NORTHERN ARIZONA
INTRODUCTION

The Grand Canyon provides an unusual opportunity to view a variety of geological features in three dimensions and with near-perfect exposure. The canyon ranges in depth from several hundred meters to more than 1500 meters. When viewed over the 140 or so km of the float trip, the canyon is a remarkable full-scale geologic cross section that depicts with great clarity not only stratigraphic, structural, and geomorphic features, but also the processes by which rivers are born and canyons eroded.

The drive to the embarkation point at Lees Ferry illustrates many features of the open, low-relief terrane typical of much of the Colorado Plateau. These features, described below, are different from those found in the Canyon country, a contrast important for understanding the birth and evolution of the Colorado River and the Grand Canyon. Because of this, a condensed road log is provided for the drive from Flagstaff to Lees Ferry.

This guidebook is intended to introduce and complement information contained in the river guide (Hamblin and Rigby, 1968) and separates that are part of the field-trip package. This format is chosen to avoid duplicating the river guide already available. The information provided in the guidebook consists chiefly of the description of sites that are especially important in understanding the geology of Marble Canyon and the upper Grand Canyon. Also included are analyses of problems of general interest such as the history of the Colorado River and its canyon. Mile-by-mile description of geologic features along the river is left to the river guide.

Geologic topics of particular interest in the field trip to Marble Canyon and the upper Grand Canyon include the following:

1. The Paleozoic shelf sequence of the Grand Canyon.
2. The Lower Pennsylvanian (?) and uppermost Mississippian Surprise Canyon Formation, a recently defined rock-stratigraphic unit (Billingsley and Beus, 1985).
3. The Middle Proterozoic sedimentary rocks of the Grand Canyon Supergroup.
4. Characteristics and depositional environment of the Middle Proterozoic Cardenas Lava.
5. Characteristics and contact metamorphism of Middle Proterozoic diabase sills.
6. Igneous and metamorphic rocks of the Proterozoic basement.

ROADLOG FROM FLAGSTAFF TO LEES FERRY

This roadlog focuses primarily on aspects of stratigraphy, structure, and geomorphology that are especially relevant to the discussion of the history of the Colorado River and the Grand Canyon. For a more complete treatment of the geology along the route, consult Holm and Ulrich (this volume) and Karlstrom and others (1974). A stratigraphic column of rocks visible on the trip is presented elsewhere in this field guide.

The roadlog begins at the junction between highway US 89 and the Leupp—Winona road. This junction is just north of Camp Townsend, one of the last outskirts of Flagstaff north on US 89.

A map (Figure 1) is provided to help locate principal features.

Odometer 0, MP 420—421.
Leupp—Winona road turnoff. Set odometer to zero, and continue straight on US 89. The turnoff is just beyond Camp Townsend, one of the last outskirt of Flagstaff north on US 89.

In the roadlog, 'MP' refers to the green mileposts located along the highway at one-mile intervals. Milepost readings are given to help correct for discrepancies in the calibration of individual odometers.

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FIGURE 1. Index map for northern Arizona. Arrows indicate beginning (B) and end (E) of river-trip route.
forms the eastern boundary of the Coconino Plateau and is bounded on the east by a monocline that passes southward into the Mesa Butte fault. The Coconino Plateau is the southern continuation of the Kaibab Plateau, from which it is separated by the Grand Canyon. These two plateaus form a north-trending belt of high ground that lies across the path of the Colorado River, giving rise both to difficulties in developing theories on the origin of the course of the river, and several concepts of how rivers cross topographic and structural barriers.

On clear days, Navajo Mountain is visible far to the north, where it straddles the Arizona-Utah boundary. This isolated mountain is a dome cored by a Tertiary laccolith and one of the four sacred mountains of the Navajo Indians.

20.3 MP 441

View at 2 o'clock of the valley of the Little Colorado River. This northwest-trending strike valley occupies the cuesta trough at the featheredge of the soft Triassic Moenkopi Formation on the underlying and very resistant Permian Kaibab Formation. The valley is at the foot of the northwest-trending cliffs that form its northeast flank. These cliffs are held up by the Triassic Shinarump Member of the Chinle Formation, which is more resistant to erosion than the underlying Moenkopi. The cliffs are retreating northeastward down the structural slope, leaving behind a stripped surface on top of the Kaibab. This surface slopes gently northeast and forms the subdued southwest side of the valley. Cliffs and mature strike valleys such as the ones visible here are typical of the Plateau country not yet affected by canyon cutting. These features are representative of the old pre-canyon-cutting topography.

24.1 MP 444-445

Turnoff on right to Wupatki National Monument.

29.4 MP 450

View of Gray Mountain at 10:30, with the Coconino monocline on its east flank. The monocline becomes more subdued southeastward and passes into the Mesa Butte fault. The continuous lava cap ends, and the road now traverses red beds of the Triassic Moenkopi Formation. Prominent flow on skyline to east of road is the Black Point flow (2.39 Ma), which reaches the Little Colorado River to the north.

36.4 MP 457

Gray Mountain, a settlement just outside the southern boundary of the Navajo Indian Reservation. At 11 o'clock, in the distance, is the dark mass of Shadow Mountain (620 Ka, Condit, 1974), a basaltic volcano. Beyond, the Echo Cliffs are visible on the skyline. The road follows the western base of these cliffs.

36.8 MP 457-458

Top of Black Point monocline, which splays eastward from Gray Mountain monocline. Road is built on surface at top of the Kaibab Formation, whose great resistance to erosion causes topography to follow structure over much of the southwest Colorado Plateau. Red beds ahead are the Moenkopi, which is retreating down the structural slope. Cream-colored bed on top of the Moenkopi is the Shinarump, at the base of the Triassic Chinle Formation. What one sees here is analogous to the situation at the valley of the Little Colorado River but exaggerated in comparison because of the steeper structural slope on the monocline.

Erosional remnants of the Shinarump are visible along the highway for several miles.

41.5 MP 462

Excellent view of scarps forming the northeast side of the Little Colorado River valley. These scarps are typical of those retreating down structural slopes throughout the Colorado Plateau wherever there is a hard-over-soft lithologic couplet. These scarps have had a major influence on drainage patterns.

44.6 MP 465+

Turn off to south rim of the Grand Canyon (State Highway 64). Continue straight ahead on US 89.

46.4 MP 467

Cameron trading post and bridge over the Little Colorado River. The bridge is near the eastern limit of canyon incision along this drainage. From here west (downstream) the river is in a canyon that quickly becomes deeper and narrower downstream. To the east there is instead the wide and mature valley of the Little Colorado that was visible earlier. The bridge is at the interface between the "Canyon country" and the older and much narrower "Plateau country." The Little Colorado is one of the rivers that bring muddy water to the Colorado River. If the Little Colorado is not flowing here, chances are improved for having clear water during the float trip.

49.4 MP 470

Beginning of Painted Desert, characterized by peculiar and picturesque greenish, yellowish, and purplish hillocks eroded in the Chinle Formation (Figure 2), which contains much petrified wood. To the east, the prominent scarp is developed in the soft Chinle capped by the more resistant sandstone of the Moenave Formation. This is another of the scarps typical of the Plateau country and developed on hard-over-soft lithologic couplets. Shadow Mountain to west.

54.7 MP 475-476

View straight ahead of the Echo Cliffs, on the downthrown side of the Echo Cliffs monocline, which begins near this point. These cliffs represent inversion of topography (structurally lower side is higher topographically), here produced by a hard-over-soft couplet (Moenave and Wingate Sandstones over Chinle shale) migrating down a structural slope and parking in a structurally low area. The couplet thus forms a cliff and high terrain in the structurally low area, whereas the structurally high area to the west has been stripped down to the Kaibab Formation. A similar situation probably existed formerly along the Kaibab-Coconino plateaus, whose axial areas thus would have been lower, or at least no higher, than surrounding terrain still underlain by Mesozoic rocks (Lucchitta, 1972, 1984). This would help explain how the Colorado River crossed the belt of high country across its path.
Good view of the Echo Cliffs and the strike valley developed at their foot in greenish and purplish Chinle shale. The Shinarump conglomerate crops out west (left) of the road. Beds east of road and forming the east side of the valley are the Moenave (small cliff), Kayenta (slope), and Wingate and Navajo (large cliff), all Mesozoic.

Greenish basaltic dike cutting Chinle shale at foot of cliff east of road.

The Gap, Navajo settlement named after the conspicuous gap in the Echo Cliffs. East of this gap are prominent river gravels preserved in linear outcrops (channels) that trend toward the gap. The gravels may have been deposited by drainages that were part of the ancestral Colorado River system or, alternatively, by drainages emptying into the Pliocene and late Miocene Bighocchi Lake near Winslow and Holbrook (Hereford, oral communication, 1986).

Transition from strike valley cut in the Chinle to one cut in the Moenkopi. The slope west of road is now in the Kaibab, whereas the red beds to the east are Moenkopi. The cream-colored Shinarump is visible above the Moenkopi.

Cedar Ridge settlement. In the past several decades, many Navajos have clustered in such settlements instead of being dispersed as before. Drainage divide. The road now follows washes that flow north and northwest into Marble Canyon.

Back into Chinle strike valley. Shinarump and Moenkopi are now on left.

Prominent outcrops of irregular-weathering, light-colored aeolian Navajo Sandstone in upper part of cliff on right. The strike valley is again in the Moenkopi here. Imposing cliffs in the distance to the north are the Vermilion Cliffs, on the northwest side of the Colorado River.

Toreva block (rotational landslide) capped by light-colored Navajo is exposed on face of Echo Cliffs just north of mouth of small canyon at 2 o'clock. Such blocks are common along the cliff face. A conspicuous block occurs about 3 miles north of here, where the cliffs change trend.

Turnoff to Page and Lake Powell (US 89). Continue straight ahead on US 89A toward Fredonia and Grand Canyon. Incoherent landslides are prominent on cliff face several miles north of turnoff. Near this stretch of road, washes flow in shallow gullies but become deeply incised within a few hundred yards to the west (downstream) and join the Colorado River in Marble Canyon within a few miles. The road here is at the interface between Canyon and Plateau terrain.

