized surficial geologic maps of eroded debris fans. (a) Moderately truncated Palisades Creek debris fan (Fig. 19.1, no. 6; Hereford, 1996) showing radiating pattern of fan forming debris flow deposits, reconstructed outline of intermediate-age debris-flow deposits, and trace of erosional scarp at base of fan. (b) Severely truncated lower Tanner Canyon fan (Fig. 19.1, no. 10) showing two erosional scarps at margin of older and intermediate-age deposits, respectively. (Modified from Hereford et al. 1996, Fig. 19.6.)

etches the surfaces of carbonate boulders. Thus, the depth of dissolution pits is related to the numerical age of a particular fan surface. The problem with using dissolution pits to estimate exposure time is that the rate of pit deepening must be established from independently dated surfaces, which are few in number.

Using the average depth of dissolution pits measured on four independently dated surfaces, the average rate of pit deepening was estimated to be 2.60 ± 0.12 mm/ka for debris fans in Grand Canyon. The procedure used to estimate deepening rate, methods of measuring and sampling pit depth, and statistical treatment of the data are in Hereford et al. (1997a). The above deepening rate was recalculated from the data in Hereford et al. (1997) using an improved computational procedure.

Limestone weathering is a linear function of time for at least the past 3 ka, as reported by numerous empirical studies of archeologic sites and gravestones [Hereford et al. (1997a) summarize the literature]. The physical processes of limestone weathering proceed at a constant rate—unlike weathering of silicate rocks, which decreases with time (Colman 1981). It is important to realize that the dissolution process is cumulative (Lipfert 1989). The influence on dissolution rate of increases and decreases in precipitation, therefore, averages out over time, maintaining an essentially constant long-term average rate.

Briefly, 6883 individual pit measurements were made on 618 boulders of Redwall Limestone on 71 fan surfaces at 26 tributary debris fans (Fig. 19.1). The measurements are summarized in Figure 19.6 by average depth and the corresponding 95 percent confidence interval for each of the 71 surfaces. The results are consistent with field relations indicating the relative age of fan segments at a particular fan. In other words, surfaces with deep dissolution pits appear older based on stratigraphic position and surface weathering criteria without considering pit depth (Table 19.1).

Ages estimated with the 2.60 mm/ka deepening rate range from 500 to about 7000 cal yr B.P. (calibrated or calendar years before present, i.e., 2001). The oldest surface on a debris fan, river mile 124.2L (Fig. 19.1, no. 17), has dissolution pits averaging 12.22 mm deep. This corresponds to about 4700 cal yr B.P. Even older surfaces are located at the mouths of Forster and Fossil Canyons and in the tributary canyon at river mile 220R. The three surfaces have dissolution pits averaging 17.29, 17.41, and 17.42 mm, respectively; this is about 6700 cal yr B.P. These surfaces are debris-flow levees preserved in the mouths of the tributary canyon, where they are protected from the erosional effects of the Colorado River.

The youngest surfaces datable by this method are at 209 Mile Canyon, Granite Park Wash, Palisades Creek, and Nankoweap Canyon (Figs. 19.1 and 19.9). Dissolution pits on the four surfaces average 1.2 to 1.47 mm deep, which is probably between 500 and 600 cal yr B.P. An average depth of around 1.2 mm is probably the minimum amount of surface pitting detectable by this method; carbonate boulders on fan-forming surfaces younger than this do not have measurable dissolution pits (Table 19.1).

The typical time between fan-forming debris flows at a particular fan is estimated from the differences in pit depth calculated for tributary fans with more than one surface (23 of the studied fans). The median difference ($n = 45$) in average depth of dissolution pits on the 23 fans is 2.13 mm, and the interquartile range is 1.32 to 3.22 mm. This suggests that surfaces of the large fans studied here differ in age by about 820 years within a range of 500 to 1200 years. This 820-year average interval is much longer than the 10- to 50-year average recurrence interval (Chapter 20, this volume) of channelized debris flows during historic time.

Several widely spaced tributaries have debris-fan surfaces with exposure times that cannot be shown to be different based on average pit depth; these are connected by the five horizontal lines in Figure 19.6. The surfaces are evidently time correlative within the limitations of this method, suggesting that conditions leading to formation of debris-fan surfaces influence the entire Grand

Canyon region, although these conditions do not necessarily trigger debris flows in every tributary.

Based on average pit depth, most of the surfaces fall into five broadly defined age clusters between about 800 and 4000 cal yr B.P. Five episodes of widespread debris-flow activity are indicated by our mapping, which delineates five to six surfaces of different age canyon-wide (Hereford et al. 1997, 2001). The surfaces in each cluster are shown by the pattern of alternating solid and open symbols with heavy lines in Figure 19.6. Surfaces were assigned to a particular cluster using *k*-means cluster analysis (Davis 1986, pp. 513–514). Surfaces at the younger and older limits of a cluster are not clearly separated from adjoining clusters. However, in most cases, surfaces within about ± 0.5 mm of the cluster centroid (vertical bars on horizontal lines in Fig. 19.9) are clearly separated from those in nearby clusters, based on nonoverlapping confidence intervals.

The five clusters are around 2-, 3.9-, 5.5-, 7.5-, and 10.4-mm average pit depth, which corresponds to about 800, 1500, 2100, 2900, and 4000 cal yr B.P., respectively. Thus, five broadly defined episodes of heightened debris-flow activity are discernible (Fig. 19.6), although the occurrence of fan-forming debris flows was highly variable during the late Holocene. These in turn may correspond to episodes of increased precipitation with each episode lasting perhaps several centuries.

A striking feature of the correlation chart is that most of the fan surfaces are younger than about 4.7 ka. The surfaces are mostly late Holocene, and 75 percent of them are younger than 2.8 ka. Preservation of prehistoric debris fans is linked to interaction between the frequency of fan-forming debris flows and removal of the deposits by the Colorado River. This interaction is probably controlled by the climate of the Grand Canyon region through its effect on local debris-flow activity and by the climate of the Colorado River drainage basin, which controls the size and frequency of mainstem floods. Only fans deposited in the last 3 to 4 ka are preserved (Fig. 19.6), suggesting that gradual, flood-related shifting of the main channel eventually removed early and mid-Holocene deposits.

An alternative explanation for the lack of early and mid-Holocene deposits in the river corridor is that they are present, but are buried by net aggradation of the fans. This seems unlikely, however, given that older surfaces have the highest elevation relative to younger surfaces on any particular fan. Early Holocene debris-flow deposits are present (Fig. 19.6) in protected sites in the mouths of tributary canyons. However, the surfaces of these deposits project downstream to the debris fan where they are topographically above younger surfaces. Thus, it seems likely that the absence of early and mid-Holocene debris fans resulted from lack of preservation in flood prone parts of the river corridor.

SUMMARY

The Holocene surficial geology of the Colorado River in Grand Canyon consists mainly of terraces and related alluvium, tributary debris fans, and sand dunes. These deposits are well-developed where the river corridor is at least 200–400 m wide. The deposits are mainly late Holocene; mid-Holocene deposits are uncommon, and early Holocene deposits are as yet unknown. The absence of older Holocene alluvium and debris fans probably results from poor long-term preservation along the narrow, flood-dominated channel of the Colorado River.

Perhaps the most important deposits are those forming prehistoric tributary debris fans, because they control the position of the channel and sites of high-level alluvial deposition. These bouldery, roughly fan-shaped deposits resist erosion, producing rapids and forcing the river to flow around the fans. Areas of low-current velocity develop around debris fans enhancing formation of high-level terraces. These alluvial deposits are in turn reworked by wind, forming dunes, sheets, and mounds of wind-blown sand.

Large debris fans of late Holocene age with multiple surfaces are present throughout Grand Canyon. These are composed of sediment derived from tributaries that was deposited in the river corridor by debris flow. The deposits are very poorly sorted and extremely coarse grained, boulders up to several meters in maximum diameter are locally abundant. The geomorphology of the fans consists of the inactive-primary fan surface and the entrenched, active debris-flow channel. Deposits related to the primary fan surfaces are termed fan-forming debris-flow deposits, and those in the channel are referred to as channelized debris-flow deposits. The fan-forming deposits are mainly of prehistoric age, although four of them were deposited as recently as 90–200 cal yr B.P. In almost every case, the channelized deposits are of historic age, having been deposited within the past 5–100 cal yr B.P.

The fan-forming deposits are divided into three relative age categories based on the degree of surface weathering. The youngest category is essentially unweathered with only incipient rock varnish and no subsurface soil development. Intermediate-age surfaces have rock varnish on most sandstone boulders, boulders that are weakly to moderately disintegrated, weakly developed Stage I soil-carbonate morphology, and carbonate boulders with relatively shallow dissolution pits. The older category has sandstone boulders with relatively dark patination, moderately developed Stage I soil carbonate, heavily disintegrated boulders, and carbonate boulders with relatively deep dissolution pits.

The numerical ages of fan surfaces are difficult to determine using standard dating methods, because organic material is extremely rare in the coarse-grained deposits. Several numerical dates, however, have been obtained from debris-flow deposits. The depth of dissolution pits on carbonate boulders is observed to increase systematically with degree of surface weathering. Using the average pit depth on carbonate boulders of dated fan surfaces, the rate of pit deepening is estimated to be 2.60 ± 0.12 mm/ka, and the rate is independent of time through at least the late Holocene. Using this rate, we determined the ages of 71 primary fan surfaces at 26 debris fans throughout Grand Canyon.

The oldest surface on a debris fan is 4700 cal yr B.P. Three debris-flow levees in protected sites just within the mouth of tributary canyons are the oldest dated surfaces at about 6700 cal yr B.P. The youngest surfaces datable by this method (several centuries are required for dissolution pits to develop) formed between 500 and 600 cal yr B.P. For the most part, 75 percent of the dated surfaces are younger than 2.8 ka. At a particular fan, the average time between fan-forming debris flows is about 820 years, indicating that the primary fan surfaces are relatively dormant compared with debris-flow activity in the entrenched channel. We believe that a primary fan surface results from an unusually high-volume debris flow or from a large number of small debris flows. Comparable in volume to historic-age flows, frequent relatively low-volume debris flows could eventually overtop the entrenched channel and spread over the primary surface. In either case, conditions leading to deposition on the primary fan surface are probably unusual as they occur infrequently.

Sandy alluvial deposits associated with large tributary debris fans range in age from 2500–1300 B.C. to the present. Five deposits form terraces that record the depositional activity of the Colorado River before completion of Glen Canyon Dam in 1963. These are informally named the striped alluvium (2500–1300 B.C. to 300 A.D.), alluvium of Pueblo II age (700–1200 A.D.), upper mesquite terrace (1400–1880 A.D.), lower mesquite terrace (1884 A.D. to early 1920s A.D.), and the pre-dam alluvium (early 1920s A.D. to 1957–1958 A.D.). A number of important archeologic sites are associated with the alluvium of Pueblo II age, which was deposited for the most part during Anasazi occupation of eastern Grand Canyon.

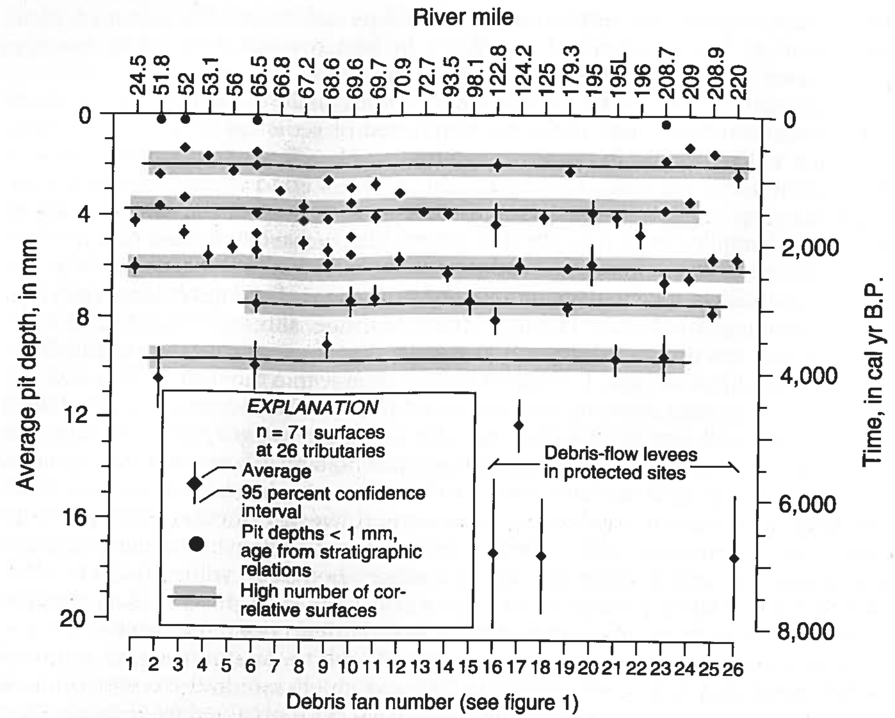


Figure 19.6. Age and correlation chart of debris-fan surfaces inferred from average depth of dissolution pits. From top to bottom, alternating solid and open symbols with heavy lines are surfaces clustered about the five horizontal lines, respectively. (Modified from Hereford et al. 1997a.)

Two major periods of erosion and nondeposition separate the late Holocene alluvia. The oldest, between A.D. 300–700, intervened between deposition of the striped alluvium and alluvium of Pueblo II age. The youngest, between 1200 A.D. and 1400 A.D. separates the alluvium of Pueblo II age from the alluvium of the upper mesquite terrace.

Broadly speaking, the alluvial deposits resulted from flood-related aggradation of the river banks; in places the terraces are extensive and may have been floodplains. Indeed, evidence suggests the floodplain of the alluvium of Pueblo II age was locally cultivated by the Anasazi. The erosional episodes isolated the floodplains from the active channel, which formed the terraces. The causes of alternating deposition and erosion are poorly understood, although climate, through its control on flood regimen and sediment load, is probably a first-order factor. Indeed, the broad similarity between alluvial chronologies of southern Colorado Plateau streams and the alluvial history of the Colorado River suggests a common causal mechanism.

Finally, the channel system is out-of-balance with regard to fan-forming debris flows. Over the life span of Glen Canyon Dam, one or more fan-forming events are expected to occur at any major tributary. In the recent past, when such debris flows entered the channel, perhaps forming a waterfall-like obstruction or greatly increasing the severity of a rapid, the debris was eventually removed by large mainstem floods. In the future, however, such channel obstructions will be difficult to remove given the limitations imposed by Glen Canyon Dam.

• 20 •

Debris Flows and the Colorado River

Robert H. Webb, Theodore S. Melis, and Peter G. Griffiths

INTRODUCTION

Debris flows are a type of flash flood that occur throughout Grand Canyon National Park (Webb et al. 1988, 1989, 1996) and elsewhere in the United States where unconsolidated deposits reside on steep terrain (Sharp and Nobels 1953; Glancy 1969; Williams and Guy 1971; Johnson and Rodine 1984; Osterkamp et al. 1986; Kochel 1987; Wells 1987; Wieczorek et al. 1989; Wohl and Pearthree 1991; Whipple and Dunne 1992; Hammack and Wohl 1996). These slurries of water, mud, and rock move with spectacular swiftness through the tributary canyons of the Colorado River. The colorful, sculptured cliffs enclosing Grand Canyon attract millions of tourists every year; these same cliffs, when subjected to intense summer thunderstorms, are the reason that debris flows are common here. The steep exposures of unstable bedrock, containing abundant clay, are ideal for initiation of debris flows, and debris-flow deposition is the primary reason for rapids in the Colorado River.