Excellent view of Vermilion Cliffs on left. Most of the cliff face visible from here is in the Chinle and younger units. The dark line near the base of the cliffs is the Shinarump. Many landslides and debris-flow deposits are visible on the face of the cliffs. Prominent maroon outcrop on the right of the highway is Moenkopi capped by Shinarump.

Coarse debris-flow deposits of Holocene (?) age on right. Similar deposits are present in several places to the north of here along the highway.

Reentrant in cliffs straight ahead is formed by the junction of the Echo and the Vermilion Cliffs. A subdued anticline is visible in this area. The reentrant probably is the result of Coarse debris-flow deposits of Holocene (?) age Chinle and younger units. The dark line near the base of the cliffs is the Shinarump. Many landslides and debris-flow deposits are visible on the face of the cliffs. Prominent maroon outcrop on the right of the highway is Moenkopi capped by Shinarump.

Navajo Bridge, built over the Colorado River in 1929 to replace the old Lees Ferry. The next place downstream where the river can be crossed is at the Phantom Ranch footbridge in the Grand Canyon, 85 mi (135 km) downstream from here. The nearest place downriver where it is possible to drive across is at Hoover Dam, about 310 mi (500 km) away.

Marble Canyon trading post and junction with Lees Ferry road. Turn right. The Lees Ferry road winds for several miles through picturesque outcrops of maroon Moenkopi capped by Shinarump, which forms the light-colored boulders littering the Moenkopi slopes. Landslides are visible in places to left of highway.

View on right of the Paria riffle on the Colorado River.

Paria River crossing. This is the second river that can dump muddy water into the otherwise blue-green Colorado. If the Paria is not flowing, the Colorado is likely to be clear at least to its junction with the Little Colorado.

Boat ramp and put-in at Lees Ferry. The Shinarump Member is at river level here and crosses the river at the boat ramp. The Grand Canyon, called Marble Canyon in this area, begins a short distance downstream where the river starts incising the Kaibab Limestone.

END OF LOG

### GUIDE TO SELECTED ROCK UNITS AND STOPS 1-13 FOR FLOAT TRIP THROUGH MARBLE CANYON AND EASTERN GRAND CANYON

<table>
<thead>
<tr>
<th>AGE</th>
<th>ROCK UNIT</th>
<th>LITHOLOGY</th>
<th>ENVIRONMENTAL INTERPRETATION</th>
</tr>
</thead>
<tbody>
<tr>
<td>PETRIAN</td>
<td>KAIAB FORMATION</td>
<td>320 ft (100 m)</td>
<td>shallow marine shelf near shore</td>
</tr>
<tr>
<td></td>
<td>TOWEREAP FORMATION</td>
<td>300 ft (90 m)</td>
<td>near shore to off shore</td>
</tr>
<tr>
<td></td>
<td>COWHIDE SANDSTONE</td>
<td>340 ft (104 m)</td>
<td>near shore to off shore</td>
</tr>
<tr>
<td></td>
<td>HERMIT SHALE</td>
<td>500 ft (150 m)</td>
<td>onshore dunes</td>
</tr>
<tr>
<td></td>
<td>ESPLANADE SANDSTONE</td>
<td>100 ft (30 m)</td>
<td>coastal lowland swamp</td>
</tr>
<tr>
<td></td>
<td>WINGATE FORMATION</td>
<td>100 ft (30 m)</td>
<td>shallow marine to intertidal</td>
</tr>
<tr>
<td></td>
<td>SHINURO FORMATION</td>
<td>400 ft (130 m)</td>
<td>shelf and/or dunes</td>
</tr>
<tr>
<td></td>
<td>MOONEY FALLS FORMATION</td>
<td>500 ft (150 m)</td>
<td>coastal lowland streams</td>
</tr>
<tr>
<td>CAMPBRIAN</td>
<td>TAFITE SANDSTONE</td>
<td>0-200 ft (60 m)</td>
<td>shallow marine shelf near shore</td>
</tr>
<tr>
<td></td>
<td>NAVAJA SHALE</td>
<td>500 ft (150 m)</td>
<td>coastal lowland</td>
</tr>
<tr>
<td></td>
<td>SHINURO QUARTZITE</td>
<td>420 ft (130 m)</td>
<td>coastal streams and delta</td>
</tr>
<tr>
<td></td>
<td>BASS LIMESTONE</td>
<td>220-320 ft (70-100 m)</td>
<td>supratidal to subtidal delta</td>
</tr>
</tbody>
</table>

### EXPLANATION

- **LIMESTONE**
  - Ripples
  - Mud cracks
  - Cross-bedding
  - Tidal channels
  - Supratidal to subtidal

- **SANDSTONE**
  - Cross-bedding
  - Tidal channels
  - Supratidal to subtidal

- **SHALE**
  - Mud cracks
  - Cross-bedding
  - Tidal channels
  - Supratidal to subtidal

- **SILICITE**
  - Mud cracks
  - Cross-bedding
  - Tidal channels
  - Supratidal to subtidal

- **GRANITE**
  - Mud cracks
  - Cross-bedding
  - Tidal channels
  - Supratidal to subtidal

**FIGURE 3.** Paleozoic and Precambrian stratigraphic section, eastern Grand Canyon.
TABLE 1. DAILY SCHEDULE FOR FLOAT TRIP

<table>
<thead>
<tr>
<th>DAY</th>
<th>MAJOR STOPS</th>
<th>RIVER MILE</th>
<th>SITES AND GEOLOGIC FEATURES</th>
<th>MILES TRAVELLED PER DAY</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>1</td>
<td>11.2</td>
<td>Soap Creek: Hermit Shale and upper Supai Group.</td>
<td>11.2</td>
</tr>
<tr>
<td>2</td>
<td>2</td>
<td>20.5</td>
<td>North Canyon: Basal conglomerate and structures in Esplanade Formation, Supai Group.</td>
<td>23.6</td>
</tr>
<tr>
<td>3</td>
<td>2</td>
<td>24.5</td>
<td>Supai Group/Redwall Limestone contact; Surprise Canyon Formation.</td>
<td>17.8</td>
</tr>
<tr>
<td>4</td>
<td>3</td>
<td>33</td>
<td>Redwall Cavern: Thunder Springs Member, Redwall Limestone.</td>
<td>12.6</td>
</tr>
<tr>
<td>5</td>
<td>4</td>
<td>34.8</td>
<td>Nautiloid Canyon: Fossils in Redwall.</td>
<td>11.3</td>
</tr>
<tr>
<td>6</td>
<td>5</td>
<td>38.4</td>
<td>Channel-fill of Temple Butte Formation.</td>
<td>5.0</td>
</tr>
<tr>
<td>7</td>
<td>6</td>
<td>52.6</td>
<td>Nankoweap delta: Bright Angel Shale; rock fall.</td>
<td>6.0</td>
</tr>
<tr>
<td>8</td>
<td>7</td>
<td>61.5</td>
<td>Little Colorado River junction: Tapeats Sandstone.</td>
<td>12.6</td>
</tr>
<tr>
<td>9</td>
<td>8</td>
<td>64.8</td>
<td>Carbon Creek to Lava Canyon hike: &quot;Great unconformity&quot;; Dox Formation; Chuar Group; Butte fault.</td>
<td>5.0</td>
</tr>
<tr>
<td>10</td>
<td>9</td>
<td>69.1</td>
<td>Basalt Canyon: Cardenas Lavas; Nankoweap Formation</td>
<td>6.0</td>
</tr>
<tr>
<td>11</td>
<td>10</td>
<td>76.5</td>
<td>Hance Rapides: Dike and sill; asbestos mine; another &quot;great unconformity&quot;; Bass Limestone; Hakatai Shale.</td>
<td>6.0</td>
</tr>
<tr>
<td>12</td>
<td>11</td>
<td>81.5</td>
<td>Grapevine: Vishnu Group; erosion of river beaches.</td>
<td>5.0</td>
</tr>
<tr>
<td>13</td>
<td>12</td>
<td>87.5</td>
<td>Kaibab bridge: Bright Angel Creek; hike out South Kaibab trail to South Rim (6.5 miles).</td>
<td>6.0</td>
</tr>
</tbody>
</table>

*The following description of selected rock units and features is supplemental to that of the river guide by Hamblin and Rigby (1968).

Table 2. Formations in the Supai Group, Marble Canyon

<table>
<thead>
<tr>
<th>Formation</th>
<th>Thickness (m)</th>
<th>Age</th>
</tr>
</thead>
<tbody>
<tr>
<td>Esplanade Formation</td>
<td>100</td>
<td>Wolfcampian</td>
</tr>
<tr>
<td>Wescogame Formation</td>
<td>30</td>
<td>Virgilian</td>
</tr>
<tr>
<td>Manakacha Formation</td>
<td>59</td>
<td>Desmoinesian</td>
</tr>
<tr>
<td>Watahomigi</td>
<td>46</td>
<td>Morrowan and Early Atokan</td>
</tr>
<tr>
<td>Total Supai Group</td>
<td>235</td>
<td></td>
</tr>
</tbody>
</table>

Pennsylvanian-Permian Supai Group

Redbeds of the Supai Group form the canyon walls in Marble Canyon at river level from mile 11 to mile 24. McKee (1975; 1982) elevated the original Supai Formation to group status and recognized four formations (Table 2). Each of the formations is distinct in lithologic and topographic expression throughout most of the Grand Canyon. However, formation boundaries are locally obscure in the nearly vertical walls of Marble Canyon.
Limestone is everywhere sharp and well defined. The upper contact with the overlying Beus and Lucchitta (some places obscure. In the Marble Canyon area it is (Figure 4). The basal contact with the underlying Redwall purple-gray mudstone more typical of the lower placed where pale red-brown mudstone is overlain by red-brown sandstone with conglomerate lenses (Figure 3). Interbedded red-brown sandstone, siltstone, and composed mostly of red-brown mudstone a few meters Surprise Chert-pebble conglomerate overlain by about 2 m of pisolitic ironstone that grades upward into a pale dark red-brown strata filling shallow to deep channels (up to 122 m) cut into the top of the Redwall Limestone (Figure 3). In central and western Grand Canyon, channel-fill beds of limestone, sandstone, and conglomerate contain a variety of plant and animal fossils indicative first of fluvial- and then marine-dominated estuary conditions during deposition of the Surprise Canyon in latest Mississippian (Chesterian) time.