Debris flow is a generic term for a fluid with a sediment concentration of about 80 percent or greater (Pierson and Costa 1987). *Mudflows* contain more clay and fewer boulders, and other terms, such as *debris torrent* and *mudslides*, have special connotations. Debris flows in the southwestern United States were first described by Blackwelder (1928). Péwé (1968) briefly described a 1967 debris flow at Lees Ferry. The first scientific descriptions of debris flows in Grand Canyon resulted from the flood of December 1966 in Crystal and Lava-Chuar Creeks (Cooley et al. 1977). Ford et al. (1974) report several small debris flows near the mouth of the Little Colorado River in. More than 25 debris flows occurred in Grand Canyon between 1983 and 1996 (Melis et al. 1994; Webb et al. 1996).

Most debris flows are related to fault uplift or landscape disturbances such as volcanic eruptions, forest fires, or poor land-use practices (Pierson 1985; Gallino and Pierson 1985; Meyer et al. 1995). In Grand Canyon, as well as in nearby river canyons, debris flows occur because the Colorado River has rapidly downcut through bedrock, leaving ideal conditions for debris-flow initiation. In addition, the arid climate minimizes erosion of the limestone cliffs, thereby creating steep topography; summer thunderstorms provide energetic rainfall that is required for debris-flow initiation. Debris flows transport boulders and finer sediment relatively long distances because the bedrock channels of most tributary canyons confine the flow to minimize losses to deposition.

Many large debris flows have swept down drainages in Grand Canyon during the last century (Table 20.1). Boulders deposited during the 1939 debris flow,

TABLE 20.1. Major Debris Flows in Grand Canyon Between 1890 and 1997 that Deposited Sediment in the Colorado River

Tributary	*Rapid*	*Miles*	*Side*[a]	*Year or Range of Largest Flow*	*Effects*
Badger Creek	Badger Creek	7.9	R	1897–1909	Changed right side of rapid
Soap Creek	Soap Creek	11.2	R	1923–1934	Made navigation easier
Rider Canyon	House Rock	16.8	R	About 1966	Increased constriction
22-Mile Wash	22-Mile	21.4	L	Unknown	Unknown
Unnamed canyon	24-Mile	24.1	L	1989	Changed rapid completely
Tiger Wash	Tiger Wash	26.6	L	Unknown	Increased constriction
Unnamed canyon	None	30.2	R	1987?	Deposited new debris fan
South Canyon	Unnamed	31.6	R	1940–1965	Completely changed fan
Unnamed canyon	President Harding	43.7	L	1984	Changed left side
Unnamed canyon	New	62.5	R	1990	Created new rapid
Lava Creek	Lava Canyon	65.5	R	1966	Unknown
Palisades Creek	Lava Canyon	65.5	L	1990	Increased constriction
75-Mile Creek	Nevills	75.5	L	1990	Deposition on debris fan
Hance Creek	Sockdolager	78.7	L	Unknown	Deposition on left side
Monument Creek	Granite	93.5	L	1984	Increased constriction
Hermit Creek	Hermit	95.5	L	1996	Increased constriction
Boucher Creek	Boucher	96.7	L	1950s	Changed rapid
Crystal Creek	Crystal	98.3	R	1966	Changed rapid
Waltenberg Canyon	Waltenberg	112.2	R	1940s	Increased constriction
Unnamed canyon	New	127.6	L	1989	Created new rapid
128-Mile Creek	128-Mile	128.5	R	1890–1923	Increased constriction
Specter Chasm	Specter	129.0	L	1989	Increased rapid severity
Bedrock Canyon	Bedrock	130.5	R	1989	Increased constriction
Unnamed canyon	Unnamed	133.0	L	Unknown	Deposition on debris fan
Kanab Canyon	Kanab	143.5	R	1923–1942	Increased constriction
Prospect Canyon	Lava Falls	179.3	L	1939	Changed rapid
205-Mile Canyon	205-Mile	205.4	L	1937–1956	Changed rapid
Diamond Creek	Diamond Creek	225.5	L	1984	Increased constriction
231-Mile Canyon	231-Mile	230.8	R	Unknown	Increased constriction

[a]L, left side of the river when viewed from upstream; R, right side of the river.

Source: Modified from Webb (1996).

the largest historic event, completely changed Lava Falls Rapid (Webb et al. 1996, 1997). The better-known 1966 debris flow in Crystal Creek changed a benign rapid into challenging whitewater (Cooley et al. 1977; Kieffer 1985; Webb et al. 1989). No one saw these debris flows, but two others were witnessed. The eyewitness accounts vividly describe this fascinating type of flash flood.

South Canyon, July 1889

In the summer of 1889, Robert Brewster Stanton's expedition to determine the feasibility of a railroad at river level through Grand Canyon had failed, with three men lost to drowning. On July 18, Stanton and his remaining crew stashed their remaining gear, released their boats, and hiked up South Canyon (mile 31.5). A mid-morning thunderstorm overtook them and, according to Stanton

> As the rain commenced to fall we heard some rocks roll down the slope behind us, when we looked up, and it seemed as if the whole slopes of the gorge had begun to move at the top. Little streams of water came over the top, and in a moment they changed into streams of mud; and as they came down they gathered strength and turned to streams of mud and rock, undermining larger rocks; and starting them they plunged ahead, and in a few moments the whole sides of the canyon seemed to be moving down upon us with a roar and awful rumbling noise; and as the larger rocks plunged ahead of the streams, they crashed against other rocks, breaking into pieces; and the fragments flew in to the air in every direction, hundreds of feet above our heads; and as these came nearer the bottom where we were, it seemed as if we were to be buried in an avalanche of rock and mud (Smith and Crampton, 1987).

The debris flow Stanton described is similar to ones that occur every year somewhere in Grand Canyon. Intense rainfall dislodges rocks and unconsolidated sediment from the steep slopes and cliffs, which mixes with storm runoff and forming a slurry that typically travels less than a kilometer. These small debris flows do not have much energy and usually stop on gentle slopes, loading channels with debris that can then fuel larger debris flows in the future.

Diamond Creek, July 1984

On July 20, 1984, two river-rafting companies completed their Grand Canyon trips at Diamond Creek in western Grand Canyon. At 3:30 P.M., the guides began driving two trucks up the road, which is little more than a channel bed smoothed by bulldozers in Diamond Creek. A flood the day before had washed out the road. The guides saw no sign of the severe thunderstorm that had just drenched the headwaters of this large drainage. About a kilometer up Diamond Creek, one of the trucks stalled in the middle of the channel. What happened next was described in detail by Ghiglieri (1992) and Webb (1996).

A guide walking up the road for help watched as a "wall of water" spread over a wide section of Diamond Creek, knocking trees over like they were matchsticks. In the constricted reach downstream where the trucks were stalled, the other guides heard the roar of the approaching flood, which to them sounded like a bulldozer. As the flood bore down on the trucks, the guides scrambled up nearby slopes to safety. The debris-flow front, nearly 2-m high, was described as a dark slurry containing wood but no visible boulders. The lead truck rose and pirouetted into the flow, accompanied by the sound of breaking glass. The second truck was quickly swept away by the flood, and both vehicles were last seen floating upside down with their wheels out of the slurry. At the Colorado River, other witnesses saw a wave 5–6 m high form when the debris flow entered the rapid, and the vehicles rose on the wave. The vehicles were found a few months later at the bottom of Diamond Creek Rapid (Ghiglieri 1992, p. 276).

The Diamond Creek flood demonstrated the awesome transport power of debris flows. The front that hit the vehicles, called the *snout*, was a mixture of wood, mud, and boulders; the larger particles were hidden behind a curtain of mud and floating wood. The trucks became particles entrained in the debris flow, floating like corks on the surface of the flow despite their weight. The slurry pushed the Colorado River to the side, changing the left side of Diamond Creek Rapid. Ultimately, the river reclaimed most of its channel, moving the trucks and other particles to the foot of the rapid.

THE PROCESS OF DEBRIS FLOW

Debris flows in Grand Canyon have three distinct phases: initiation, transport, and deposition. These phases are well illustrated by the 1984 debris flow in Monument Creek (Webb et al. 1988). Intense rain pelted a slope of Hermit Shale, and slurries of mud and colluvium from the overlying units moved quickly down the steep slopes. A large block of Esplanade Sandstone gave way as an avalanche and mixed with the slurries as they plunged over a 300-m cliff of Redwall Limestone. At the base of the cliff the mixture of boulders, ground-up rock, and muddy water had enough energy to flow down the lower gradient stream channel, leaving behind a deposit of the avalanche material. Boulders and finer sediment from the channel margins were entrained into the debris, causing it to "bulk up." As the debris flow approached the Colorado River, the slurry no longer was confined by bedrock. Most of the sediment and water decelerated, spread out over the debris fan, and stopped. The snout had enough momentum to enter the Colorado River, and deposition of large boulders diverted river flow from the left side of the rapid. The entire process, from initiation to deposition, was extremely fast; Webb et al. (1988) estimated that the main debris-flow pulse lasted only 3 minutes.

Initiation

Initiation of debris flows in Grand Canyon requires intense rainfall on steep slopes resulting in a mass movement, or slope failure, in consolidated or unconsolidated sediment. The intense rainfall and height of the initiation point above the river are the sources of energy. Slope failures supply the bulk sediment and plunge pools at the bases of cliffs, or long chutes through colluvial wedges, are where the mixing of sediment and water takes place. Although debris flows can be mobilized in any type of sediment, the ones that travel the longest distances are those derived from certain shales and their associated colluvial wedges.

The intense, sometimes protracted thunderstorms of July through September initiate most of the debris flows in Grand Canyon. Thunderstorms are either (a) widespread, affecting numerous tributaries, or (b) concentrated over one tributary. Few rain gages are present in areas where debris flows begin (Griffiths et al. 1997). Rainfall intensity during historical debris flows has ranged from 10 to 40 mm/hr at remote climate stations, and rainfall may last several hours. Debris flows do not necessarily occur during above-average rainfall seasons, although precipitation during the preceding weeks or month is typically well above normal (Melis et al. 1994; Webb et al. 1996; Griffiths et al. 1997).

In addition to summer thunderstorms, certain types of regional storms also cause debris flows. Thunderstorms from a dissipating tropical cyclone caused debris flows in Prospect Canyon in 1939 (Webb et al. 1996). The Crystal Creek flood of December 1966 resulted from a warm winter storm, with rainfall falling on a preexisting snowpack (Cooley et al. 1977). Historically, this type of storm spawned debris flows only in 1966 and 1995. Winter storms that have initiated debris flows typically have had lower intensities, and debris-flow initiation is likely a function of both rainfall intensity and duration (Wilson and Wieczorek 1995). In the Grand Canyon region, a closing period of higher intensity precipitation defines storms that produce debris flows; otherwise, streamflow floods result (Webb et al. 1996).

TABLE 20.2. Clay Mineralogy of Shales, Colluvium, and Debris-Flow Deposits in Grand Canyon

Formation	*Member*[a]	*Depositional Environment*	*Illite*[b] *(%)*	*Kaolinite*[b] *(%)*	*Smect ite*[b] *(%)*	*Other*[b] *(%)*
		Shales not Producing Debris Flows				
Chinle Formation	Petrified Forest	Lacustrine	41	11	42	6
Hakatai Shale[c]	Middle	Near-shore	52	15	2	31
Tropic Shale		Marine	16	19	57	8
mean ± standard deviation			36 ± 18	15 ± 4	34 ± 28	15 ± 14
		Shales Producing Debris Flows				
Bright Angel Shale			68	22	0	10
Chinle Formation	Monitor Butte?	Fluvial	32	44	7	17
Dox Formation	Lower Middle	Fluvial	65	14	0	21
Esplanade Sandstone	Basal shale	Fluvial	50	40	2	8
Galeros Formation	Carbon Canyon	Lacustrine	63	23	0	14
Galeros Formation	Jupiter	Lacustrine	35	42	2	8
Hermit Shale		Fluvial	54	41	0	5
mean ± standard deviation			52 ± 14	32 ± 12	1 ± 3	14 ± 7
		Colluvial Wedges				
Mean ± standard deviation ($n = 16$)			34 ± 13	43 ± 19	5 ± 5	18 ± 8
		Debris Flow Deposits				
Mean ± standard deviation ($n = 10$)			41 ± 15	28 ± 9	6 ± 11	25 ± 11

[a]Nomenclature follows summary in Elston et al. (1989).

[b]Minerals identified by semiquantitative x-ray diffraction. Margin of error ± 20% by weight (Starkey et al. 1984).

[c]Typically not high enough on cliffs to produce debris flows.

If initiation is to occur, intense rainfall must trigger slope failures in bedrock or loose colluvium. Shales form unstable slopes, whereas sandstones and limestones form cliffs. Massive, uniform rock layers such as the Redwall Limestone are very stable, whereas units comprised of alternating limestone, sandstone, and shale layers, such as strata of the Supai Group, are unstable and susceptible to failure. Susceptibility has less to do with "hard" and "soft" units and more to do with thickness of strata and presence of underlying shales. In other words, sandstones from the Supai Group may be as resistant or more so than the Redwall Limestone, but stratification in the Supai Group produces the critical instability of thin, resistant layers being undercut by erosion of the weaker layers.

Certain types of clay minerals increase the propensity of bedrock failure and debris-flow mobilization. A mixture of sand, silt, clay, and water fills the interstices between larger particles. This fluid has a higher density and apparent viscosity than water. Bedrock units that contain significant amounts of clay are the most important units for the initiation of debris flows. Failures in the Hermit Shale, for example, have contributed to most of the largest historical debris flows in Grand Canyon. The Muav Limestone, a silty dolomite that grades downward into the Bright Angel Shale, is another important stratum producing slope failures. Many units of the Precambrian Grand Canyon Series, particularly the Dox Sandstone, contain enough fine particles to produce debris flows.

Not all shale units present in the Grand Canyon region produce debris flows. The clay mineralogy of shales greatly affect their stability (Trask 1959; Hampton 1975; Griffiths et al. 1996). Shales that produce debris flows are composed primarily of illite and kaolinite and are of terrestrial origin (Table 20.2). These single-layer, nonswelling clays allow deep penetration of rainfall to failure surfaces, whereas the cracks in multilayer, swelling clays, such as smectites, quickly close after wetting, preventing deep percolation. The only terrestrial shales that do not produce debris flows are the lacustrine strata in the Chinle Formation. These strata are diagenetically altered volcanoclastic sediment, which results in a high content of smectites. Marine units, such as the Tropic Shale in the vicinity of Lake Powell, are high in smectites (Table 20.2). Colluvial wedges that produce debris flows, and the debris-flow deposits themselves, have clay mineralogies similar to the terrestrial shales.

Major cation chemistry of bedrock units affects their proclivity for failure (Hampton 1975; Pierson and Costa 1987). Sodium causes clay minerals to disperse, sealing cracks and inhibiting deep percolation. Shales produced in a marine environment typically are high in sodium. Terrestrial shales typically are low in sodium and high in magnesium. All shales in the Grand Canyon region are high in calcium, and the presence of dispersed gypsum greatly increases the potential for slope failures. For example, the Fossil Mountain Member of the Kaibab Formation in western Grand Canyon does not contain significant shale layers, but the Harrisburg Member, a prominent gypsum-bearing unit, produces small debris flows.

Even the most failure-prone lithologic units do not produce debris flows when they are close to river level. The longitudinal profile of the Colorado River in upper Marble Canyon (Fig. 20.1) illustrates the relation between height of

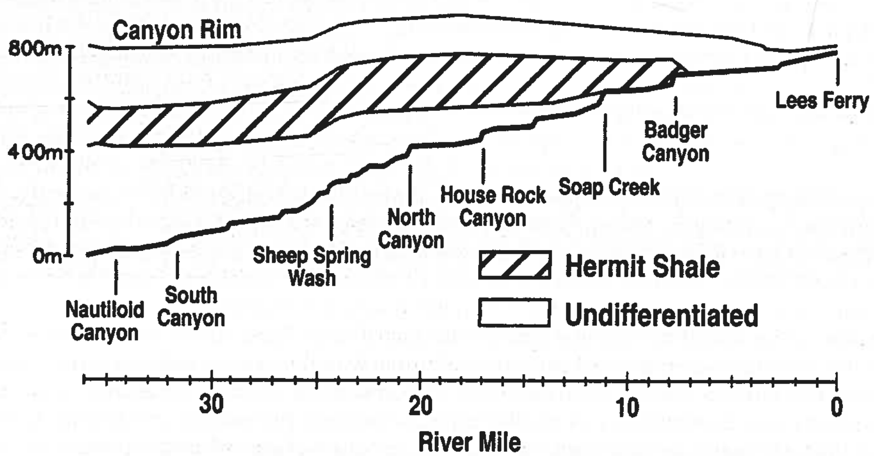
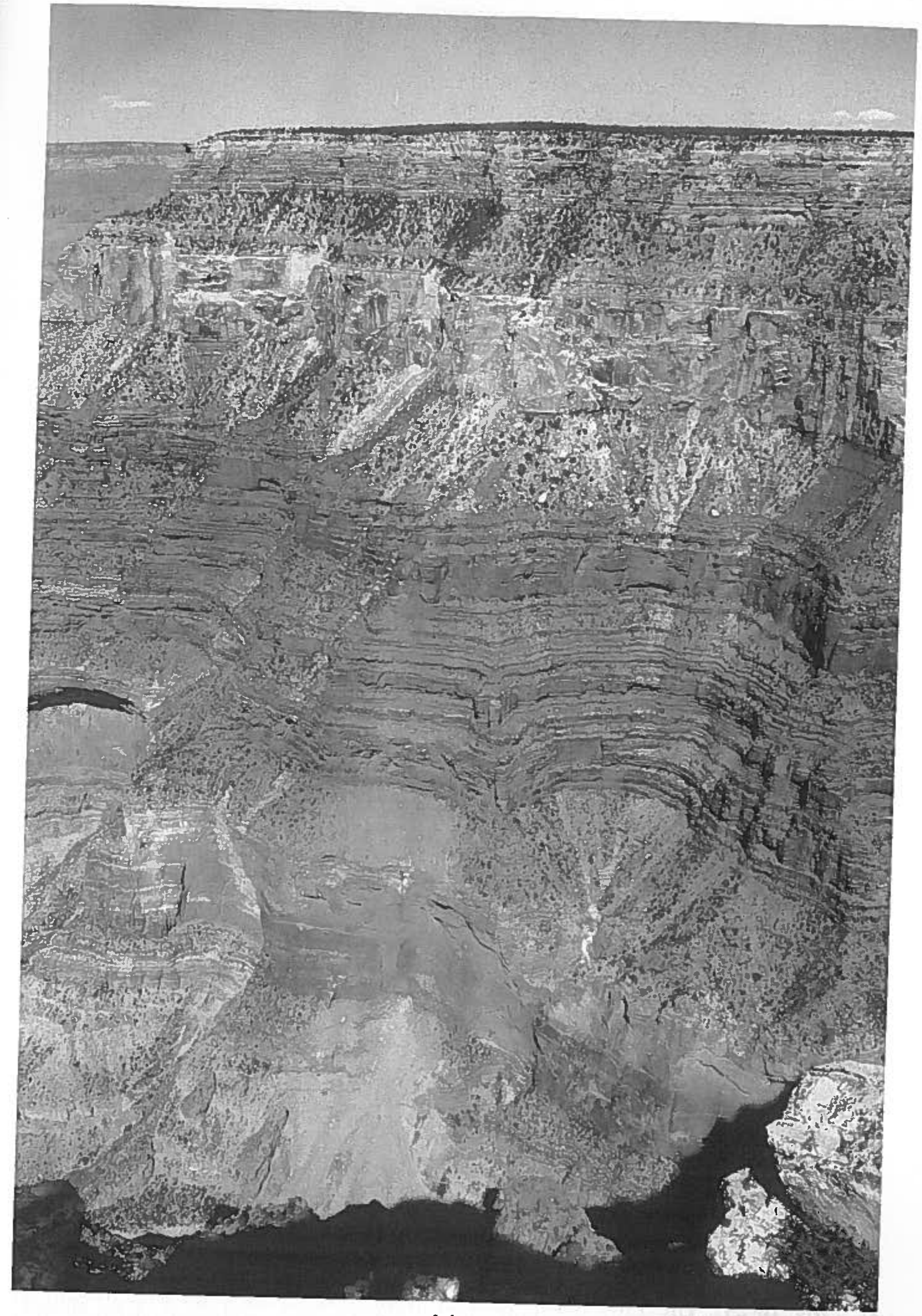
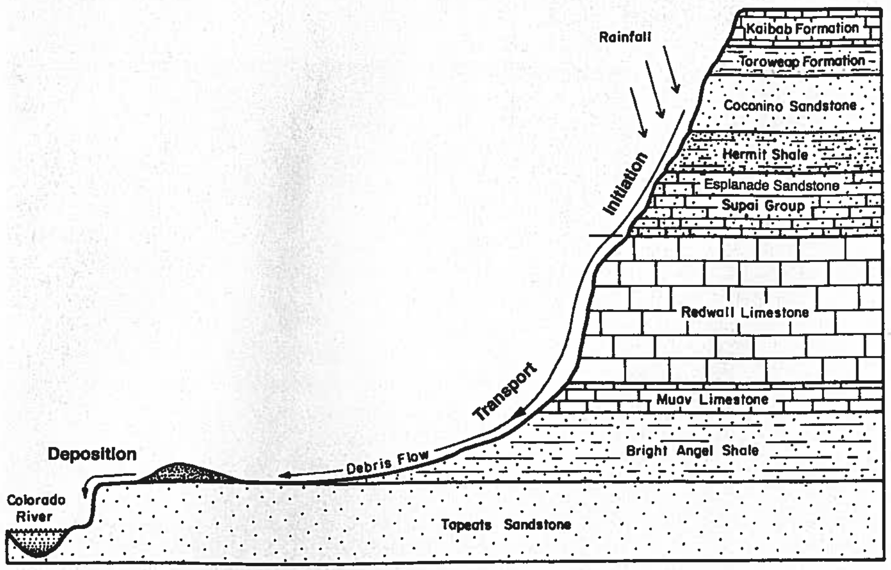


FIGURE 20.1. Longitudinal profile of the Colorado River in upper Marble Canyon showing the relation between the height of the Hermit Shale and rapids.



(a)

FIGURE 20.2. (a) The initiation area for debris flows in Monument Creek (Mile 93.5) consists of the interbedded Supai Group, primarily sandstones, siltstones, and shales (middle of the photograph), overlying the near-vertical Redwall Limestone and underlying the Hermit Shale. The 1984 debris flow was initiated in the Hermit Shale and in the Esplanade Sandstone Member of the Supai Group (Webb et al. 1988; photo by R. Webb). (b) Schematic diagram illustrating the initiation of debris flows by the failure of bedrock—usually the Hermit Shale and (or) members of the Supai Group—during intense rainfall.



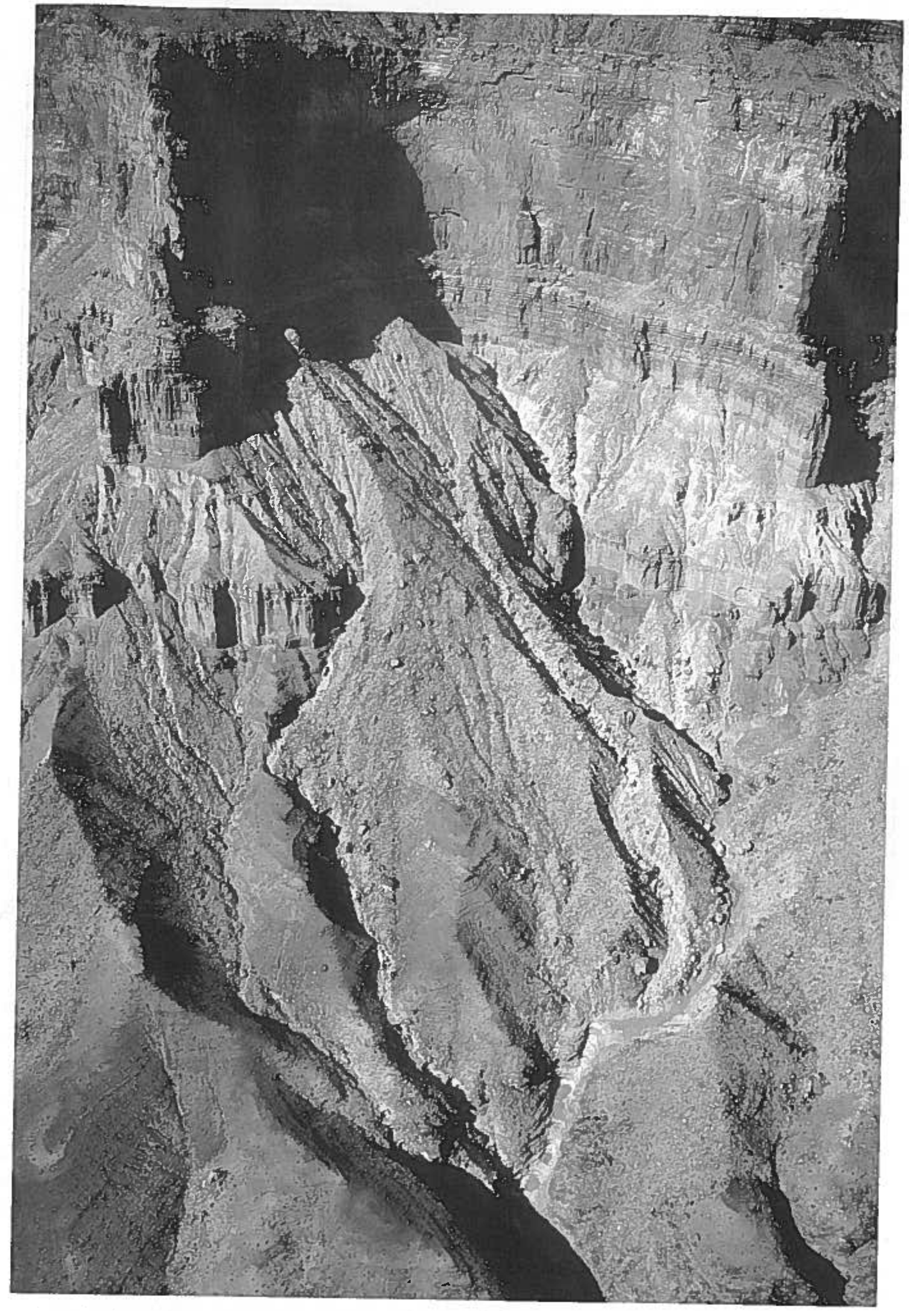
(b)

FIGURE 20.2. *(Continued)*

shales above the river and the presence of rapids, an indicator of debris flows. In the first 20 miles of river downstream from Lees Ferry, only five rapids are present. Only two of these—Badger and Soap Creek Rapids—are significant navigational hazards. The debris flows that created these rapids came from shales, which were deposited in a fluvial environment, of the Chinle Formation at the base of the Vermillion Cliffs. Downstream from about Colorado River mile 15, the Chinle Formation is not close enough to the river corridor to produce debris flows that reach the Colorado River. By mile 20, the Hermit Shale, which first appeared at river level at about mile 6, has risen over 300 m above the river. With higher elevation comes the potential energy necessary to mobilize falling debris into slurries. The result is more debris flows: Seven rapids and numerous riffles are in the Colorado River between mile 20 and 27, and the slopes above the river are strewn with loose boulders.

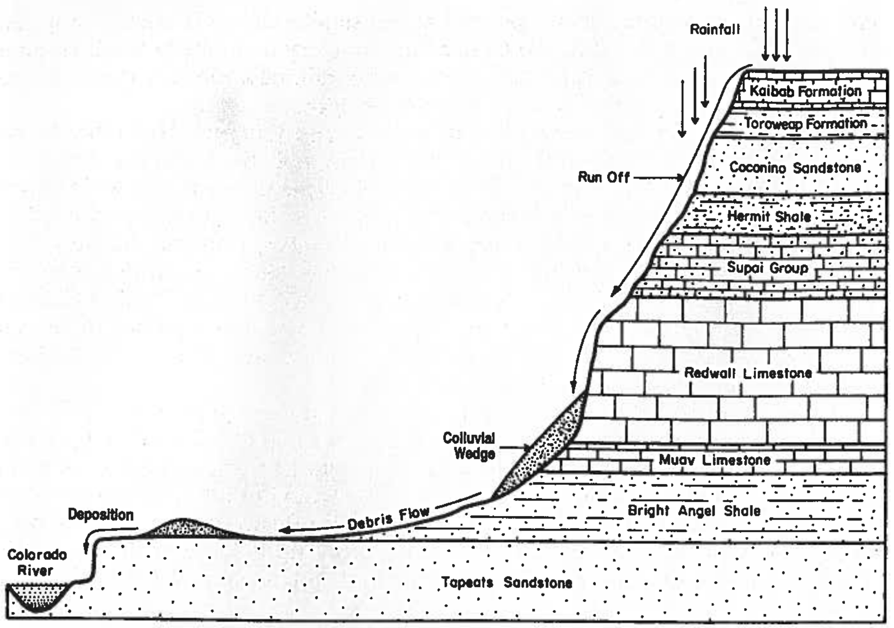
Bedrock failures (Fig. 20.2) are associated with 13% of historic debris flows in Grand Canyon. These failures are responsible for many of the largest flows of the last century (Melis et al. 1994; Griffiths et al. 1996, 1997). In December 1966, 11 slope failures in the Hermit Shale, Supai Group, and Muav Limestone contributed to the debris flow in Crystal Creek (Cooley et al. 1977). These failures occurred 1400–2000 m above and 20 km from the Colorado River. Of 93 slope failures throughout Grand Canyon that resulted from the storm in December 1966, 70 percent were in the Hermit Shale and Supai Group (Webb et al. 1989, p. 24). The 1984 debris flow in Monument Creek (Fig. 20.2) began as an avalanche from the Esplanade Sandstone of the Supai Group and flowed 5 km in its 1000-m descent (Webb et al. 1988).