In eastern Grand Canyon and Marble Canyon, Surprise Canyon outcrops are smaller and less abundant and composed mostly of red-brown mudstone a few meters thick. An outcrop of the formation on the right bank just below 24 1/2-mile rapids consists of a basal chert-pebble conglomerate overlain by about 2 m of interbedded red-brown sandstone, siltstone, and pisolitic ironstone that grades upward into a pale red-brown sandstone with conglomerate lenses (Figure 4). The basal contact with the underlying Redwall Limestone is everywhere sharp and well defined. The upper contact with the overlying Supai Group is in some places obscure. In the Marble Canyon area it is placed where pale red-brown mudstone is overlain by purple-gray mudstone more typical of the lower Supai (Figure 4).

Each formation in the Supai Group is interpreted by McKee (1982, p. 23) to record a distinct marine transgression. All four units include both marine and nonmarine sediment and each is bounded by a hiatus marked by a basal limestone and sandstone-pebble conglomerate. We will examine an example of the basal conglomerate of the Wescogame Formation a short distance up North Canyon (mile 20.5, right bank). Marine fossils, including index fusulinids, are widespread in the Supai Group, particularly in the Watahomigi Formation where Atokan fusulinids occur as far east as the Eminence fault area of Marble Canyon, about mile 44 (McKee, 1982, p. 259). Abundant marine fossils, together with extensive limestone and the absence of large-scale cross-strata, attest to a depositional environment of low-energy marine or brackish water conditions for the Watahomigi.

The overlying Manakacha Formation has extensive cross-stratified sandstone and limestone beds and fewer marine fossils. The Manakacha was formed in higher energy, nearshore, marine to estuary conditions that became more nonmarine at the end (McKee, 1982, p. 210).

The Wescogame Formation contains intact marine fossils only in westernmost Grand Canyon. Highly cross-stratified sandstone and the occurrence of quadruped footprints in the eastern Grand Canyon suggest a fluvial environment for most of the Wescogame, particularly in eastern Grand Canyon and Marble Canyon. The Esplanade Sandstone is chiefly composed of large-scale cross-stratified sandstone everywhere except in the westernmost Grand Canyon, where the sandstone grades into a marine carbonate unit referred to as the Pakoon Limestone by McNair (1951). The marine fossil record of the Esplanade, like that of the Wescogame, is mainly limited to western Grand Canyon, although bioclastic debris of echinoderms, bivalves, and small foraminifers occurs in eastern Grand Canyon and Marble Canyon (McKee, 1982, p. 259). Deposition of the Esplanade occurred in a high-energy environment, probably under both marine and nonmarine conditions.

MISSISSIPPIAN SURPRISE CANYON FORMATION

The Surprise Canyon Formation (Billingsley and Beus, 1985) forms isolated lens-shaped outcrops of dark red-brown strata filling shallow to deep channels (up to 122 m) cut into the top of the Redwall Limestone (Figure 3). In central and western Grand Canyon, channel-fill beds of limestone, sandstone, and conglomerate contain a variety of plant and animal fossils indicative first of fluvial- and then marine-dominated estuary conditions during deposition of the Surprise Canyon in latest Mississippian (Chesterian) time.

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MISSISSIPPIAN REDWALL LIMESTONE

The Redwall Limestone typically forms a sheer 150-m red-stained cliff throughout the Grand Canyon (Figure 5). It is exposed at river level for 14 miles beginning at mile 22 and forms the imposing vertical walls that characterize Marble Canyon. Four members of the Redwall are recognized by McKee and Gutschick (1969) throughout northern Arizona.

The lowermost Whitmore Wash Member is a fine-grained, thin- to thick-bedded, oolitic, and fossiliferous dolomite in eastern Grand Canyon. It forms the lower fifth of the Redwall cliff and rests unconformably on the Muav or Temple Butte. The Thunder Springs Member is the most prominent unit in the Redwall. It is 28 m thick in Marble Canyon and characterized by a distinct banded appearance owing to the alternation of gray thin-bedded crinoidal limestone and thin beds or lenses of light- to dark-gray chert. Thin-section analysis indicates that the chert beds are silicified by oolites, wackestone and mudstone and the intervening carbonate beds are mainly crinoidal grainstone or packstone. Brenner (1986) suggested that the chert beds are preferentially silicified algal mats that served as a baffie to trap delicate bryozoan fragments. She considered the interbedded carbonate units to be crinoidal sands that periodically draped and buried the algal mats. The Whitmore Wash and Thunder Springs together are interpreted as the record of a major
FIGURE 5. Aerial view of south wall of Grand Canyon near the South Kaibab trail. B, Bass Limestone; Ba, Bright Angel Shale; R, Redwall Limestone; P, rimrock cliff of Kaibab Formation, Toroweap Formation, Coconino Sandstone.

marine transgression-regression cycle (McKee and Gutschick, 1969). The gradational boundary between the two members can be examined above the nautiloid horizon at Nautiloid Canyon (mile 34.8).

The Mooney Falls Member is the thickest unit in the Redwall and makes up 78 m, or more than half the Redwall cliff. The Mooney Falls is composed chiefly of thick-bedded limestone containing oolites, peloids, and a variety of skeletal fragments dominated by crinoid plates.

The Horseshoe Mesa Member is the highest and least extensive unit of the Redwall. It is about 35 m thick in Marble Canyon and consists of light-gray thin-bedded limestone and minor chert. This unit is considered the record of a slow marine regression following a major and prolonged transgression recorded by the underlying Mooney Falls Member.

The Redwall Limestone is fossiliferous throughout and is dated as Kinderhookian through Meramecian by index brachiopod, foraminiferid, and coral fossils (McKee and Gutschick, 1969).

MIDDLE PROTEROZOIC CHUAR GROUP

General Statement

The Chuar Group was named by Walcott (1883) and subdivided into formations and members by Ford and Breed (1973a). It is about 2,000 m thick (Figure 6) and exposed only in a belt measuring 24 by 6 km west of the Colorado River in eastern Grand Canyon. Chuar outcrops are cut off from the Colorado River by the Butte fault but are accessible from the river by foot traverses up east-draining tributary canyons including Nankoweap, Kaapang, Carbon, and Lava (Figure 7).

FIGURE 6. Stratigraphic section of Precambrian Grand Canyon Supergroup, eastern Grand Canyon.

Galeros Formation

The Galeros Formation forms the lower 1,300 m of the Chuar Group. The lowermost Tanner Member includes a basal dolomite 12-24 m thick and about 175 m of blue-gray micaceous shale. The top of the member, as defined by Ford and Breed (1973a, p. 1247) is placed below a prominent stromatolitic limestone bed that forms the base of the overlying Jupiter Member. The discoid carbonaceous fossil Chuaria circularis, a possible acritarch, occurs near the top of the Tanner Member (Ford and Breed, 1973b).
The Jupiter Member of the Galeros includes a basal stromatolitic limestone and several dolomite beds in the lower part. However, multicolored shales make up most of the 420 m of the member.

The Carbon Canyon Member of the Galeros consists of alternating limestone beds up to 2 m thick and red or green mudstone and shale of similar thickness. Mudcracks, ripple marks, and salt-crystal casts are common structures in the shale. The Carbon Canyon Member is about 470 m thick and extensively exposed in Carbon Creek, where we will examine it in our foot traverse of Carbon and Lava Canyons (Figure 7).

The uppermost Duppa Member of the Galeros Formation is mainly a purple to gray micaceous shale but includes numerous thin beds of limestone or calcareous siltstone rarely more than 20 cm thick. The Duppa is about 140-170 m thick and includes a single stromatolite horizon in a 60-cm limestone bed 70 m below the top (Ford and Breed, 1973a, p. 1250).

Kwagunt Formation

The Kwagunt Formation is multicolored micaceous shale with subordinate sandstone, limestone, and dolomite. A conspicuous 3.5-m stromatolitic limestone containing the form Boxtonia occurs about 76 m above the base of the formation at the base of the middle Awatubi Member (Ford and Breed, 1973a, p. 1251). Flask-shaped chitinozoanlike microfossils were reported from shale of the upper Walcott Member of the Kwagunt by Bloeser and others (1977).

Sixty Mile Formation

At the top of the Chuar Group is a 60-m sequence of conglomerate, sandstone, and breccia designated as the Sixty Mile Formation (Ford and Breed, 1973a, p. 1253). Outcrops are limited to the north fork of Sixty Mile Canyon and the summit of Nankoweap Butte (Elston and McKee, 1982). Although the formation appears conformable with the underlying Kwagunt Formation, the Sixty Mile clearly records a major change from the low-energy marginal marine conditions reflected by most of the Chuar Group. The coarse clastic deposits of the Sixty Mile Formation are considered to mark the onset of folding, faulting, and uplift associated with the Grand Canyon disturbance that terminated Chuar deposition (Elston and McKee, 1982).

Structure

Strata of the Chuar Group are folded into a broad synclinal structure originally recognized by Walcott (1890). The suncline axis trends nearly north and more or less parallels the Butte fault to the east. Hunt and others (1986) have mapped additional smaller folds paralleling the main syncline. Dips on the west limb of the major fold are gentle but become very steep on the east limb near the Butte fault.

As shown in Figure 8, episodes of movement on the Butte fault have produced as much as 3,200 m of downthrow to the west in pre-Paleozoic time (Elston and McKee, 1982, p. 683), and an estimated 820 m of downthrow to the east in post-Paleozoic time (Walcott, 1890). Local juxtaposition of units along the Butte fault has produced double drag so that nearly vertical beds of the Chuar Group are in contact with vertical Cambrian strata bottom to bottom across the fault (Figure 9).
Age of the Chuar Group

From an analysis of available paleomagnetic, isotopic, and paleontological data Elston and McKee (1982) concluded that the Grand Canyon disturbance, which marked the end of the Chuar deposition, began about 820 Ma. The Chuar is thus older than 820 Ma and younger than the Cardenas Lavas, which are 100 m below the base of the Chuar Group and have yielded a whole-rock Rb-Sr isochron of 1070 ± 70 Ma (Elston and McKee, 1982, p. 689). The Chuar Group is probably the correlative of the Uinta Mountain Group of Utah and the Little Dal Group of the Mackenzie Mountains of northwest Canada.