Most debris flows in Grand Canyon result from failure of colluvial wedges at the base of Redwall Limestone cliffs (Fig. 20.3). The boulders and cobbles in these colluvial wedges are mostly derived from the Redwall Limestone, Kaibab Formation, and Supai Group sandstones; sand, silt, and clay are contributed by the Hermit Shale and fine-grained strata of the Supai Group. This poorly sorted mixture makes an ideal source for debris flows. Two types of failures begin in



(a)

FIGURE 20.3. (a) Failure scars caused by the "firehose effect" along gullies through colluvial wedges at Mile 62.5. The vertical cliffs at the top of the photograph are Redwall Limestone; the colluvial wedges overlie Muav Limestone and Bright Angel Shale. (Photo by T. Melis.) (b) Schematic diagram illustrating the initiation of debris flow by colluvial wedge failure during intense rainfall.



(b)

Figure 20.3. (*Continued*)

colluvial wedges. Runoff from intense rainfall erodes gullies into the colluvial wedges, where more and more material is entrained until a slurry forms. This type of direct failure has contributed to 20% of historical debris flows in Grand Canyon. The firehose effect (Johnson and Rodine, 1984) involves water pouring off cliffs onto colluvial wedges, causing failures that contributed to 37% of historical debris flows in Grand Canyon (Griffiths et al. 1997).

Debris flows initiated in colluvial wedges are small and flow only short distances, except in the case of failures caused by the firehose effect. The 1987 flow in 18-Mile Wash was caused by a flash flood cascading over a 100-m cliff in the Kaibab Formation and falling onto an unstable slope of Hermit Shale and strata of the Supai Group. In 1990, runoff cascading over the Redwall Limestone onto colluvial wedges overlying slopes of Muav Limestone caused several debris flows between mile 62 and 64 (Melis et al. 1994). Flood waters in Prospect Canyon, a large drainage that meets the Colorado River at Lava Falls Rapid (mile 179.4), must pour over a 325-m cliff onto a scree slope of loose basalt boulders and other colluvium, which then can mobilize and flow a short 1.6 km to the Colorado River. Prospect Canyon yielded 6 debris flows between 1939 and 1995, all initiated by the firehose effect (Webb et al. 1996).

Transport

In the vicinity of the South Rim, the source areas for debris flows are 1000–1300 m above and many kilometers south of the Colorado River. Resistant bedrock units, such as the Redwall Limestone, form bedrock-floored canyons that act as confining chutes for the moving debris, allowing long-distant transport with minimal loss of mass by deposition. Most side canyons preserve little evidence of past debris flows except an occasional lateral levee in a wide section of the tributary canyon or boulder fields deposited where smaller debris flows had stopped short of the river. Debris flows in Grand Canyon carry a relatively small volume of material delivered in a large pulse; without confining channels, debris flows would lose most of their volume to levee deposition.

Debris flows can move long distances in Grand Canyon. The 1966 debris flow in Crystal Creek flowed 21 km from its initiation points to the Colorado River (Cooley et al. 1977; Webb et al. 1989). Prehistoric debris flows, which have left depositional evidence in Shinumo and Kanab Creeks, could have flowed as far as 40 km. Most debris flows travel shorter distances, primarily because few tributary canyons are longer than several kilometers. More typical debris flows, like the 1984 flood in Monument Creek and the 1987 flood in 75-Mile Canyon (Melis et al. 1994), travel 3–6 km. The exact mechanism of how debris flows can travel such distances is not well understood (Johnson and Rodine 1984; Costa 1984) but is currently under intense study.

Three types of debris flows occur in Grand Canyon. Melis et al. (1997) reported hypothetical hydrographs developed from the combination of stratigraphic analyses of debris-fan and channel-margin deposits and reconstruction of peak discharges. Type I flows consist of a single unsustained pulse of debris flow followed by recessional streamflow (Fig. 20.4). Type I flows often occur in Grand Canyon and typically are the smallest debris flows. The 1990 debris flows at mile 62.5 and 62.6 are examples of this type of runoff (Melis and Webb 1993; Melis

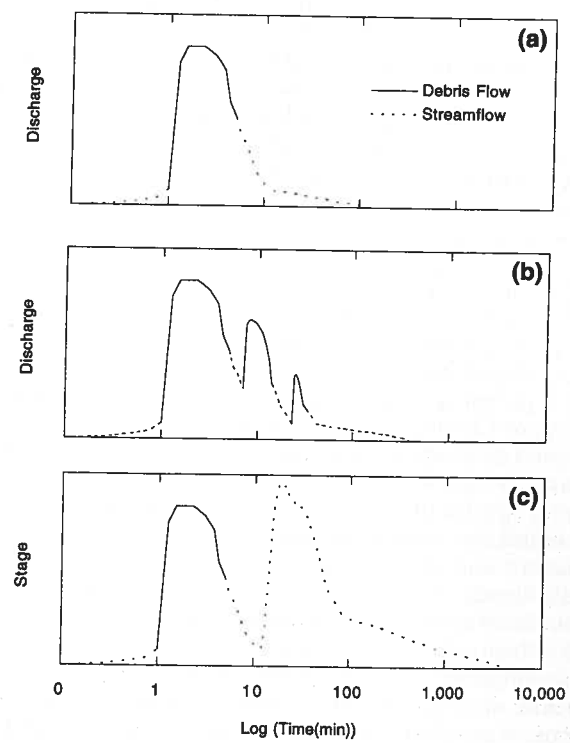


Figure 20.4. Generalized hydrographs of three types of debris flows in Grand Canyon reconstructed from depositional evidence in channels and on debris fans. (a) Type I flow. (b) Type II flow. (c) Type III flow. (From Melis et al. 1997.)

et al. 1994). Type II flows consist of a main debris-flow pulse, followed by alternating pulses of streamflow and debris flow. Depositional evidence on debris fans provides the basis for this type of hydrograph; the 1987 debris flow at 18-Mile Wash and the 1989 debris flow in a left tributary at 127.6 mile are excellent examples. Type III flows consist of a debris-flow pulse followed by a higher-stage streamflow that removes most of the debris-flow evidence (Fig. 20.4). The streamflow discharge is not necessarily higher than the debris-flow discharge, because debris-flow deposition may raise the channel bed significantly, allowing smaller streamflow discharges to erase the evidence of larger debris flows (Melis et al. 1997). The 1989 floods in Fossil and Forster Canyons are good examples of type III flows (Melis et al. 1994).

Peak discharges for debris flows in Grand Canyon range from 100 to 1000 m^3/s (Melis et al. 1994). These discharges are well within the known realm of debris-flow size (Pierson and Costa 1987). Using the total depositional volume, Webb et al. (1988, 1989) and Melis et al. (1994) showed that the duration of debris-flow pulses ranges from 30 seconds to 3 minutes. Peak velocities for debris flows range from 2 to 10 m/s (Melis et al. 1994), and channel slope is not the most important factor. Higher velocity does not result in higher discharge. For example, the Crystal Creek debris flow of 1966, which had a discharge of about 310 m^3/s, had velocities that ranged from 3 to 6 m/s; other debris flows in steep-angled chutes with discharges of about 100 m^3/s had velocities closer to 7 m/s.

Peak discharges of debris flows are similar to those of post-dam flows in the Colorado River except that debris flows typically contain 70–90% sediment. Water content of the 1966 debris flow in Crystal Creek was about 30 percent by weight; therefore, the peak discharge contained about 220 m^3/s of sediment. Melis et al. (1994) found that the water content of 20 debris flows reconstituted in the laboratory ranged from 5 percent to nearly 40 percent, depending on the amount of silt and clay carried in the debris flow. The variability in water content for a given flow is 1–5 percent; in other words, water content varies considerably among debris flows, but for any one debris flow the water content falls within a fairly precise range.

The particle-size distribution of sediment carried in debris flows shows a very wide range of sizes (Fig. 20.5). Gravel and cobbles (2- to 256-mm diameter) constitute the largest amount of debris flows (50–90 percent); 10–25 percent of a typical debris flow is sand (Melis et al. 1994). The clay content, which is critical for long-distant transport, ranges from <1% to 5%. Although boulders typically comprise <20% of debris flows, they are the most visible and important aspect of debris-flow transport (Beatty 1989; Rodine and Johnson 1976), particularly with respect to Colorado River geomorphology.

As yet, the exact mechanism of boulder transport in debris flows is unknown, but several hypotheses have been advanced. A large part of the mechanism is buoyancy, or more specifically the difference in density between fluid and rock. The density of sandstones, limestones, and basalts in Grand Canyon ranges from 2600 to 2700 kg/m^3, and debris flows have densities between 2200 and 2400 kg/m^3. The small density difference indicates that buoyancy forces should provide considerable lift to boulders in a moving debris flow. Despite this, boulders still should sink, albeit slowly. Another upward force on boulders is created by particle collisions in the slurry. In the jostling among all those buoyant boulders, collisions may cause sinking particles to move upward and (or) ascending particles to sink. In pure water, forces are transmitted only in the fluid. In debris flows, forces are transmitted both in the fluid and by particle-to-particle collisions.

Early models treated debris flows as deformable plastic moving as plug flow with little internal deformation (Johnson and Rodine 1984). Plug flow models,

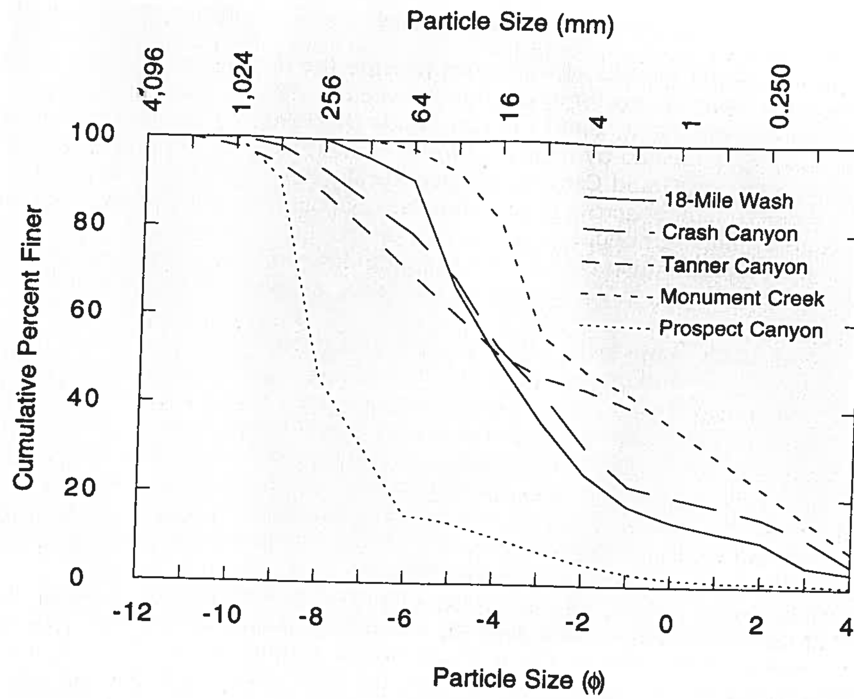


FIGURE 20.5. Particle-size distributions for selected debris flows in Grand Canyon. 18-Mile Wash, the 1987 debris flow at 18-Mile Wash; Crash Canyon, the 1990 debris flow at Mile 62.6; Tanner Canyon, the 1993 debris flow at Tanner Canyon, Mile 68.5; Monument Creek, the 1984 debris flow in Monument Creek (Mile 93.5); and Prospect Canyon, the 1995 debris flow in Prospect Canyon (Mile 179.4).

also known as Bingham models, require the assumption that below a certain point, called the yield stress, debris flows could withstand pressures without deforming. One indication of yield stress is critical thickness; when a debris flow stops, it has a finite thickness that is a function of yield stress. When slurries move in straight channels, most of the mass flows as a plug with little internal deformation (shearing). The only shearing in the flow is assumed to occur very close to the sides and bottoms of the flow. Using the plug-flow model, with no internal shear and a finite yield stress, rocks could bob along in the flow primarily as a result of buoyancy forces.

The problem is that debris flows moving in natural channels have considerable internal shear, although not necessarily as much as turbulent water flow. Something else is helping to support those boulders. One suggestion of what that force might be came from experiments in which landslides were created in a flume (Iverson and LaHusen 1989). High-frequency fluctuations in pressure, on the order of one cycle per second, occurred in the interstitial fluid at the base of the landslide. These pressures could have occurred for many reasons; they have resulted from flow over a rough bed like the bottom of a typical Colorado River tributary. The pressures generated were of sufficient magnitude to support the media; in other words, transient pressures generated within the fluid could provide the supporting forces for boulders.

Deposition

Debris fans at the mouths of tributaries provide the depositional site for most debris flows and a setting for interaction between fluvial and hillslope processes. Debris fans create most Grand Canyon rapids (Howard and Dolan 1981); only two rapids were created by rockfalls (Webb et al. 1988). Melis (1997) described 444 debris fans in Grand Canyon that periodically have debris-flow deposition. The tributary channel across most debris fans is confined, not by bedrock, but by levees of mud and boulders left by past debris flows. Terraces on the higher parts of debris fans, untouched by the Colorado River in recent years, are clearly of debris-flow origin, indicating earlier Holocene debris-flow activity (Hereford et al. 1997a).

Some debris flows lack sufficient mass and energy to reach the Colorado River. Their mass and energy are expended during deposition of the confining levees on the debris fan. Type I debris flows typically create a uniform depositional area on part of the debris fan, whereas type II flows create a complex array of debris-flow and streamflow deposits (Melis et al. 1997). The surfaces of recent deposits are rough with pressure ridges, which reflect pulses in debris flow (Major 1997), and levees. Boulder trains, which are reworked levees, are also common. Only residual boulder trains (some very large) and streamflow deposits mark deposition by type III debris flows.

Most debris flows reach and constrict the river to some extent, although the effect of debris flows on the mainstem varies with reach morphology, particularly channel width (Melis 1997). No twentieth century debris flow has fully dammed the river or flowed downstream; the 1939 debris flow from Prospect Canyon constricted Lava Falls Rapid by about 80 percent, whereas the 1966 debris flow from Crystal Creek had a similar effect on Crystal Rapid (Webb et al. 1997). Debris flows lose too much energy and mass on debris fans; in addition, the turbulent water in rapids dilutes the snout, further reducing its momentum. In most Grand Canyon rapids, the largest boulders are at the head, which is also the closest point to the mouth of the tributary canyon. Boulders near the bottom of rapids are generally pushed there by the force of the Colorado River.

Some of the boulders deposited in the Colorado River are enormous. At mile 62.5, one boulder transported into the river in 1990 weighs 280 Mg. Boulders transported in some debris flows that occurred after 1983 range between 70 and 100 Mg (Melis et al. 1994). In Crystal Creek, a boulder weighing 45 Mg did not reach the Colorado River in 1966, and larger ones deposited in the river are what increased the severity of Crystal Rapid. The size of boulders transported is not necessarily related to the discharge of debris flows; boulder size varies tremendously among debris flows of about the same size (Webb et al. 1989).