Middle Proterozoic Unkar Group - Dox Formation

The Dox Formation, originally the Dox Sandstone of Noble (1914), is extensively exposed at river level for 11 miles between mile 63.5 and 74.7 along the "big bend" where the Grand Canyon turns from south to west trends. The Dox is the thickest unit (950 m) in the Unkar Group (Figure 6) and forms most of a large gently tilted homoclinal block of Unkar Group strata. The position and relatively low resistance to erosion of the Dox account for the widening of the canyon and the smooth dark-red to purple-gray slopes of the inner Grand Canyon below Desert View.

Four members of the Dox Formation recognized by Stevenson and Beus (1982) are, in ascending order, the Escalante Creek, Solomon Temple, Comanche Point, and Ochoa Point members (Figures 6 and 10). The lowermost Escalante Creek Member is tan to brown lithic and arkosic sandstone interbedded with green to brown mudstone for a total thickness of 390 m. The lower quartzitic sandstone of the Escalante Creek Member is the most resistant part of the Dox Formation. This unit is very similar to the underlying Shinumo Quartzite even to the occurrence of similar contorted bedding. The two formations are separated by a 12-m dark-green to black shale, which forms a recess at the base of the Dox Formation.

The Solomon Temple Member is a 280-m-thick succession of red siltstone and fine-grained sandstone that mostly forms smooth slopes. The upper 66 m of the member are more resistant channel-sandstone beds having prominent tabular-planar cross-beds.

The Comanche Point Member is 160-188 m thick and composed of interbedded red quartz sandstone and siltstone in thin beds. The Comanche Point occupies more than half the total Dox outcrop in eastern Grand Canyon and forms relatively smooth red-brown slopes. Five distinct light-gray to grayish-red marker beds 3-12 m thick and composed of bleached sandstone occur in the upper two-thirds of the member. A stromatolitic dolomite bed occurs directly above the lowest light-gray marker bed. Mud cracks, ripple marks, and salt-crystal casts are common in the Comanche Point Member.

The uppermost Ochoa Point Member is composed of 91 m of reddish-brown micaeous siltstone with thin interbeds of fine-grained sandstone. Siltstone predominates in the lower half of the member but decreases upward as the sandstone content increases, so that the upper 30 m are mostly sandstone and form alternating ledges and slopes. Mud cracks and salt-crystal casts occur in the lower half; asymmetrical ripples and small-scale cross-beds are more common in the upper sandstone beds of the Ochoa Point Member.

Deposition of the Dox Formation probably occurred in a variety of marginal marine settings beginning with lagoonal and delta-front environments and ending with low to high tidal flats. Dox deposition ended when the overlying Cardenas Lavas flowed out over the intertidal sand flats (Hendricks and Lucchitta, 1974; Lucchitta and Hendricks, 1983). The age of at least the upper part of the Dox is approximately that of the overlying lavas, which are dated at about 1070 Ma.
MIDDLE PROTEROZOIC CARDENAS LAVA

Description

The basaltic Cardenas Lava of the Unkar Group (previously called Cardenas Lavas) forms somber cliffs (Figure 11) in the lower part of the canyon on both sides of the Colorado River from Chuar Lava Hill (mile 65) downstream to near Unkar Creek (mile 72). The lava is about 300 m thick and occurs near the middle of the 4000-m-thick Grand Canyon Supergroup of Middle Proterozoic age (Figure 6).

The age of the lava is discussed by Lucchitta and Hendricks (1983), as are the general characteristics of the section. Briefly, a Rb-Sr isochron has yielded an age of about 1.1 Ga, which is considered a cooling age. K-Ar ages of about 800 Ma are thought to be reset ages reflecting maximum burial under rocks of the Chuar Group, the highest unit of the Grand Canyon Supergroup. More recently, Larson and others (1986) suggest eruption and diagenesis dates of 950 and 715 Ma, respectively, on the basis of variation of major and trace elements.

The lava column is made up of at least 7 to as many as 15 flows. The number varies from place to place and commonly is difficult to determine exactly, owing to poor exposures.

The lower 90-100 m of the lava consist of a highly weathered, slope-forming, bottle-green member that contains a high proportion of fractured and devitrified glass, now with a high sodium content probably owing to syn- or postdepositional alteration. Ophitic texture is preserved in places. Spheroidal nodules are common and probably the result of weathering.

The upper 200 m of lava is composed of basalt and basaltic andesite. Several distinctive members can be recognized in this part of the section, the most noteworthy being a lapillite containing bombs. The members are laterally persistent and can be traced through most of the area of outcrop. All units display alteration, in part probably contemporaneous with emplacement, in part due to burial. In spite of the alteration, many primary structures are remarkably well preserved.

FIGURE 11. View of Cardenas Lava within tilted Middle Proterozoic section of the Grand Canyon Supergroup, viewed from the south. Basalt Canyon near center of photograph.
The lava and the rest of the Grand Canyon Supergroup are exposed in fault blocks tilted northeast and truncated by the angular unconformity beneath the nearly horizontal Tapeats Sandstone of Cambrian age (Figure 11). Because of this, progressively younger Proterozoic units are preserved to the northeast. Within the Cardenas, the lapillite is the highest member preserved in the southwesternmost exposures (Basalt Cliffs), whereas the same unit occurs only a little above the middle of the section at Chuar Lava Hill, at the northern limit of exposure.

Individual units are thickest at Ochoa Point, where the lava section also reaches its greatest cumulative thickness. Vent-facies rocks are present locally in this area. These features suggest that the vents for the lavas now preserved were in the area of Ochoa Point.

The base of the Cardenas Lava is conformable with the underlying Dox Sandstone, with which it interfingers locally. Interbeds of sandstone occur throughout the lava section. Those in the lower part are well sorted and fine grained and resemble the Dox. Many of these sandstone beds are laterally persistent; others are very restricted. Abrupt channels whose fill does not differ markedly from the material elsewhere in the beds are a common feature of the sandstone beds. A reasonable interpretation of these features is that they are tidal-flat deposits.

The basal contact, the sandstone interbeds, and quenching features such as fractured glass in the matrix and closely spaced joints all suggest that the lava was emplaced in the shallow and saline Dox sea, whose floor sank at the same rate as the lava pile built up so that the top of the lava remained near sea level. Only late in the time represented by the lava section did the vents rise above sea level to produce the tephra represented by the lapillite unit.

Alternatively, this unit may represent an unusually explosive eruption from a vent that remained covered by shallow sea water.

**Basalt Canyon**

Basalt Canyon (Mile 69.5) is an excellent place to view a representative section of the Cardenas Lava (Figure 12). The part of the canyon that is easily accessible represents a modest walk of little difficulty for most people. Only three places may cause problems and need to be pointed out. The first occurs at a major fork of Basalt Canyon, about one mile upstream from its mouth. The right-hand fork is the one that usually contains water. Take the left fork (going upstream). The next place is at a waterfall formed by a 6-m sandstone bed. This waterfall can be circumvented to the left. The final place is a second waterfall, in basalt, several hundred yards above the first. This waterfall can be climbed, but it is easier and safer to bypass it on the left by going up a tributary wash about 100 yards downstream from the waterfall, then contouring back to the main canyon above the waterfall. The end of the walk occurs in a box canyon from which there is no exit without technical climbing.

It is well to keep in mind that rattlesnakes are common in this area during the warm season, especially near the lower waterfall. An additional hazard is posed by the lava, which weathers into hard sloping surfaces mantled by loose pebbles that act as ball bearings and make for precarious footing.

The first part of the walk is through the Dox Sandstone, which shows interesting sedimentary structures, including salt casts. Debris-flow deposits are conspicuous in the wash, a reminder that such flows are a common occurrence in this canyon.

**Figure 12.** Representative section of Cardenas Lava. Sandstone beds within the lava are shown stippled and thicker than scale for clarity.

The Dox-Cardenas contact is well exposed on the left side of the wash. This outcrop shows the baked zone in the Dox, the chilled zone in the lava, and the planar and conformable character of the contact. The outcrop, however, does not show the soft-sediment deformation in the Dox at the contact, nor the intra­Dox flow. These features, which demonstrate that the upper Dox and lower Cardenas are contemporaneous, are impressively exposed elsewhere, but these areas are remote and not easy to reach.

The rubbly slopes above the contact are underlain by the bottle-green lower member of the Cardenas. Such slopes are typical for this poorly resistant member. As the canyon narrows and exposures improve, spheroidal nodules can be seen in the canyon walls. These nodules have been interpreted variously as the product of weathering and as pillows. The former interpretation is most likely, but some nodules do show radial and concentric cracks and glassy selvages suggestive of pillows. Nodules showing pillowlike structures are particularly well exposed in the right fork of Basalt Canyon (mentioned above).

The prominent 6-m sandstone at the lower waterfall marks the top of the bottle-green unit. Ripple marks are well developed on the lower face of this sandstone. Directly above the waterfall are good exposures of the fan-jointed member, which is laterally persistent and reminiscent of flows that have been rapidly quenched.

The second waterfall is made by dense basalt flows with few vesicles and typically forming cliffs on hill slopes. These flows probably were emplaced as fluid lava: traces ofropy lava are preserved on the left a few tens of meters stratigraphically above the
top of the waterfall, andropy lava is prominent in this part of the section in other exposures. Copper streaks are visible upon careful examination at the top of the waterfall.

The lapillite is exposed on the right side of the canyon beyond a prominent fault, about 1/4 mile upstream from the upper waterfall, and about 100 m above it stratigraphically. The unit has a distinctive gnarly appearance and a pinkish cast. These characteristics make it stand out even at a distance. Ash and highly vesicular lapilli are readily visible. The larger fragments are bombs, some of which are fusiform.

The lapillite is the highest unit that can be reached easily and safely. Above the lapillite at Basalt Canyon are about 50 m of basaltic flows and one sandstone interbed. The sedimentary unit above the Cardenas is the Nankoweap Formation, part of the Grand Canyon Supergroup. The lowest part of this formation, visible on the rim of the cliff, is a purplish-to-dirty-white quartz sandstone grading downward into a basal conglomerate derived chiefly from the underlying Cardenas Lava. Cardenas and Nankoweap are separated by an erosional surface that is a disconformity to slightly angular unconformity.

Returning down the canyon, note how the sandstone interbeds in the upper part of the section are coarser grained, more poorly sorted, and more cross-bedded than those in the lower part. The upper beds also contain rubble and grains derived from the lavas. The source for the quartz sand in the sandstone beds is an interesting and as yet unresolved problem because the beds are separated from the Box Sandstone, which contains similar sandstone, by the lava, and from the basement rocks, which are an obvious source, by about 1600 m of Unkar sedimentary rocks.

Excellent views of channels in the sandstone beds can be seen from good vantage points on the way down.

**DYNAMIC RIVER BEACHES**

Sandy alluvial terraces and bars along the Colorado River in the Grand Canyon serve as convenient campsites and lunch stops for some 15,000 people each year and thus constitute a major recreational resource in the National Park. The terraces and bars were formed by a dynamic river system that deposited the beaches mainly during seasonal high-water flows before completion of Glen Canyon Dam. Since completion of the dam in 1963, the river regimen in Grand Canyon has changed dramatically. Whereas predam seasonal high-water flows annually reached about 86,000 cfs*, post-dam discharge rates have been carefully controlled and have rarely been permitted to exceed 28,000 cfs (level for maximum power generation at the dam). Beach profile surveys during the past 11 years (Howard, 1975; Dolan, 1981; Beus and others, 1985) indicate that many beaches are gradually losing sand that may never be replaced under the present controlled flow of the river and in the absence of periodic high-water floods of sediment-enriched water. In the summer of 1983 and again in 1984, unexpected high and rapid spring runoff produced a full reservoir in Lake Powell and required releases of high-water flows through the Grand Canyon. These high-water releases attained 40,000 to 96,000 cfs and were maintained for periods of several weeks or months. The river washed over all the campsites and lunch stops for some 15,000 people each year and thus constitute a major recreational resource in the National Park.

* cubic feet per second

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**FIGURE 13**. Grapevine beach, mile 81.1. View upstream to the east.

**FIGURE 14**. Topographic beach profiles measured across downstream end of Grapevine beach.

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During the 11-year period (Beus and others, 1985), essentially all the gain came during the 1983 and 1984 high-water periods.

Grapevine beach (mile 81.1), where we plan to camp on day 6, has experienced both gain and loss from the high-water "spills" in 1983 and 1984 (see Figures 13 and 14). There was almost no change at Grapevine between 1974 and 1980, a major addition in 1983, a loss in 1984, and a minor addition in 1985. Clearly the beach has not been very stable under river-flow conditions of the past 3 years.

A preliminary conclusion from studies in progress is that wise management of the Colorado River flow through Grand Canyon may require periodic high-water discharges at as-yet undetermined rates and lengths to assure adequate sand on campsites in the future.
HISTORY OF THE COLORADO RIVER AND ITS GRAND CANYON

Like the Nile River of Africa, the Colorado obtains most of its water from mountain ranges that are far removed from the parched lands that border much of its course. Because of this, the river has become a lifeline that makes it possible to cultivate large areas that otherwise would be desert and to establish or maintain cities such as Los Angeles, Las Vegas, and Phoenix that could neither flourish nor reach their present size in its absence.

The great burden placed on the Colorado, combined with the finiteness of its waters, has resulted in a host of social, political, and engineering problems that have centered chiefly on how many reservoirs should be built and who should get the water. For geologists, however, the river has been the source of quite different interests and controversies ever since it was discovered. The most pressing of these questions are:

When and how did the river come into being?

When did canyon cutting and correlative uplift occur?

How quickly was the Grand Canyon cut?

Why and how did the river so perversely cut across the many belts of high ground that sit astride its course?

Most recently, another question has arisen in connection with environmental impacts produced by Glen Canyon Dam, which impounds Lake Powell. This question concerns the destruction and construction of beaches along the river edge and will be discussed elsewhere in this guide.

To understand clearly what follows, it is important to remember that there are two contrasting views regarding the history of the Colorado River. The first is that the river from birth has been part of an integrated drainage system with a course approximating the present one; the second is that the river is a dynamic entity changing through time and constructed from segments of preexisting drainages with courses potentially quite different from the present one.

According to the first view, the river came into being pretty much as it is today at some particular time such as the Eocene. A statement that is made about any part of the river applies to the river as a whole: the entire river is young, or old, as the case may be. The second view is that one cannot talk about the birth of a river because most rivers are continually changing entities that have evolved from various ancestors and will continue evolving into progeny whose configuration depends on factors such as tectonism and climate. The answer, then, to the question "When was the river born?" can only be another question: "How much departure from the present configuration is one willing to tolerate and still speak of the Colorado?"

Lastly, it is important to remember that the river traverses two contrasting terranes in Arizona. The first is the Canyon country, typified by the Grand Canyon. It is highly dissected and commonly has much relief. The second is the Plateau country, typified by most of the Navajo Reservation and characterized by low relief, wide mature valleys, and scarps developed on beds of contrasting resistance that areretreating down structural slopes. The Plateau country is more widespread than the Canyon country. The two terranes are in uneasy equilibrium at their boundaries with each other. The drive from Flagstaff to the embarcation point at Lees Ferry is along one such boundary.

For the first 60 years or so after Powell's journey of discovery (1875), geologists subscribed to the idea of a river with a simple history: it was born with the same course that it has now. The question was: when did this happen and when did the uplift of the region, which was considered responsible for the cutting of the canyons, occur? Impressed by the observation that the Plateau country is an area of pervasive erosion, these geologists inferred that this erosion was also deep — the "Great Denudation" of Dutton (1882) — and thus old. Thus, the canyon cutting that produced the erosion and the uplift ultimately responsible for both must have occurred a long time ago, presumably shortly after retreat of the great inland seas at the beginning of the Tertiary Period. According to this view, the Colorado River, the uplift, the canyon cutting, and the Great Denudation all began in Eocene time and perhaps even earlier in the Tertiary.

The origin of the river and the Grand Canyon seemed safely established. Attention, therefore, was focused on geomorphic problems highlighted by the textbooklike character of the Grand Canyon region, where sparse vegetation and simple structure make it possible to see landforms clearly and trace them for great distances. These characteristics led to the development of several concepts of fundamental importance in geomorphology, among which are the principles of antecedence, superposition, consequence, and anteposition, all having to do with the problem of the relation between drainage systems, structure, and topography (Davis, 1901, 1903; Babenroth and Strahler, 1945; Strahler, 1946).

Storm warnings signalling danger for these views were hoisted in the 1930's and 1940's by people mapping in the Basin and Range country, where interior-basin deposits of late Miocene and Pliocene age are common throughout the course of the Colorado River, and there is no evidence for an older drainage system that could be called the Colorado. In conformity with the concept of a monophase history for the river, these people concluded that the entire river, and thus the Grand Canyon as well, were no older than late Tertiary (Blackwelder, 1934; Longwell, 1936, 1946).

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

Under the sponsorship of E. D. McKee and the Museum of Northern Arizona, Lucchitta and Young attempted to shed some light on the paradox by studying critical areas at and near the mouth of the Grand Canyon. These authors (Lucchitta, 1966; Young, 1966) jointly showed that indeed there is no stratigraphic or morphologic evidence at the mouth of the Grand Canyon for a through-flowing drainage system during deposition of Miocene interior-basin deposits related to Basin-Range deformation. Nor could this difficulty be bypassed by looking elsewhere along the course of the lower Colorado River or the southwest margin of the Colorado Plateau in Arizona. In this area interior-basin deposits are ubiquitous, and deposits older than the Basin-Range event record drainage northeastward from what is now the Basin and Range Province onto what is now the Colorado Plateau. The northeast drainage direction existed as recently...
as the time of the Peach Springs Tuff, a 17-18 Ma ignimbrite that flowed onto the Colorado Plateau. Before Basin-Range faulting, therefore, drainage was not to the west or southeast, as would be required for a river with a course similar to that of the present Colorado, but in the opposite direction, to the northeast.

At this point the idea of a polyphase history for the river gradually began to take hold. Hün (1969) contributed the concept of drainage systems initially departing markedly from the present Colorado River, but gradually evolving into this configuration. However, Hunt, as well as Lovejoy (1980), continued to advocate an ancient river flowing westward from the Colorado Plateau even before Basin-Range deformation, a concept contradicted by the evidence mentioned above.

In 1967, an important contribution by McKee and others for the first time presented in fully developed form the concept of a polyphase history for the Colorado River. These authors accepted the antiquity of the upper part of the drainage system, as documented by Hunt, but could not accept a continuation of this drainage westward through the Grand Canyon into the Basin and Range Province because there is no evidence to support such a course. Instead, they proposed that the ancestral Colorado followed roughly its present course as far as the eastern end of the Grand Canyon, but then continued not westward, but southeastward along the course of the present Little Colorado and Rio Grande Rivers into the Gulf of Mexico. In Pliocene time, a youthful stream, emptying into the newly formed Gulf of California and invigorated by its shortness and consequent steep gradient, eroded headward and captured the sluggish ancestral river somewhere in the eastern Grand Canyon area. It was then that the river became established in its present course and the Grand Canyon was carved.

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

One aspect of the paper by McKee and others (1967) has not stood the test of time because evidence accumulated since the paper was written does not support the concept of drainage southeast along the Little Colorado and Rio Grande Rivers. On the other hand, evidence has continued to accumulate that an ancient river could not flow through the western Grand Canyon region into the nearby Basin and Range Province (Lucchitta, 1972, 1975; Young, 1970; Young and Brennan, 1974).