FREQUENCY OF DEBRIS FLOWS

Estimating the frequency of debris flows requires an approach that is considerably different from traditional flood-frequency analysis. Until the last few decades, no one recorded debris flows in most of Grand Canyon. The history of debris flows must be reconstructed from their deposits or their effects on the Colorado River. Several methods are available, but most have limitations. For example, radiocarbon dating of organic materials is the standard way to determine the age of sediments deposited during the last 40,000 years, but very little organic material was deposited with the mud and boulders. Under certain circumstances, such as association with archaeological sites, debris flows can be dated using radiocarbon analysis (Hereford et al. 1996). Using twigs extracted from mud plastered against an overhanging wall, Melis et al. (1994) dated a 5400 yr B.P. debris flow at river mile 63.3. Limestone particles dissolve with time, and the depth of pitting in limestone boulders can be used to date prehistoric debris flows (Hereford et al. 1997a). However, development of measurable pitting requires several centuries, rendering the technique useless for twentieth century debris flows.

Repeat photography is the best technique for documenting historical debris flows (Webb 1996). Before closure of Glen Canyon Dam in 1963, floods on the Colorado River eroded away most newly deposited debris within five years. Residual accumulations of boulders remain that are obvious in repeat photographs (Fig. 20.6). Repeat photography can also reveal the number of debris flows from a tributary in the last century, but only under certain conditions. A sequence of historical photographs can be used to reconstruct a history of debris-flow activ-

(a)

Figure 20.6. Repeat photographs of Boucher Rapid (mile 96.7). (a) (February 8, 1890; R.B. Stanton). Historical evidence of debris flows in Grand Canyon largely comes from repeat photography (Webb 1996). Boucher Rapid is a moderate-sized rapid in the Inner Gorge that has had several debris flows in the last century. In this 1890 view, fresh-looking gravel in the channel of Boucher Creek suggests that a flash flood, but probably not a debris flow, had occurred a short time before this photograph was taken. (b) (February 18, 1992; R. Webb). Boucher Rapid is not considered a very formidable reach of whitewater today. The debris flow of 1966 at Crystal Rapid, only 2 km downstream, raised the river level sufficiently to drown out the tailwaves of Boucher Rapid (Péwé 1968). Similarly, Boucher Creek had a debris flow in the early 1950s that drowned out the tailwaves of Hermit Rapid, about 2 km upstream. The debris flow in Boucher Creek caused deposition of boulders over the entire debris fan. Boulders 1–2 m in diameter were displaced, and new ones of about the same size were deposited.

(b)
Figure 20.6. *(Continued)*

ity, such as in Prospect Canyon at Lava Falls Rapid (Webb et al. 1996, 1997). A single historical view can also be used if it was taken from a vantage point that looks down on a debris fan. In this case, changes in the depositional patterns on debris fans may indicate many past debris flows.

Melis et al. (1994) discriminated 525 tributaries of the Colorado River between Lees Ferry (mile 0) and Diamond Creek (mile 225) that periodically yield debris flows; Griffiths et al. (1996) extended the definition downstream to mile 246 to include 600 tributaries. The debris fans of 164 of these tributaries are visible in 483 photographs taken after 1872 (Webb 1996; Griffiths et al. 1996). By examining changes in the arrays of boulders on the debris fans, we found 96 tributaries that had one or more debris flows in the last century, or 56 percent of all debris fans recorded in the repeat photography. One half of the 63 largest tributaries (>10 km^2) had at least one debris flow in the last century. We also observed debris flows in 17 small chutes or gullies that otherwise are insignificant.

Using logistic-regression analysis of the data obtained from the repeat photography, Griffiths et al. (1996) estimated the probability of debris-flow occurrence throughout Grand Canyon (Fig. 20.7). Because the physiography of western and eastern Grand Canyon are considerably different, Griffiths et al. (1996) divided the canyon at Hermit Rapid (mile 95.0) and modeled each half separately. A calibration model was developed using the repeat photography information from 164 tributaries. A verification model was also developed from 214 tributaries that included those used in the calibration model combined with additional tributaries with frequency information obtained by methods other than

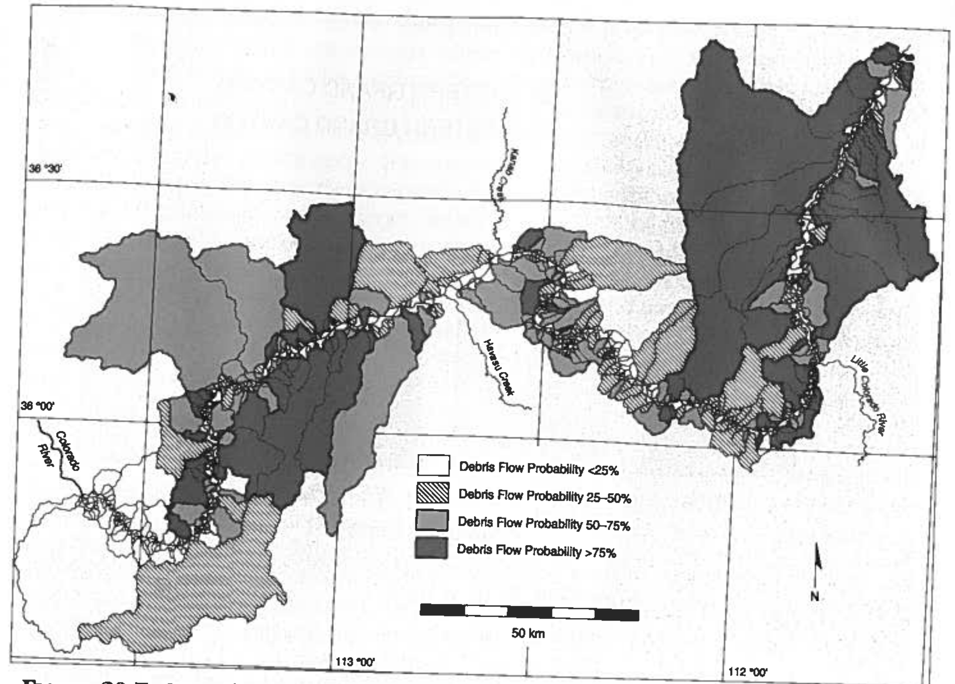


Figure 20.7. Map showing four probability classes of debris-flow frequency during the last century in 600 tributaries of the Colorado River in Grand Canyon. (From Griffiths et al. 1996.)

repeat photography. Griffiths et al. (1996) used 21 morphometric and lithologic variables that were suspected of influencing debris-flow frequency.

Based on the calibration data, the logistic-regression model of Griffiths and others (1997) had an accuracy of about 70% in eastern Grand Canyon and 74% in western Grand Canyon. Of the morphometric variables, tributary drainage area and channel gradient, the aspect of the river corridor at the mouth of the tributary, and the channel gradients to the Hermit Shale and Muav Limestone were statistically significant and were used to estimate frequency. The resultant map of probabilities of debris flows during the last century in Grand Canyon (Fig. 20.8) indicates that debris flows are more frequent in eastern than in western Grand Canyon. The highest frequency of debris flows has occurred between the Little Colorado River and Hance Rapid (mile 61.5 to 77). The reach between Havasu Creek (river mile 157) and Lower Granite Gorge (river mile 213) had the lowest frequency of debris flows. Debris-flow frequency is clustered in distinct reaches along the Colorado River, particularly where the river's course is southwesterly and where Hermit Shale is between 100 and 1000 m above and relatively close to the river.

For individual tributaries, debris flows recur every 10–50 years, on average, in 60 percent of the tributaries of the Colorado River. Some tributaries had no debris flows during the last century; Prospect Canyon has had six debris flows clustered between 1940 and 1995, but none in the preceding 50 years (Webb et al. 1996, 1997). A large flow may destabilize enough sediment to supply additional debris flows until the supply is exhausted. Another major slope failure oc-

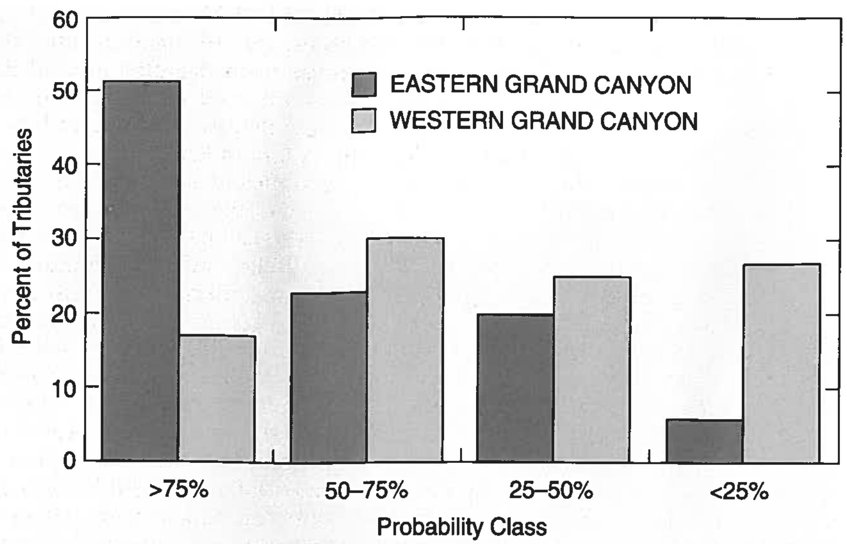


Figure 20.8. Histograms of the probability of debris-flow occurrence in eastern and western Grand Canyon. (From Griffiths et al. 1996.)

curs—perhaps a half century later, perhaps centuries later—and the process repeats itself. Some tributaries, such as 217-Mile Canyon, do not appear to have experienced a debris flow for perhaps several centuries. Whereas these tributaries have small floods on a frequent basis, debris flows are rare. The notable exception is Prospect Canyon, one of the most active debris-flow producers in Grand Canyon (Webb et al. 1996, 1997).

THE GEOMORPHIC FRAMEWORK OF THE COLORADO RIVER

River runners and scientists have long recognized parts of the association between tributary canyons, whitewater, and sand bars (Howard and Dolan 1981; Schmidt and Graf 1990). Major John Wesley Powell first recognized that the boulders in rapids were deposited during tributary floods (Powell 1876); other early explorers also noticed the close association of tributaries and rapids. But Powell believed that differential erosion of bedrock somehow affected the river; his crew learned to dread the appearance of granite and limestone along the river because it indicated, so they thought, intense rapids ahead. They were wrong. The lithology at river level affects mostly channel width, not slope (Hamblin and Rigby 1968; Howard and Dolan 1981). However, channel width dramatically affects debris fan stability and evolution, including fan shape, size, spacing, and composition (Melis 1997).

Piles of boulders in the river channel create hydraulic controls on the Colorado River. Debris flows transport boulders from the cliffs to the river; river floods transport much of the newly deposited material downstream where it accumulates in islands, rock gardens, or debris bars (Webb et al. 1989, 1997a; Melis 1997). This process is called reworking. What remains after repeated debris flows from the tributaries and reworking by the Colorado River is a debris fan, typically comprised of boulders on the surface (Melis and Webb 1993); a rapid full of large boulders that form various waves and holes; and an island or other deposit of well-sorted cobbles or boulders just downstream. Mantled around the debris fan on its upstream and downstream sides are sand bars, typically the best camping beaches in Grand Canyon. Debris-flow deposits, rapids, and sand bars make up the geomorphic framework of the Colorado River.

Rapids exist because the Colorado River lacks sufficient stream power to remove all boulders deposited from debris flows. Most of the sediment deposited by debris flows is removed; typically, the river will erode all mud and sand whenever the debris fan is inundated, and most of the cobbles and small boulders in the river are transported away. Whereas clay, silt, and sand may be transported a long distance from the source debris fan, the eroded cobbles and boulders are transported only a short distance downstream. Just below the rapid, the cobbles and small boulders accumulate in neatly sorted piles. These deposits, properly termed debris bars and informally called "islands" and "rock gardens" (Kieffer 1985, 1987), form secondary riffles and rapids. These stretches of whitewater are known informally as the "Son of . . ." rapids (e.g., Son of Hance) in recognition of their genetic link to the parent rapid. Howard and Dolan (1981) refer to well-sorted, coarse-grained deposits as cobble bars. Cobbles, which have diameters between 64 and 256 millimeters, are not the only constituents of these bars; sand and boulders also are present. Gravel bars are common between river mile 65 and 72, and the only exposed rock garden is downstream from Crystal Rapid (Kieffer 1985). The majority of reworked debris fans in Grand Canyon consist of Redwall and Muav Limestone boulders (Melis 1997).

Debris fans, rapids, and sand bars are all controlled, directly or indirectly, by the process of debris flows. Debris fans that are reworked remain relatively stable until modified by additional debris flows (Melis 1997). In terms of navigation, rapids change with each change in water level of the river, but the underlying configuration of boulders that create the waves and holes change little until the next debris flow. Rock gardens and islands remain until the next debris flow and reworking flood in the Colorado River. Finally, the sand bar, the least stable element of this geomorphic framework, can change with fluctuations in pre-dam flow or the pattern of flow release from Glen Canyon Dam. But the locations of most sand bars are fixed relative to piles of boulders, and the general configuration of sand bars is dependent on when the last debris flow occurred.

The presence of debris fans and rapids has broad implications as to whether a bedrock canyon is downcutting or not. The rapids account for about 10 percent of the length while causing 50 percent of the fall in the Colorado River through Grand Canyon (Leopold 1969). In the relatively warm and dry climate of the present, the Colorado River likely is expending its erosive energy on removing boulders from the rapids instead of eroding downward through bedrock.

Since completion of Glen Canyon Dam in 1963, the delicate balance between tributaries and river has shifted in favor of aggrading debris fans and narrower rapids. Concern about the decreased competence of the regulated Colorado River was raised shortly after the dam was completed (Péwé 1968). Graf (1980) concluded that debris fans were aggrading in bedrock canyons of the Green River in Utah. Reportedly, 25 percent of debris fans in Grand Canyon aggraded between 1965 and 1973 (Howard and Dolan 1981). Our observations and repeat photography (Melis et al. 1994; Webb 1996) suggest that this is too high a percentage for such a short period of time, but debris fans are aggrading at even the largest rapids (Webb et al. 1996).

The depth of the Colorado River is not controlled by resistant bedrock (at least not during the Holocene). Instead, the 444 debris fans (not all of the 525 tributaries have debris fans at their mouths) along the length of the river corridor inhibit further downcutting. The energy of flowing water is expended on removing all the massive boulders thrown into it by minuscule tributaries, not eroding the underlying bedrock. It is a titanic struggle between small tributaries and a large river: Tributaries push boulders in, and the river tries to transport them downstream or dissolve them. The tributaries currently are winning, because rapids are present along the river corridor. Now, operation of Glen Canyon Dam is accelerating that aggradation of boulders and cobbles (Webb et al. 1997a).