Analysis of deposits along the course of the lower Colorado River in the Basin and Range Province confirmed that this part of the river is no older than latest Miocene, and showed that the capture of the ancestral Colorado River is documented by the sudden appearance within river deposits in the Salton trough (upper Tertiary Imperial Formation) of coccoliths otherwise found only in the Cretaceous Mancos Shale of the Colorado Plateau (Lucchitta, 1972).

An attempt to synthesize information currently known led Lucchitta (1975, 1984) to postulate that the ancestral Colorado did not flow southeast as proposed by McKee and others. Instead, the river crossed the Kaibab upwarp along the present course of the Grand Canyon, then continued northwestern along a strike valley in the area of the Kanab, Uinkaret, or Shiwits Plateaus to an as yet unknown destination. After the opening of the Gulf of California, this ancestral drainage was captured west of the Kaibab upwarp by the lower Colorado drainage. According to this concept, the upper part of the Grand Canyon in the Kaibab Plateau area is old and related to the ancestral river, whereas the lower part of the canyon in this area and all of the western Grand Canyon postdates the capture and was carved in a few million years, a process aided by nearly 1 km of regional uplift since inception of the lower river (Lucchitta, 1979).

The hypothesis is based on three observations. (1) Gravels of probable river origin are present in the area of the Kanab, Uinkaret, and Shiwits Plateaus. (2) Northwest-trending drainages along strike valleys were common and persistent before canyon cutting, as evidenced by fossil valleys preserved under Miocene lavas in many places in the southwestern Colorado Plateau and also by ancient valleys in the plateau country. Examples of such valleys are those of Cataract Creek and the Little Colorado River, which predate canyon cutting and have not yet been appreciably affected by it. (3) If Mesozoic rocks eroded in the last few million years are restored, a river such as the ancestral Colorado could have flowed westward across the Kaibab upwarp with no great difficulty and would have done so in an arcuate racetrack located just where the current arcuate Grand Canyon occurs.

The history of the Colorado River, as currently understood, contains many instructive lessons regarding geologic processes associated with drainage systems. One is that the process of downcutting, even in the hard Paleozoic rocks of the Grand Canyon, can occur at astonishingly high rates: 1 1/2 km in just a few million years. The key here, perhaps, is that in canyons the volume of material removed per unit of downcutting is small compared to that for more open valleys. In other words, the rate at which material is carved from a canyon is small in relation to the rate at which the floor of the canyon is lowered (Lucchitta, 1966).

Even more interesting is the concept that physical processes such as drainage evolution can obey a Darwinian law based on natural selection and survival of the fittest, in close analogy to what happens in biological systems. With rivers, the agent of change is tectonism, not random mutations; competition occurs through changes in gradient: rivers whose gradient has been reduced are handicapped; those whose gradient is increased are favored; and the weapons with which the battles for survival are fought are headward erosion and capture. It is by such means that drainage networks constantly change and evolve from ancestors that may bear but little family resemblance to their scions. Any particular configuration is merely the crystallization of a specific moment within the chain of evolution. And when that particular configuration was born is a matter of definition, not of time.
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INTRODUCTION
This trip into the western Grand Canyon—down the lower Colorado River to Lake Mead—gives us the opportunity to see a classic geologic cross section. The Colorado River in western Grand Canyon has exposed the Precambrian and Paleozoic rocks of the southwestern margin of the Colorado Plateau Province. Cenozoic rocks (Paleocene to Holocene) are present in several tributary canyons, on adjacent plateaus, and in the Grand Wash trough immediately west of the mouth of Grand Canyon.

This field guide briefly illustrates and discusses the geologic history and some of the wide spectrum of geologic rock units and structural features of the western Grand Canyon region. (The structural geology is emphasized in another field guide in this volume by Peter Huntoon and others.) Also, trip participants will receive three geologic maps, each containing detailed descriptions of the Paleozoic geology and other geologic information.

The field trip (all in Arizona) begins in Flagstaff. Our bus will follow Interstate 40 west to Seligman and continue on State Highway 66 (old US 66) to Peach Springs. Several geologic points of interest along the way will be briefly discussed on the bus. From Peach Springs, we will follow a dirt road north down Peach Springs Canyon for 20 miles to the Colorado River. Here we will board river rafts and begin the float trip into the depths of the Lower Granite Gorge of western Grand Canyon. On the third day we will leave the Grand Canyon and the Colorado Plateau and enter the Basin and Range Province. The raft trip ends at Pierce Ferry, where we will make a final stop at Airport Point for an excellent view of the Colorado Plateau-Basin and Range Transition. We will then return to Flagstaff by bus. Our route and stops are shown in Figure 1.

PRECAMBRIAN GEOLOGY
Precambrian rocks are exposed in the depths of Grand Canyon along the Lower Granite Gorge and some tributary canyons. Malcome D. Clark, who has conducted the only study of these rocks (1976), summarized them as follows: "The Precambrian crystalline rocks in the Lower Granite Gorge are divided into widespread schists and associated lithologies, and a suite of intrusive granitic rocks, similar to the crystalline rocks in the Upper Granite Gorge of eastern Grand Canyon. It is therefore convenient to retain the names Vishnu Group and Zoroaster Plutonic Complex in this region for both of these groups respectively."
Figure 1. Location map of field-trip route and stops.
PALEozoIC AND MESozoIC GEOLOGY
Detailed descriptions of the Paleozoic rocks are given in the handouts. These rocks are sedimentary strata consisting of sandstone, shale, and limestone that accumulated with several hiatuses, from Early Cambrian to Early Permian time. (Strata of Ordovician and Silurian age are not present in the Grand Canyon area; their position in the section is marked by a regional disconformity.) Cambrian, Devonian, and Mississippian rocks are the most widely exposed Paleozoic units on the Hualapai Plateau (see Figure 1 for plateau locations). Rocks of Pennsylvanian and Permian age form the surface of the Sanup Plateau. The resistant Fossil Mountain Member of the Kaibab Formation (Permian) forms the surface of the Shiwits Plateau that is in turn partly covered by Tertiary basalt flows.

Outcrops of Mesozoic rock are confined to a small erosional channel in upper Surprise Canyon on the Shiwits Plateau. This rock is a conglomerate of the Timpoewap Member of the Moenkopi Formation of Early Triassic age.

All of the eras represented in the western Grand Canyon are summarized in Figure 2.

CENOZOIC GEOLOGY
The beveled surface of the Hualapai Plateau was eroded during the Laramide Orogeny (Late Cretaceous to Eocene; Young, 1985). The erosional valleys carved into the plateau are partly filled with Tertiary sediments and Miocene volcanic rocks (Figure 3). The Tertiary valleys originated south and west of the Colorado Plateau in the western and central highlands of Arizona between Kingman and Prescott (Young, 1966, 1970, 1985; Young and McKee, 1978). The paleodrainages flowed north and northeast down a structural and topographic slope, cutting through northeast-dipping Paleozoic strata. After losing most of their gradient, the drainage meandered out onto a lowland area of gently dipping Paleozoic and Mesozoic strata, now the southwestern margin of the Colorado Plateau. The meandering loops of one ancient Tertiary drainage near Peach Springs Canyon can be seen on the geologic maps of Billingsley and others (1986b). The posttectonic paleodrainages flowed perpendicular to structurally and lithologically controlled northwest-trending cuestas and strike valleys (Lucchitta, 1975, 1979; Young, 1979, 1982, 1985; Young and Hartman, 1984). Erosion into the Precambrian crystalline rocks south of the lowland area brought clasts of schists, quartzites, and granites into the paleovalleys after major Laramide tectonic activity slowed about 55 million years ago (Huntoon, 1981; Young, 1985). Shortly after Laramide time, erosion had cut down into Cambrian rocks in the Peach Springs Canyon area and elsewhere on the Colorado Plateau. Subsequent tributary drainages continued to erode headward along strike valleys on the Colorado Plateau, locally depositing Paleozoic clastic material. Farther north in the Peach Springs Canyon area, the Tertiary valley floors were cut into younger Paleozoic rocks because the gradient of the valley was not as steep as the dip of Paleozoic rocks. Evidence for paleodrainages north of Diamond Creek (Figure 1) is missing because of the relatively recent erosion of the modern Grand Canyon. However, about 20 miles northeast of Peach Springs and Diamond Creek Canyons, Tertiary deposits consisting of Paleozoic and Precambrian lithologic clasts are found in paleovalleys cut into Permian rocks (Koons, 1948, 1964).

Transportation of sediments from the south and southwest down the Tertiary valleys was gradually terminated by middle Tertiary faulting (down to the west) along the Grand Wash fault (represented by the trend of the Grand Wash Cliffs in Figure 1), but before the valleys filled with several hundred feet of materials derived from Precambrian and Paleozoic rocks. After the Tertiary valleys were disrupted along the Grand Wash fault, some reversed drainage direction and began draining southwest. Headward erosion on the Colorado Plateau accelerated aggressively into the retreating scarps of the now uplifted plateau during the development of the relatively younger Grand Canyon drainage system (Young, 1985; Lucchitta and Young, 1986).

During the post-Laramide valley filling, tectonic activity occurred mainly south and west of the Colorado Plateau. Minor movement occurred along the Hurricane fault, the largest in the area, just prior to development of the paleodrainage along the strike of the fault. This movement resulted in monocline folding down to the east, in a manner consistent with many of the other major structures on the Colorado Plateau. Later periods of faulting would partly explain the disruption of the highland paleovalleys east of Peach Springs Canyon (Billingsley and others, 1986a). Subsequent drainage capture along the Hurricane fault cut off all the meanders on the east side of Peach Springs Canyon.

About 1,000 feet of the Buck and Doe Conglomerate and the Westwater Formation nearly filled the paleovalleys in upper Peach Springs Canyon prior to emplacement of the Peach Springs Tuff about 19.3 to 19.6 million years ago (Glazer and others, 1986, p. 842). These relations are generalized in Figure 3. Major movement along the Hurricane fault probably began in late Miocene or early Pliocene time and continued into the Pleistocene. All of the pre-Pleistocene sediments in the Peach Springs Canyon area are offset as much as 40 miles, erosion of the underlying Paleozoic rocks.