SUMMARY

Debris flows occur in 525 tributaries of the Colorado River in Grand Canyon between Lees Ferry and Diamond Creek (river miles 0 to 225). An episodic type of flash flood, debris flows transport poorly sorted sediment ranging in size from clay to boulders into the Colorado River. Debris flows in Grand Canyon are initiated by slope failures that occur during intense rainfall. Failures in weathered bedrock, particularly in the Hermit Shale and Supai Group, have initiated many historic debris flows in Grand Canyon. A second mechanism, termed the "firehose effect," occurs when runoff pours over cliffs onto unconsolidated colluvial wedges, triggering a failure.

Interpretation of repeat photographs spanning 125 years yielded information on the frequency of debris flows in 168 tributaries. Of these, 96 contain evidence of debris flows that have occurred since 1872, whereas 72 tributaries have not had a debris flow during the last century. The oldest, dated debris flow occurred 5400 ^{14}C years ago at mile 63.3. The frequency of debris flows ranges from one every 10 to 15 years in certain eastern tributaries, to less than one per century in 40 percent of the tributaries. Debris flows are more frequent in Marble Canyon and eastern Grand Canyon than in western Grand Canyon.

Debris flows in Grand Canyon have three types of hydrographs. Type I events consist of a single, unsustained pulse of debris flow, whereas type II events have multiple debris-flow pulses. Type III events consist of a debris flow followed by a streamflow flood of larger stage. Although peak discharges of most debris flows range from 100 to 300 m^3/s, the largest debris flow in Grand Canyon during the last century—the 1939 debris flow in Prospect Canyon—had a peak discharge of about 1000 m^3/s. The water content of debris flows ranges from 10 to 25 percent.

Debris flows create and maintain debris fans and the hundreds of associated riffles and rapids that partly control the geomorphic framework of the Colorado River downstream from Glen Canyon Dam. Before regulation, debris fans aggraded by debris flows were periodically reworked by large river floods. Operations of Glen Canyon Dam have reduced flood frequency in the Colorado River, which has limited reworking of recently aggraded debris fans. The presence of rapids indicates that the bed of the Colorado River is aggrading, not eroding through bedrock, in the Holocene.

• 21 •

SIDE CANYONS OF THE COLORADO RIVER IN GRAND CANYON

Andrew R. Potochnik and Stephen J. Reynolds

INTRODUCTION

The majesty of the Grand Canyon emanates from the magnitude of its impressive dimensions. The Colorado River has cut an exquisite gorge; its tributaries are no less spectacular. Trails from the rim are few, the hike is lengthy, and access to much of the canyon from above is limited; however, a river trip through the Grand Canyon, with ample time for hiking, enables one to explore many side canyons with relative ease. Unlike a view from the rim, a view from the river provides an inside-out perspective (Fig. 21.1). The rim view paints an overall picture of the grand-scale geological scene. In the side canyons, it is possible to examine the details of geological features and relationships that tell a more complete story. The sequence of rocks in the Grand Canyon can be divided into four groups, each separated by a major unconformity (Fig. 21.2). The oldest group includes Proterozoic igneous and metamorphic rocks (e.g., Vishnu Group) that were formed during a major episode of deformation and metamorphism about 1.7 billion years ago (Ilg et al. 1996). These rocks are overlain by the Grand Canyon Supergroup, a series of tilted sedimentary rocks and interlayered mafic flows and sills that formed between 1.3 and 0.8 billion years ago (Larson et al. 1994; Elston and McKee 1982). Extensive erosion of these Proterozoic rocks formed a conspicuous, regionally planar unconformity on which the canyon's characteristic, horizontally stratified Paleozoic rocks were deposited between 550 and 250 million years ago. The fourth group of rocks in the canyon includes a veneer of late Cenozoic sediments and volcanic rocks that were deposited mostly since six million years ago, during and after the main period of canyon cutting. The classic physiography of the canyon and surrounding Colorado Plateau is largely a signature of Cenozoic erosion in response to post-mid-Cretaceous uplift of the region to a height of about two miles (3.2 km) above sea level.

From its source in north-central Colorado to its mouth at the Gulf of California, the Colorado River crosses four major physiographic provinces: the Rocky Mountains, the Colorado Plateau, the Transition Zone, and the Basin and Range Province. On the final leg of its long journey across the Colorado Plateau, the river follows a sinuous course in northern Arizona through the Grand Canyon. The Grand Canyon, the Transition Zone, and the Colorado Plateau end abruptly at the Grand Wash Cliffs, where the river flows into the Basin and Range Province

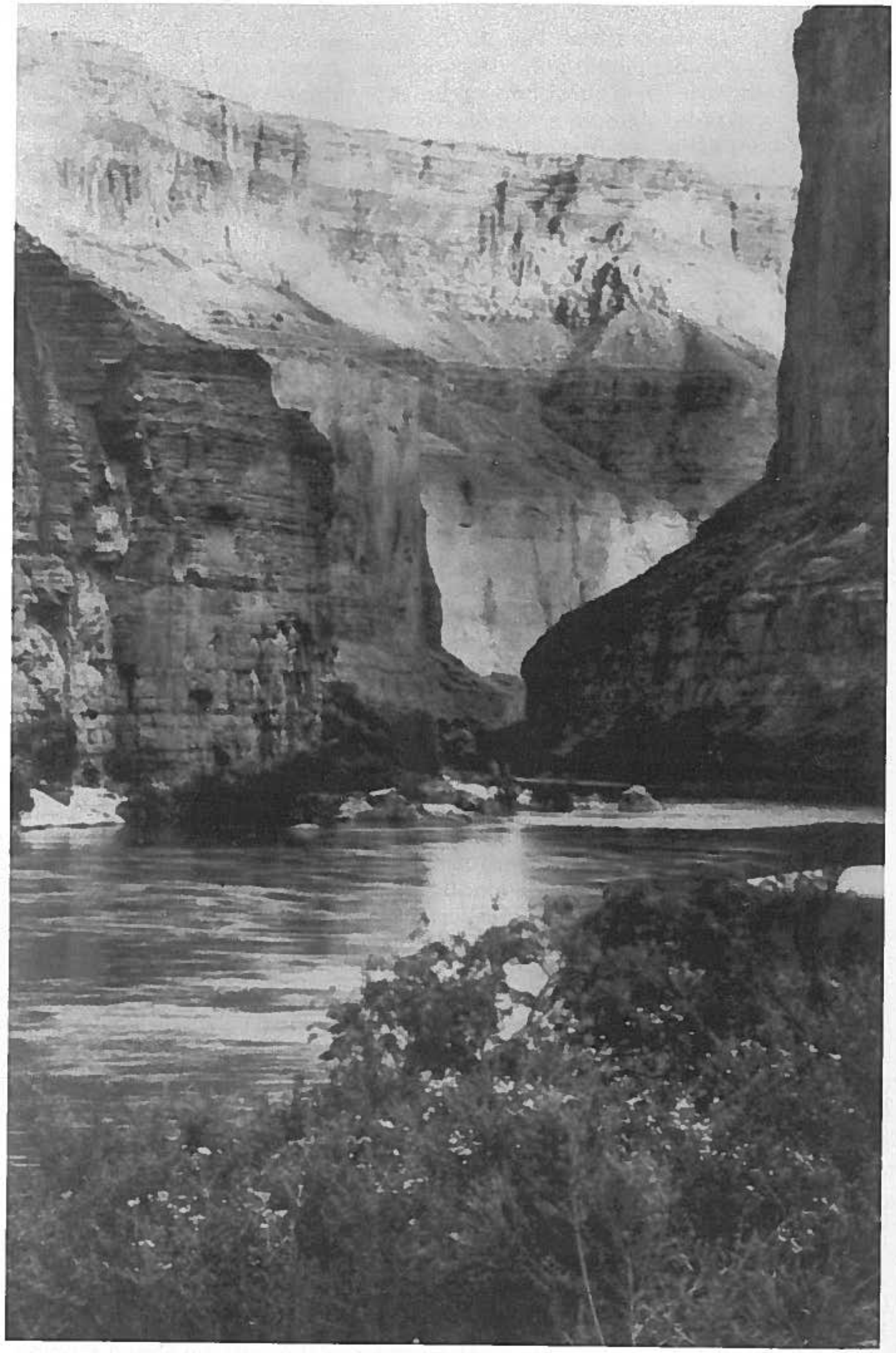


FIGURE 21.1. The Colorado River within Marble Canyon, as seen from Saddle Canyon. (Photograph by S. Reynolds.)

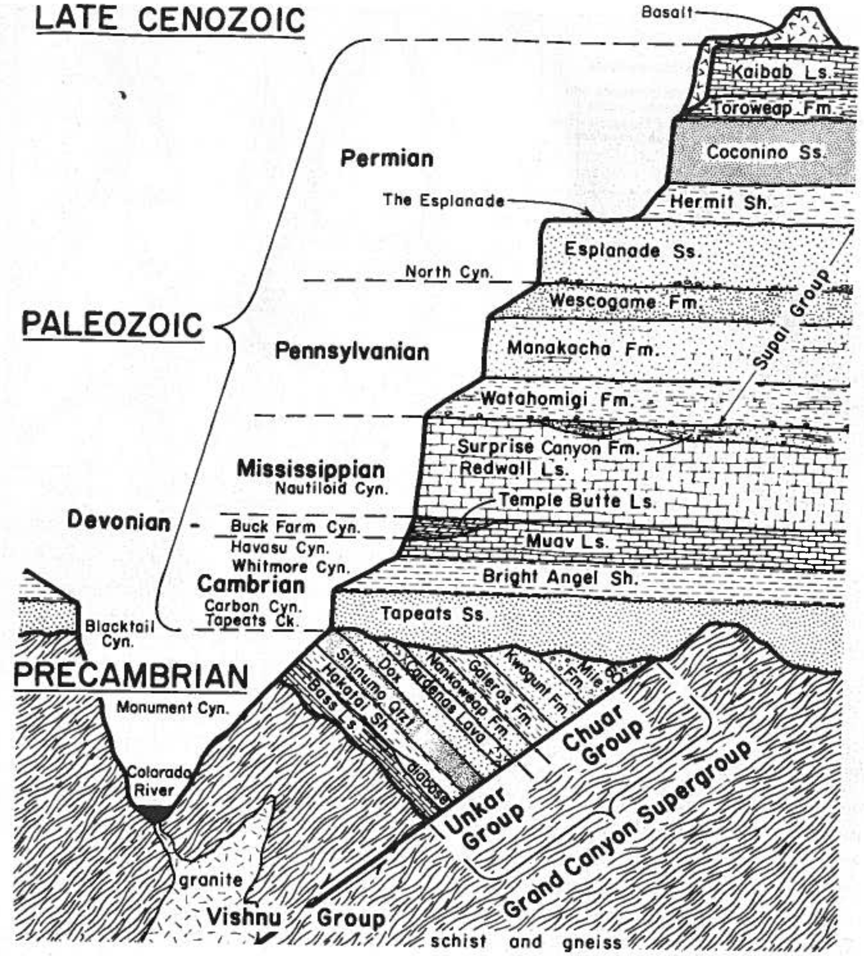


FIGURE 21.2. Generalized stratigraphic section of rock units in the Grand Canyon.

at Lake Mead. The canyon is 277 miles (444 km) long, up to 18 miles (30 km) wide, and nearly one mile (1.6 km) deep. Although many rivers in the world follow the trend of preexisting structural weaknesses, a peculiar feature of the Colorado River in the Grand Canyon is that it generally cross-cuts major north–south faults and folds and flows against the regional dip of strata through most of its length. It parallels the major structures only twice and for a mere one-sixth of its length (Fig. 21.3).

In contrast, side canyons are influenced more strongly by structural and geomorphic controls such as faults, regional dip of strata, and rock hardness. Below, we summarize these controls and show how they have resulted in the formation of segments of the Grand Canyon, each having a distinctive side-canyon morphology. Finally, we discuss geological features of some of the more interesting side canyons. This discussion is, in part, modified from Potochnik and Reynolds (1986).

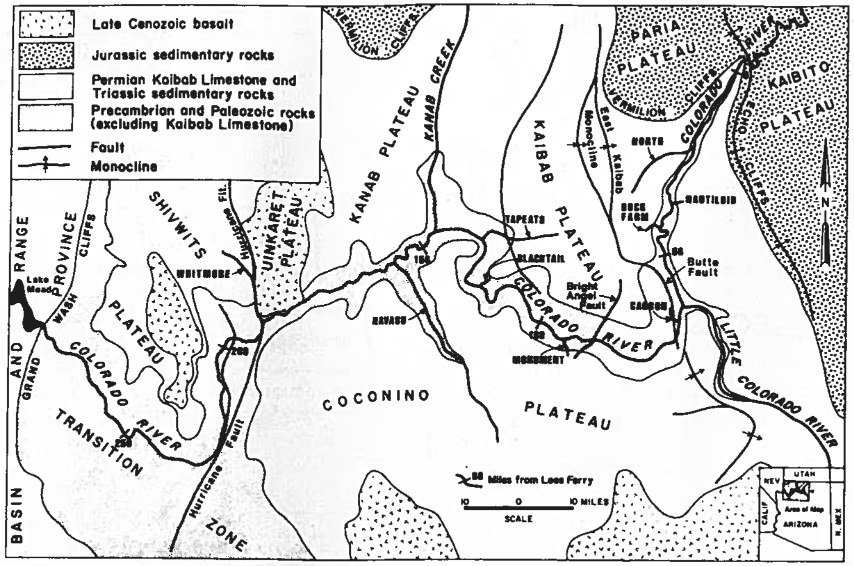


Figure 21.3. Simplified geologic map of the Grand Canyon region.

CONTROLS ON THE MORPHOLOGY OF SIDE CANYONS

The morphology of side canyons reflects various combinations of stratigraphic and structural controls, especially the regional dip of strata, folds and faults, and differential erosional susceptibility of the various rock units. Other factors influencing side-canyon morphology include climate and amount of time since the beginning of side-canyon incision.

A first-order control of side-canyon morphology is the regional or subregional dip of Paleozoic strata. Side canyons that flow down the regional dip tend to be longer and wider than those draining into the river against the dip. This is exemplified along the north rim of the canyon near Bright Angel Creek. Here, we see a more branched network of longer streams flowing with the southward dip, compared to those flowing against it from the south rim (Fig. 21.4).

A more local control of side-canyon morphology is displayed along folds and faults. Fault-controlled canyons, such as Bright Angel Creek, are recognized by distinct linear trends and longer-than-normal lengths (Fig. 21.4), attributes that reflect the easily eroded character of faulted and brecciated rock along the faults. Regional folds, such as the structurally low sag along Havasu Creek (Fig. 21.4), control the general location of some major tributaries. More structurally abrupt folds, such as the East Kaibab monocline (Fig. 21.3), cause rapid downstream changes in the morphology of a single side canyon as the stream cuts across a wide variety of rock units.

The character of side canyons is influenced strongly by differences in the erosional susceptibility of different rock units. Canyons cut in hard, resistant rocks, such as the Proterozoic Vishnu Group or the Redwall Limestone, have steep walls and narrow, V-shaped or vertical profiles. In contrast, canyons cut entirely in soft rocks, especially the easily eroded Dox Sandstone and Chuar Group, are broad and bowl-shaped, with rounded, receding walls. Most side canyons display both characteristics because of the successive alternation of relatively hard and soft rocks. A rapid lateral retreat of resistant cliffs probably begins as the underlying soft rock is exposed. As a consequence, side canyons commonly become wider once the stream has incised down through a cliff-former and into an underlying weak unit.

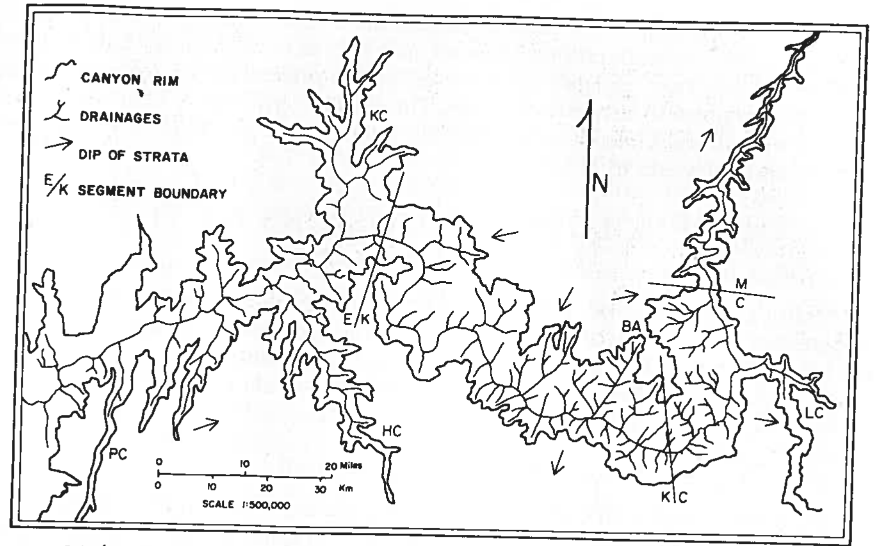


Figure 21.4. Simplified drainage-network map of the eastern Grand Canyon, showing the four segments of the canyon discussed in the text. Segments are as follows: M, Marble Canyon; C, Chuar; K, Kaibab; and E, Esplanade. Labeled tributaries are as follows: BA, Bright Angel Creek; HC, Havasu Creek; KC, Kanab Creek; LC, Little Colorado River; and PC, Prospect Canyon.

Climatic variation also causes different side-canyon morphologies. Precipitation varies considerably, according to changes in elevation across the region. Greater precipitation in uplifted regions, such as the Kaibab Plateau, provides more erosive power to streams originating in those areas. Frost action and chemical weathering also are greater in those same regions. Climatic control is best illustrated by comparing the relatively unincised Desert Facade [6000 ft (1829 m) elevation] near the mouth of the Little Colorado River to the topographically higher [7300 ft (2225 m) elevation] and more incised, pine-covered, south rim to the west. Climatic changes with time also have affected the types and magnitudes of processes responsible for incising and enlarging side canyons. For example, a change from a warmer and more humid climate in the Eocene to a colder and drier one in the Oligocene (Frakes and Kemp 1973) probably affected the rate of denudation of Mesozoic rocks from the plateau prior to any canyon cutting. Pleistocene climatic fluctuations likewise may have affected the recurrence interval of high-discharge flow events and the rates at which lava dams were scoured from the river.

Base-level control is of fundamental importance when considering the causative factors in canyon development. The local base level that controls Grand Canyon tributaries is the Colorado River. An interesting example of this effect is the lower gorge of the Little Colorado River. Upstream from the gorge, the Little Colorado River meanders across northeastern Arizona in a broad, open valley cut in Mesozoic strata. In its last 30 miles (48 km), however, the river gradient increases by about 500 percent and cuts a gorge through the entire 2900-foot (884-m)-thick Paleozoic section. Such an abrupt change in gradient probably was initiated by integration of the Little Colorado River into the main Colorado River system. Regional considerations indicate that this occurred about four million years ago, when Lake Bidahochi, centered in the present Little Colorado River Valley, was drained (Scarborough 1989). Through the process of headwater erosion, the ephemeral Little Colorado will be adjusting its gradient to this change for thousands of years to come.

MORPHOLOGICAL DIVISIONS OF THE GRAND CANYON

The Grand Canyon can be subdivided into four segments based on the general morphology of side canyons: (1) Marble Canyon, (2) Chuar, (3) Kaibab Plateau, and (4) Esplanade. The differences between these four segments can be observed readily on the Geologic Map of Arizona (Reynolds 1988) and in Fig. 21.4.

Marble Canyon Segment

This segment includes all canyons cut into the Marble Platform from below Lees Ferry to Little Nankoweap Creek (mile 52). It is characterized by fairly short, narrow, relatively unbranched gorges, such as North and Nautiloid canyons. These streams flow in accord with the regional northeast dip, while the river flows against it, resulting in distinctively barbed tributaries (Fig. 21.4). Despite fairly large watersheds, the canyons do not incise a large area of the Marble Platform and have the appearance of being newly formed.

Chuar Segment

From Little Nankoweap Creek to the beginning of the Upper Granite Gorge, canyon development is influenced strongly by the East Kaibab monocline and the Butte fault. The most distinctive feature in this segment is the abrupt appearance of enormous tributaries from the west and the virtual absence of lateral gorges from the east (Fig. 21.4). The river flows south, paralleling the base of the East Kaibab monocline, then swings broadly westward as it cuts across the monocline, deep into the core of the Kaibab Plateau. This segment is the structurally and topographically deepest part of the Grand Canyon. The large tributaries from the Kaibab uplift owe their size to the greater runoff and more significant relief of the lofty upland region. Moreover, broad, open valleys in their upper reaches are a consequence of incision into the soft Chuar Group shales and the Dox Formation. On the east rim, the absence of canyons along the Desert Facade is the result of low rainfall, resistant rock, and the regional dip of Paleozoic strata away from the river. The Little Colorado River is the only tributary from the east along this entire 48-mile (80-km)-long segment, yet it is the largest of the tributaries entering the Grand Canyon. Not far below its confluence with the Little Colorado River, the main Colorado River crosses a splay of the East Kaibab monocline, where the easily eroded Upper Proterozoic shales are exposed on both sides of the river. Here, small canyons begin to appear on the east side because of the weak shales, despite the continued dip of strata away from the river.

Kaibab Segment

This segment includes all canyons controlled by the Kaibab upwarp between Red Canyon (mile 77) and Kanab Creek (mile 143) (Fig. 21.4). Through the uplifted Kaibab Plateau, the river cuts to its deepest stratigraphic level, nearly 1800 feet (547 m) into the Proterozoic crystalline basement. Through most of this distance, the main river channel is incised in a narrow, steep-walled canyon called the Upper Granite Gorge. Snowmelt from the Kaibab Plateau produces several large, perennial streams that flow radially toward the canyon off the southern and western flanks of the plateau. The north rim tributaries commonly are long and fault-controlled and display a well-developed dendritic pattern. In contrast, southern tributaries flow against the dip of the strata and form short, steep, and generally unbranched canyons (Fig. 21.4). Outwash from these small, south-side tributaries produces many of the Grand Canyon's most tyrannical rapids (e.g., Hance, Sockdolager, Horn Creek, and Granite Falls).

Esplanade Segment

This segment is named for the characteristically broad topographic bench formed on the Esplanade Formation by the erosional retreat of the Kaibab–Coconino cliffs. It includes the entire western Grand Canyon downstream from Kanab Creek (mile 143) (Fig. 21.4). The Esplanade surface actually becomes a prominent feature rather abruptly at the lower end of the Upper Granite Gorge (mile 114) but does not become fully developed until the river leaves the Kaibab upwarp near Kanab Creek. Many side canyons in this segment are uncommonly long and linear, reflecting a strong control by north-trending faults that break the western Colorado Plateau into a series of fault blocks. Such faults control the position and trend of Prospect Creek, Peach Springs Canyon, and other side canyons, in addition to the overall course of the main Colorado River along the Hurricane fault (Fig. 21.3). The great width of these canyons is confined to the portion above the level of the Esplanade. Much narrower gorges are cut in the underlying Pennsylvanian–Cambrian rocks. The tenfold increase in width of the Esplanade west of the Kaibab Plateau suggests that canyon-cutting into the Kaibab Limestone began much earlier here than in areas to the east.

The Esplanade segment also includes two major tributaries—Kanab Creek and Havasu Creek—and side canyons with long and complex geological histories. For example, the present mouth of Prospect Canyon is a mere remnant of a wider, deeper canyon that was filled with basalt flows after its incision (Hamblin 1976). Other canyons, such as Milkweed and Peach Springs canyons, originally were formed by north-northeast-flowing drainages that predated cutting of the present Grand Canyon; the walls of these present-day side canyons contain remnants of sedimentary and volcanic rocks deposited during this earlier drainage regime.

GEOLOGICAL HIGHLIGHTS OF SELECTED SIDE CANYONS

A variety of factors have influenced the size, orientation, and morphology of side canyons. Some of these factors are local features, such as faults, whereas others are regional in extent, resulting in the four canyon segments discussed above. To illustrate the attributes of side canyons and the processes that control them, the geological highlights of nine side canyons (Fig. 21.3) are presented below. These nine canyons have been chosen for discussion because they display a wide range of geological features in a variety of settings. We have included brief descriptions of the aesthetics of most canyons in an attempt to convey the natural wonder of the Grand Canyon as viewed from river level and from the vantage point of short hikes up the side canyons.

North Canyon

In the first 20 miles (32 km) below Glen Canyon Dam, the Colorado River cuts through the red-colored Mesozoic rocks of the Vermilion and Echo cliffs into the underlying, light-colored Permian rocks that form the familiar rim of the Grand Canyon. By North Canyon (mile 20 from Lees Ferry), the river also has deeply incised the underlying, rust-colored Permian Hermit Shale and the Pennsylvanian–Permian Supai Group. North Canyon, typical of the Marble Canyon segment, is a narrow cleft cut in sculptured formations of the Supai Group, including a gray conglomerate that marks the base of the Esplanade Sandstone. The canyon walls are steep and high, with the warm colors, soft textures, and acoustics of a cathedral. Parts of the canyon have the feeling of a sanctuary because of smooth, concave walls formed by dramatic, curved fractures in the sandstone (Fig. 21.5). The unusual fractures presumably reflect exfoliation of the sandstone walls that was caused by canyon incision.

Nautiloid Canyon

Not far downstream from North Canyon, gray- and cream-colored strata appear at river level and form a small cliff along the river's shores. Through the next 12 miles (19 km), the river's course becomes confined within the towering walls of the canyon's single most formidable barrier to river-rim hiking namely, the Redwall Limestone. This fossiliferous, carbonate rock was deposited in a vast, shallow sea that inundated much of western North America during Mississippian time, more than 300 million years ago. In 1869, John Wesley Powell named Marble Canyon for the beautiful, marblelike polish of the Redwall Limestone flanking the river. The limestone's true ivory and gray colors, exposed by running water or recent rockfalls, usually are concealed by a red iron-oxide stain derived from the overlying Supai Group and Hermit Shale.

The 500-foot (152-m) vertical walls of nearly pure limestone and dolomite are highly resistant to erosion in the semiarid climate of the inner canyon. Numerous solution caverns, however, were formed by groundwater dissolution of the limestone during the geological period between deposition of the Redwall and the overlying Supai Group (McKee 1976). Many large caverns developed along vertical joints or cracks in the limestone and are conspicuous—particularly near mile 35, where a large set of joints bisects the canyon.

In this vicinity, a narrow cleft called Nautiloid Canyon enters from the east (Fig. 21.3). A short climb into this canyon reveals numerous fossilized remains of the chambered nautiloid (Fig. 21.6). These fossils were the first of their kind to be found in the Grand Canyon (Breed 1968). The polished limestone floor of Nautiloid Canyon bears many cross sections of these ancient, cone-shaped creatures, whose size can be up to 20 inches (55 cm) in length. The tentacle-like appendages on some specimens may be relicts of soft parts. These fossils, found in the cherty Thunder Springs Member of the Redwall Limestone, are but one of several types of preserved marine life.

Buck Farm Canyon

Small rapids and riffles occasionally punctuate the shady serenity of Marble Canyon as the river cuts into older Paleozoic strata downstream from Nautiloid Canyon. Only the keenest observer will notice where the river intersects the inconspicuous horizontal contact between the Mississippian Redwall Limestone and the underlying Cambrian Muav Limestone. The parallel layering and similar appearance of these two red-stained limestones belies the staggering difference in their ages. Rocks representing Late Cambrian through Early Mississippian time (nearly 180 million years of earth history) are missing at this stratigraphic boundary. A clue to this unrecorded period is evident in the canyon walls upstream from Buck Farm Canyon (mile 41), where purplish lenses of Devonian Temple Butte Limestone 50 feet (15 m) high and hundreds of feet long are present between the Redwall and Muav limestones (Fig. 21.7). Especially well-exposed examples of these lenses are present in Buck Farm Canyon.

Figure 21.5. North Canyon. Curved fractures and bedding surfaces occur in the Permian Esplanade Sandstone of the Supai Group. (Photograph by A. Potochnik.)

The bowl-shaped cross section of these lenses suggests the following evolutionary scenario. After deposition of the Cambrian Muav Limestone, the continent emerged above sea level, and streams carved channels into the landscape. Subsequent transgression of a Devonian sea filled these channels with impure limestone, which accumulated to even greater thicknesses in the western Grand Canyon (Beus 1987). Both Temple Butte Limestone and Muav Limestone then

FIGURE 21.6. Fossil nautiloid on stream-polished outcrop of Mississippian Redwall Limestone. (Photograph by S. Reynolds.)

were eroded down to a peneplain as the landmass once again emerged from the sea in Late Devonian time. This second erosional period so thoroughly leveled the landscape that the Temple Butte Limestone in the eastern Grand Canyon was preserved only at the bottom of channels in which it first had accumulated. A third marine advance from the west in middle Mississippian time blanketed the region with the Redwall Limestone. This activity preserved the channel infillings in the geological record. The relatively recent cutting of the Grand Canyon affords cross-sectional views of these channels.

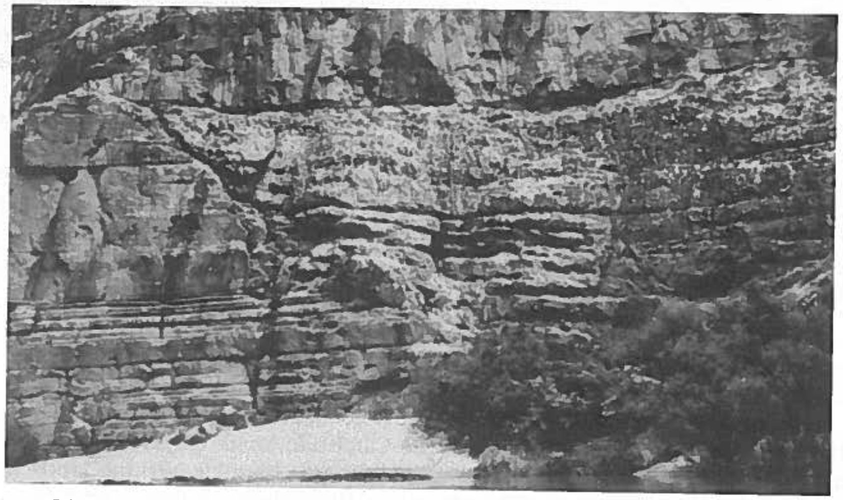


FIGURE 21.7. Lens of Devonian Temple Butte Limestone in a channel cut into the underlying Cambrian Muav Limestone. The lens is overlain by Mississippian Redwall Limestone. (Photograph by S. Reynolds.)