The development of the Colorado River system has occurred only during the last 4 million years (Lucchitta and Young, 1986). A summary of this development is presented at the last stop of the third day of the trip.

ROAD LOG
Day One
The group will leave Flagstaff and travel west for 76 miles on Interstate 40 to Seligman. From
**Description**

<table>
<thead>
<tr>
<th>Formation and Member</th>
<th>Average Thickness of Member in Feet</th>
<th>Columnar Section</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>MOENKOPI FORMATION</td>
<td>100</td>
<td></td>
<td>Light-gray conglomerate in reddish-gray matrix of siltstone and coarse-grained sandstone</td>
</tr>
<tr>
<td>KAIBAB FORMATION</td>
<td>300</td>
<td></td>
<td>Yellowish-gray to pale-red shale and medium-grained sandstone interbedded with gray, thin-beded, fossiliferous limestone and white, thick-beded gypsum</td>
</tr>
<tr>
<td>Fossil Mountain</td>
<td>250</td>
<td></td>
<td>Light-gray cherty limestone and sandy dolomia</td>
</tr>
<tr>
<td>TOROWEAP FORMATION</td>
<td>100</td>
<td></td>
<td>Gypsumous pale-red and gray siltstone and medium-grained sandstone interbedded with dark-gray, thin-beded limestone</td>
</tr>
<tr>
<td>Saligman Member</td>
<td>40</td>
<td></td>
<td>Yellowish-white to pale-red, fine-grained, thin-beded sandstone</td>
</tr>
<tr>
<td>COCONINO SANDSTONE</td>
<td>10</td>
<td></td>
<td>Light-brown to yellowish-red, fine-grained sandstone; large-scale, tabular cross-stratification</td>
</tr>
<tr>
<td>HERMIT SHALE</td>
<td></td>
<td></td>
<td>Reddish-brown and white siltstone and fine-grained, thin-beded sandstone</td>
</tr>
<tr>
<td>PAKOON LIMESTONE</td>
<td>200</td>
<td></td>
<td>White to pale-red, medium- to fine-grained, medium-beded, coarse-grained sandstone and siltstone, thin-bedded siltstone, interbedded with gray, thin-beded, fossiliferous limestone of Pakoon limestone which is not considered to be part of the Supai but is equivalent to the lower Esplanade</td>
</tr>
<tr>
<td>WESCOSGAME FORMATION</td>
<td>50</td>
<td></td>
<td>Westcoisen, Manakacha, and Watahomigi formations: Pale-red siltstone and grayish-red, fine-grained, cross-stratified sandstone, interbedded with grayish, thin-beded limestone and dolomitic sandstone; lower limestone contains lenses of red and white chert</td>
</tr>
<tr>
<td>MANAKACHA FORMATION</td>
<td>130</td>
<td></td>
<td>Supai upper part is dark reddish-brown, fine-grained sandstone and siltstone, thin-beded; some thin beds of gray limestone. Middle part is a yellowish-gray, coarsely crystalline, silty, crusty, thin-beded, fossiliferous limestone that is interbedded with underlying pale-yellow and reddish-brown, cross-stratified sandstone. Basal part is dark reddish-brown, iron stained, chert pebble conglomerate in matrix of brown, coarse-grained sandstone that grades upward into pale yellow and reddish-brown, cross-stratified sandstone</td>
</tr>
<tr>
<td>WATAHOMIGI FORMATION</td>
<td>200</td>
<td></td>
<td>Redwall limestone: Light to dark gray, inapparent, thick to medium-beded, fossiliferous limestone and dolomite, aleund beds and lenses of gray to white fossiliferous chert</td>
</tr>
<tr>
<td>SUPAI FORMATION</td>
<td>300</td>
<td></td>
<td>Upper part is dark reddish-brown, fine-grained sandstone and siltstone, thin-beded; some thin beds of gray limestone. Middle part is yellowish-gray, coarsely crystalline, silty, crusty, thin-beded, fossiliferous limestone that is interbedded with underlying pale yellow and reddish brown, cross-stratified sandstone. Basal part is dark reddish-brown, iron stained, chert pebble conglomerate in matrix of brown, coarse-grained sandstone that grades upward into pale yellow and reddish brown, cross-stratified sandstone</td>
</tr>
<tr>
<td>TEMPLE BUTTE FORMATION</td>
<td>130</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>classified dolomites</td>
<td></td>
<td></td>
<td>Middle member: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Hovasu Member</td>
<td>130</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Kanab Canyon Member</td>
<td>40</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Peach Spring Member</td>
<td>100</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Spencer Canyon Member</td>
<td>80</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Sanup Plateau Member</td>
<td>40</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Rampart Lave Member</td>
<td>100</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Flower Rock Member</td>
<td>80</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Meriwitche Member</td>
<td>100</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>Faded brown sandstone member</td>
<td>150</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>unnamed shale</td>
<td>35</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
<tr>
<td>TAPEATS SANDSTONE</td>
<td>50</td>
<td></td>
<td>Upright limestone: Light to dark gray, locally pink, mottled, inapparent, thin-beded dolomite and limestone; weathers dark gray to rusty brown with sugary texture; interbedded with gray to brown, platy siltstone and green glauconitic sandstone</td>
</tr>
</tbody>
</table>

Figure 2. Stratigraphic column of the western Grand Canyon, Arizona.
Seligman, we will continue west on old Highway 66 for 34 miles to the town of Peach Springs. Here we will turn right onto a paved street on the east side of Peach Springs and continue through a residential section toward the northwest side of town to where the Diamond Creek road begins its descent into Peach Springs Wash, a tributary to Peach Springs Canyon. Road mileage starts at the Highway 66 turnoff. Odometer 0.0. Turn right onto paved street.

0.1 Elevation 4,950 feet. Hualapai Tribal Offices on right. Continue straight ahead and then, at 0.2 mile, turn right at second street.

The boundaries of the original Hualapai Reservation were established by Executive Order of President Chester A. Arthur on January 4, 1883. Subsequent additions—in the Big Sandy River valley (about 30 miles south of Peach Springs), a small area at Valentine, and disputed lands deeded to the tribe by the Santa Fe Railroad—enlarged the reservation to nearly 1 million acres in 1968. Peach Springs was established as a railroad town about the year 1900 and is now the headquarters of the Hualapai Tribe.

The town is built on gravels of the Willow Springs Formation of Pliocene to Pleistocene(? ) age (Young, 1966, p. 51). Clasts in these gravels are locally derived from Paleozoic rocks exposed on the ridge east of town. The Paleozoic strata dip east about 8° and represent the southermmost extension of the Toroweap monocline (Sillingsley and others, 1986a,b). Most Willow Springs clasts in this area are limestone and dolomite of the Temple Butte Formation and the Redwall Limestone; a few are red sandstone and gray limestone from the Watahomigi Formation. The Willow Springs is here 40 to 75 feet thick and overlies a basalt flow of unknown age (Figure 3). Beneath the cover of Cenozoic rocks is the contact of the Devonian Temple Butte Formation and the Cambrian unclassified dolomites.

0.6 Begin descent into Peach Springs Wash.

0.8 Basalt flow in road cut originated somewhere on the higher plateau southeast of Peach Springs and flowed down Yampai Canyon. The lava spread out at the canyon mouth and accumulated to a thickness of nearly 130 feet just west of what is now downtown Peach Springs. Some basalt ponded in this area, but most of the flow traveled southwestward several miles into the Truxton Valley. Gravels and sandstones of probable Miocene age underlie the flow, which is, in turn, covered by more fanglomerates and conglomerates of the Willow Springs Formation. A series of small northeast-trending synclines and anticlines occur in the basalts and underlying Tertiary sediments. The origin of these folds is unknown, but they are postulated to be due in part to collapse by solution of finer grained calcareous Miocene sediments of the Westwater Formation (Young, 1966, p. 32). There is no evidence of folding in the nearby exposed Cambrian bedrock.

1.4 Peach Springs Tuff at 12 o'clock, forming a 12-foot, grayish-tan cliff. This tuff is an important marker bed for time correlation of Cenozoic deposits in Peach Springs Canyon and other
tributary canyons on the Hualapai Plateau. The Peach Springs Tuff (Young, 1966; Young and Brennan 1974) has been dated at about 18.5 to 19.6 million years (Glanner and others, 1986, p. 842). Below the tuff, the road cuts through fine-grained calcareous sandstone and siltstone of the Westwater Formation and conglomerates and fanglomerates of the Buck and Doe and Music Mountain Conglomerates (Figure 3).

1.5 The Spencer Canyon and Sanup Plateau Members of the Muav Limestone (Cambrian) crop out on Billingsley and others marked by large boulders, is exposed in cliffs on both sides of the wash under a veneer of lag gravels (see Figure 2 for stratigraphic position).

2.3 Music Mountain Conglomerate (rounded clasts) and Hindu Fanglomerate (angular clasts) on both sides of Peach Springs Wash (Young, 1966, p. 24, 30).

3.4 Peach Springs. Water for the town of Peach Springs is pumped from these springs. Most springs in western Grand Canyon discharge from the Rampart Cave Member of the Muav Limestone (Twenter, 1962). Here, the water may come from the top of this member or from the Sanup Plateau Member (Figure 2).

3.6 Rampart Cave Member on left side of wash.

4.3 Lower Peach Springs, an overflow from Peach Springs upstream. The water here flows a short distance on the surface on impermeable rocks of the Bright Angel Shale. An erosional unconformity, marked by large boulders, is exposed in cliffs on both sides of the wash (Figure 4). This unconformity is in the lower part of the Rampart Cave Member of the Muav Limestone. The basal contact of the Rampart Cave is placed at the change to greenish-gray, thin-bedded dolomites and shales of the Flour Sack Member of the Bright Angel Shale. The Rampart Cave here is thinning from the north and northwest, and it pinches out about 4 miles south of this location. Several other members of the Muav gradually thin and pinch out about 20 miles to the south. The Bright Angel continues a bit farther south than the Muav but eventually it too pinches out. The resulting unconformity, farther south, places the Devonian Temple Butte Formation first on the Muav, then on the Bright Angel, and eventually on the Rampart Sandstone. The Tincanebits Member of the Muav Limestone and Bright Angel Shale.