How did the landscape appear during the erosional intervals, and what lived in the Devonian sea? The flat-lying, undisturbed nature of the Paleozoic rocks tells us that dynamic crustal activity, such as mountain building, faulting, and volcanism, was absent. During the first erosional episode, the area was a bleak, featureless landscape of Muav Limestone that was incised by local stream channels. Probably no plants or animals lived on land, but early, bony-plated fishes first made their appearance during the transgression of the Devonian sea. The landmass subsequently reemerged, and the earliest land plants took root. The subsequent inundation by the Mississippian sea brought a much greater abundance and diversity of marine life—including corals, sponges, shellfish, echinoderms, and nautiloids.

Carbon Canyon

The Marble Canyon segment of Grand Canyon ends about 11 miles (18 km) downstream of Buck Farm Canyon, where the river enters the Chuar segment. The configurations of the side canyons change dramatically in response to the uplift of the broad Kaibab Plateau. Carbon Canyon, a lateral gorge entering the river from the west at mile 65, provides a fascinating structural and geomorphical perspective of the eastern boundary of the uplifted Kaibab Plateau.

The canyon walls in the lower portion of Carbon Canyon consist of purple sandstone and siltstone of the Proterozoic Dox Formation and buff-colored, coarse-grained arkose of the overlying Cambrian Tapeats Sandstone. The latter exhibits striking examples of honeycomb-weathering and colorful Liesegang banding. The purple- and rust-colored, parallel banding forms graceful, curving patterns throughout the rock (Fig. 21.8). It was caused by precipitation of iron oxides as groundwater migrated through the saturated matrix of the porous sandstone.

Evidence of the forces that uplifted the Kaibab Plateau relative to areas to the east is displayed dramatically a short distance up Carbon Canyon. Here, bedding planes in the Tapeats Sandstone become gently tilted upward as one walks up the canyon until a place is reached where the beds abruptly bend vertically. At this point, the narrow gorge opens into a wide valley of rolling hills with many small tributaries that drain the soft shales of the Proterozoic Chuar Group. The Kaibab Plateau, underlain by the entire Paleozoic sequence, is visible on the western skyline 5 miles (8 km) away and 2000 feet (610 m) higher than the elevation of the same formations along the river. The sharp upturn in the strata in Carbon Canyon is a local fold caused by the Butte fault, which parallels the East Kaibab monocline, the eastern boundary of the Kaibab Plateau. The Butte fault was a normal fault (west side down) in the Proterozoic but was reactivated as a reverse fault (west side up) during the Late Cretaceous to early Tertiary uplift of the Kaibab Plateau. Cenozoic erosion has breached the monocline, causing removal of the entire Paleozoic sequence and exposure of the underlying Chuar Group shales.

Monument Creek

Near mile 93, within the Kaibab segment, Monument Creek enters the Colorado River from the south side in the deepest section of the Upper Granite Gorge. Within the walls of this side canyon are exposures of Early Proterozoic metamorphic and granitic rocks, whose resistance to erosion is responsible for the steep-walled, "V" shape of the inner gorge (Fig. 21.9). The steep metamorphic layering and numerous convoluted folds within the rocks were formed during

Figure 21.8. Liesegang banding in Cambrian Tapeats Sandstone, Carbon Canyon. Bedding dips to the left. (Photograph by S. Reynolds.)

metamorphism and deformation that occurred during amalgamation and suturing of terranes along a continental margin to the southeast (Anderson 1986; Ilg et al. 1996). The mountains that formed during this tectonic episode were eroded away before the Cambrian sea encroached on the landscape and buried it beneath the beach sands of the Tapeats Sandstone. The "Great Unconformity" between the steeply dipping metamorphic rocks and the overlying, gently inclined Tapeats Sandstone is well-exposed and represents over one billion years of time.

In the more recent geological past, boulders carried down the present canyon of Monument Creek and deposited at its mouth have partially dammed the Colorado River, creating the major rapid of Granite Falls (Webb et al. 1989). Monument Canyon, which contains remnants of debris deposited in 1984, has the characteristics that a side canyon needs to form a large rapid in the river. These include a steep stream gradient, a narrow canyon with a flat floor, a sufficient supply of large boulders, and a source of fine-grained material to generate mudflows or debris flows (large boulders are more easily transported in a medium that is thicker than water). Monument Creek does not have a large drainage area, compared to other canyons; apparently, this factor is of secondary importance in creating a large rapid (Griffiths et al. 1996).

Blacktail Canyon

The Upper Granite Gorge ends in the vicinity of Blacktail Canyon (mile 120), where the regional westward dip of the strata causes the Tapeats Sandstone to descend to river level. Blacktail Canyon is a narrow, somewhat tubelike notch cut along the Great Unconformity between the sandstone and underlying schist of the Proterozoic Vishnu Group. The details of the unconformity are incredibly well exposed along the polished walls of the canyon (Fig. 21.10). The vertical metamorphic layering in the schist is overlain by sandstone and conglomerate derived from weathering and erosion of the schist. Although thin, vertical quartz veins in the schist were somewhat resistant to weathering, they finally were eroded into the small quartz pebbles now found in the basal sandstone. When

Figure 21.9. Upper Granite Gorge. Dark-colored walls are Proterozoic metamorphic and igneous rocks. (Photograph by S. Reynolds.)

standing here, it is easy to imagine the waves of the Cambrian sea 550 million years ago crashing onto jagged hills of schist and churning the metamorphic rock into sandy beaches as the sea advanced across the barren landscape.

Tapeats Creek

A few miles downstream, near mile 134, a cold and clear-flowing perennial stream called Tapeats Creek joins the Colorado River. The lowest formations of the Middle Proterozoic Unkar Group are exposed near river level (Fig. 21.11). These formations have a characteristic tilt that readily distinguishes these rocks from the more flat-lying Paleozoic rocks. A creek-side path through ancient ruins and garden sites of the Anasazi Indians traverses upward through the Unkar Group into the overlying Paleozoic rocks. Here, Thunder Springs bursts from a cavern high in the Muav Limestone wall. The Bass Formation, the oldest unit in the Unkar Group, contains wavy, fossilized algal mats, the oldest preserved evidence of life revealed in the Grand Canyon. Bright, reddish orange shales and siltstones of the overlying Hakatai Shale contain ripplemarks and mudcracks, features that suggest deposition in a tidal-flat environment. A gradational contact between these shales and the underlying Bass Limestone indicates that Hakatai mudflats gradually displaced the algal marine environment as the Bass sea retreated from the area. The tidal flats, in turn, were covered by a thick sequence of sand deposited near the shoreline of a sea. Consolidation of these sands formed the overlying Shinumo Quartzite, a cliff-forming unit that constitutes the steep-walled, narrow canyon of upper Tapeats Creek. A sill of dark-colored diabase approximately one billion years old occurs within the Bass Formation. As the layers in

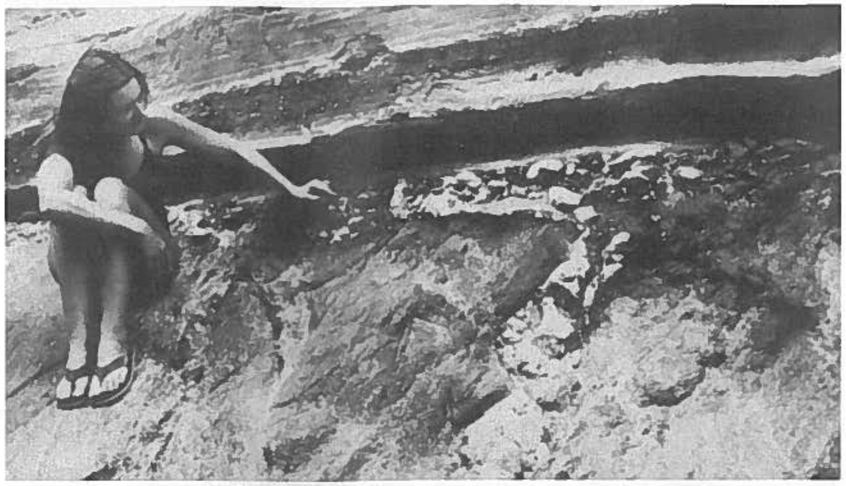


Figure 21.10. The Great Unconformity between Tapeats Sandstone and the underlying Proterozoic Vishnu Group, Blacktail Canyon. Material from the light-colored layer within the schist has been incorporated into the thin, basal conglomerate of the overlying sandstone. The unconformity represents more than one b.y. missing from the geologic record. (Photograph by A. Potochnik.)

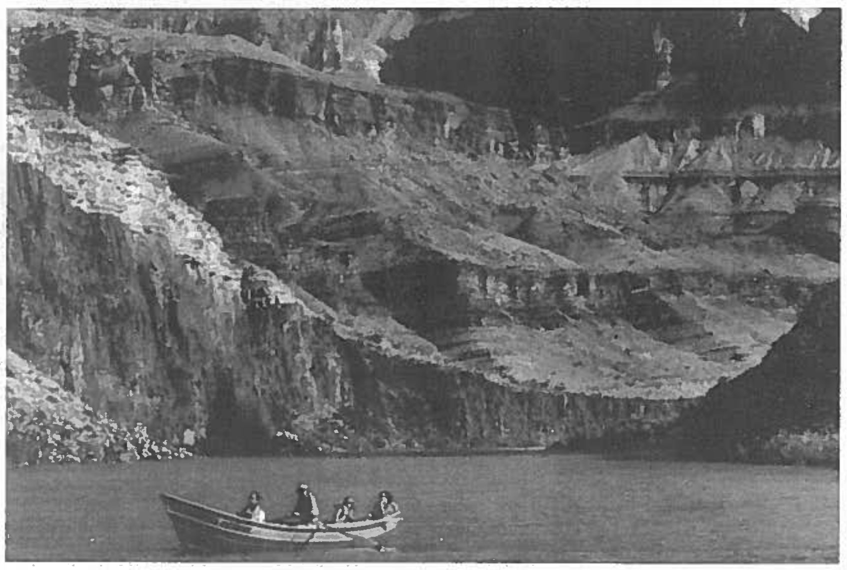


Figure 21.11. Proterozoic diabase (dark ledge at river level) and slope-forming units of the Proterozoic Unkar Group along the Colorado River near Tapeats Creek. Tilted units of Unkar Group are overlain unconformably by Cambrian Tapeats Sandstone. (Photograph by J. Blaustein.)

the Bass Formation were pushed apart to accommodate the magma, they reacted with this magma to form thin layers of green serpentine and fibrous, chrysotile asbestos.

An up-canyon view from high on the Thunder River switchbacks reveals the angular unconformity between the Shinumo Quartzite and the overlying Cambrian rocks (Fig. 21.12). The Tapeats Sandstone, a beach sand of the advancing Cambrian sea, was deposited on the shore of a Shinumo Quartzite island that stood as a large remnant of late Proterozoic erosion. As the sea deepened, the island became submerged, and offshore muds of the Bright Angel Shale lapped across the top of the former island.

Havasu Creek

Downstream from Tapeats Creek, the river turns west and quickly passes out of the Kaibab segment and into the Esplanade segment. The regional tilt of the rock layers causes the cliff-forming Paleozoic limestones, once again, to appear at river level. The confluence of Havasu Creek with the Colorado River near mile 157 is easily missed. A narrow Muav Limestone gorge obscures the enormity of this large tributary, second only to the Little Colorado River in size. Havasu Creek is known for its spectacular waterfalls. The verdant banks of this perennial, aqua-blue stream are lined with velvet ash, cottonwood, and wild grape.

Figure 21.12. Tilter Proterozoic Shinumo Quartzite (Unkar Group) overlain unconformably by flat-lying Cambrian strata, Tapeats Creek. The resistant quartzite was once a craggy island. Along its flanks, Cambrian seas deposited the sands of the Tapeats, which forms the dark ledge shown in the left center of the photograph. The island later was buried by marine muds, which were compacted into the overlying, slope-forming units of the Cambrian Bright Angel Shale. (Photograph by S. Reynolds.)

Travertine deposits are perhaps the most fascinating geological feature of Havasu Creek. The travertine is formed by the precipitation of calcium carbonate as the creek waters warm and evaporate during the long flow to the Colorado. The travertine tends to encrust the surface and to take the form of any object over which the water passes.

Distinctive features of travertine cementation are the flat-topped and sinuous dams so commonly seen in the creek. These dams form by a self-enhancing process. An obstruction tends to catch sticks and leaves, which become encrusted with calcium carbonate, thereby increasing the size of the obstruction. When the obstruction becomes large enough, mosses colonize it and provide an additional substrate that increases its width and size. Eventually, a dam forms across the channel with perhaps one or two spillways through which the stream flows. The process becomes self-restricting in the spillways because water velocity is sufficient to prevent accumulation of debris, growth of moss, and precipitation of travertine.

Whitmore Wash

Near mile 188, below the notorious Lava Falls rapids, Whitmore Wash preserves evidence of a time even more tumultuous than that experienced by the river traveler while navigating the rapids. Whitmore Wash and the area around Lava Falls contain remnants of dark, basalt lava flows that once filled the Grand Canyon to a depth of more than 1400 feet (427 m). The lava erupted from volcanic vents, such as Vulcan's Throne, that pierced the Esplanade surface approximately 3000 feet (910 m) above the canyon bottom (Hamblin 1976 and Chapter 17, this volume).

A number of volcanic vents occur near the Hurricane fault, a major, recently active, north–south fault that may have served as a conduit for the ascending lavas. Upon eruption, the flows of molten lava cascaded over the walls of the Grand Canyon into the Colorado River 3000 feet (910 m) below, creating enormous clouds of steam and filling tributary canyons that drained into the Grand Canyon prior to volcanism. A cross-sectional view of the lava-filled canyon of "old" Whitmore Wash is visible from river level where the Whitmore Trail climbs the north wall of the Grand Canyon into Whitmore Valley. Less than one-half mile downstream from the trail is the "new" Whitmore Wash, a narrow side canyon that drains the same extensive watershed as the former wash. The new wash, however, has cut into the Paleozoic limestones instead of excavating the more erosionally resistant lava that fills the old channel. In the main Grand Canyon, the dams formed by lava flows were more transient, probably surviving less than 10,000 years (McKee et al. 1968; Damon et al. 1967; Hamblin 1976).

SUMMARY

The canyons described above represent just a small sample of the geological features and natural beauty found within side canyons of the Colorado River. Each canyon is unique, both in scenary and in the array of exposed geological features. Short hikes within these canyons complement the river running, which alternates between the relaxing tranquility of long, slow-moving stretches and the burst of apprehension, excitement, and chaos within the rapids. The entire experience is difficult to describe, but impossible to forget.

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