The valley occupied by the Music Mountain sediments is nearly six times that of background values measured in nearby outcrop of the Muav Limestone and Bright Angel Shale. The valley is deeper on the right side of the Tincanebits Member and above the redbrown sandstone member. The Tincanebits Member and above the redbrown sandstone member of the Bright Angel Shale is absent in this area, but thin limestone beds of the Tincanebits Member are.

4.5 Entrance to Peach Springs Canyon; elevation 1,300 feet. The road here follows the Hurricane fault, and we will have a good view down Peach Springs Canyon. The transition member of the Rampart Cave is exposed in the road cut on right. That part of the Bright Angel on the left (west) side of road is stratigraphically between the Flour Sack member and the redbrown sandstone member of the Bright Angel. Nearly 400 feet are offset here on the hurricane fault, an amount that decreases to 210 feet at a distance 2.8 miles south at the small pinnacle on the skyline behind us. This pinnacle is capped with Peach Springs Tuff and is on the upthrown side of the Hurricane fault (the fault passes 0.2 miles to the west). Its displacement increases northward down Peach Springs Canyon.

5.7 Lower sandstone and transition member of Peach Springs Canyon; elevation 1,500 feet. The road here follows the Hurricane fault, and we will have a good view down Peach Springs Canyon. The transition member of the Rampart Cave is exposed in the road cut on right. That part of the Bright Angel on the left (west) side of road is stratigraphically between the Flour Sack member and the redbrown sandstone member of the Bright Angel. Nearly 400 feet are offset here on the hurricane fault, an amount that decreases to 210 feet at a distance 2.8 miles south at the small pinnacle on the skyline behind us. This pinnacle is capped with Peach Springs Tuff and is on the upthrown side of the Hurricane fault (the fault passes 0.2 miles to the west). Its displacement increases northward down Peach Springs Canyon.

6.9 Buck and Doe Conglomerate and Hindu Fanglomerate at 9 o'clock (left) fill a small tributary paleovalley that enters from the west. Many small caves in the recess in the lower cliff on right mark the erosional boundary between the Flour Sack Member of the Bright Angel Shale and the Rampart Cave Member of the Muav Limestone.

8.6 Water tank is on right; elevation 3,300 feet. Low ridge on right (east) is capped with the Buck and Doe Conglomerate above a resistant ledge of the redbrown sandstone member of the Bright Angel Shale. Unnamed drainage entering from the east is a rejuvenated canyon of the Tertiary meander canyon seen at last stop. This meander crosses the road here and swings north along the west wall of Peach Springs Canyon, then cuts back to the northeast a mile down the road, thus widening Peach Springs Canyon. The road follows very closely the covered Hurricane fault. Thick gravels of early Tertiary age in upper Peach Springs Wash are easily eroded and have choked the present drainage, causing the stream to erode in a braided pattern.

9.6 Bright Angel Shale. That part of the Bright Angel exposed here is stratigraphically below the Flour Sack Member and above the redbrown sandstone member. The Tincanebits Member of the Bright Angel is absent in this area, but two very thin limestone beds of the Meriwitza Member are present. Low saddle in the ridge at 2 o'clock marks position of the Tertiary drainage as it progressed northeast to form the next meander. The Hurricane fault splits here into several small splinter faults that are mostly covered by landslide debris.

10.0 Large slump blocks can be seen at 9 o'clock and 2 o'clock. The slump blocks in Peach Springs Canyon are typical of many in western Grand Canyon; the toes of almost all rest on the Bright Angel.

Figure 4. Unconformity within the Rampart Cave Member of the Muav Limestone, lower Peach Springs in Peach Springs Wash. View is northwest. Bar = Rampart Cave Member, Ebaf = Flour Sack Member of Bright Angel Shale.
Shale. Overlying Paleozoic rocks have rotated and dip against the parent wall as the giant slumps gradually slide into the canyon.

11.2 Mesquite Spring and campground; elevation about 2,960 feet. The spring is a small seep that supports a little forest of mesquite trees. A cliff of Tapeats Sandstone is behind the spring.

11.5 Precambrian metasedimentary and metavolcanic rocks are exposed at 3 o'clock on east side of the Hurricane fault.

12.3 Stop #2. Elevation is 2,840 feet. We have an excellent view up and down Peach Springs Canyon that parallels the strike of the Hurricane fault. We are about 75 feet west of the fault on the downthrown side; nearly 860 feet of displacement places Precambrian rock against the redbrown sandstone member of the Bright Angel Shale. The green shale just above the redbrown member contains trilobite trackways, worm tubes, and ripple marks. Cambrian rocks here dip west, away from the fault, because of fault drag.

Hells Canyon enters Peach Springs Canyon from the southeast at 4 o'clock. This dry canyon is a reentrant of an old Tertiary meander that has been rejuvenated for about 2 miles by headward erosion of Peach Springs Canyon. The old paleovalley turns north from here, but all traces of its existence to the north are missing because of younger Grand Canyon erosion. The northernmost possible remnant of this old valley, probably a partial meander as the old channel continued northward, is a slight indentation in the east wall of Peach Springs Canyon about 1 mile downstream at 2 o'clock and 1,200 feet up. Now the indentation is filled with landslide debris.

Lower Lost Man Canyon to our left (west) has also been rejuvenated for nearly 1 mile; it ends abruptly in sheer walls nearly 1,200 feet above the valley floor. Farther west, upper Lost Man Canyon is a wide relict of a major Tertiary paleodrainage from the southwest, now nearly filled with Tertiary sediments; the easternmost has the greatest displacement and the westernmost is about 130 feet west of the road.

13.0 Two thin gray ledges of Muav Limestone are seen in the Bright Angel slope above the redbrown sandstone member at 9 o'clock. The lower ledge is the Tincanebits Member and the upper ledge is the Marilvmites Member. We will see the redbrown sandstone member at road level increase in thickness and become redder as we go down the canyon.

13.7 Road is on the Hurricane fault at this bend.

14.3 Large slump block at 10 o'clock.

14.6 View of Diamond Peak directly ahead.

15.9 The road here is on a block of Tapeats Sandstone. The Hurricane fault has split into three segments; the easternmost has the greatest displacement and the westernmost is about 130 feet west of the road.

16.3 Stop #3. Tapeats Sandstone on both sides of the road. A short walk across the wash will provide a close-up view of the Great Unconformity between the Precambrian Vishnu Group and the Cambrian Tapeats Sandstone, a break representing about 1,100 million years. Stratigraphy of the west wall of Peach Springs Canyon at this stop is shown in Figure 5.

17.0 Travertine in roadcut on right.

17.4 Tapeats Sandstone on both sides of wash.

Figure 5. The west wall of Peach Springs Canyon at stop 3, looking southwest. Mr = Redwall Limestone; Dtb = Temple Butte Formation; Gu = unclassified dolomites; Gm = Muav Limestone (h = Havasu Member, g = Gateway Canyon Member, k = Kanab Canyon Member, p = Peach Springs Member, s = unnamed shale, sp = Spencer Canyon Member, sa = Sanup Plateau Member, r = Rampart Cave Member); Gba = Bright Angel Shale; St = Tapeats Sandstone.

17.9 Precambrian rocks both sides of wash.

17.2 Seeps on the left at 9 o'clock in the Tapeats Sandstone are depositing salt and calcium carbonate to form travertine. The travertine has cemented cobbles and pebbles derived from Peach Springs Wash at a time when the wash was a few tens of feet above its present position. A segment of the Hurricane fault cuts Precambrian rock on east side of drainage. The main fault continues down the wash.

18.7 The campground on the left, at the junction of Peach Springs and Diamond Creek canyons, is the site of the Farley or Diamond Creek Hotel, the first hotel in the Grand Canyon (Figure 6a,b). It was run by J.H. Farley between 1884 and 1889, until tourist accommodations were developed on the South Rim in what is now Grand Canyon Village in Grand Canyon National Park.

18.8 Diamond Creek enters Peach Springs Canyon from the right (east); elevation here is about 1,560 feet. The water in Diamond Creek comes from two springs about 9 miles upstream that emerge from the base of the Rampart Cave Member of the Muav. Diamond Peak is at 1 o'clock; the top is 1,950 feet above us. The Hurricane fault goes through the low notch capped with Tapeats Sandstone east of Diamond Peak (Figure 2). Displacement on the fault here is nearly 1,280 feet; its displacement in Grand Canyon is about 2,400 feet at a distance 10 miles farther north at Three Springs Canyon (Billingsley and others, 1986a).
19.2 Entrance to the narrows of lower Diamond Creek offers view of the Diamond Creek tonalite pluton on the west side of the Three Springs fault; Vishnu metavolcanic and metasedimentary rocks are on the east side of the fault.

20.1 Stop #4. Colorado River, elevation 1,343 feet. End of bus ride. Prepare for river trip and mass confusion in general. We are now in the Lower Granite Gorge of western Grand Canyon, Colorado River mile 225.6 (Mileage on the river is measured from Lees Ferry, Arizona, where river trips through Grand Canyon begin.)

Nearly a half mile upstream on the Colorado River is a remnant of a black basalt flow on the west bank, about 100 feet above the river. The basalt flow originated from Pleistocene volcanoes on the north rim of Grand Canyon about 35 miles upriver (Hamblin, 1970). This flow establishes a minimum age for the Colorado River, which is considered to be about 700,000 years old by Donald Elston (personal communication, 1986) or as much as 1.2 million years old by McKee and others (1968). We will see remnants of the basalt flow from here to river mile 254.2.

RIVER TRIP LOG

Conclusion of Day One

Mile 225.6 Diamond Creek Rapid is formed by boulders washed into the Colorado River from Diamond Creek. The rapid drops 25 feet in the next 0.4 mile. On July 19, 1984, a flash flood nearly 20 feet deep rushed through the narrows and deposited several tons of boulders into the rapids along with a 2-ton truck (with all the river equipment it contained), and a pickup truck. No one was hurt.

Mile 227.0 End of the Diamond Creek pluton, beginning of the 229-mile gneiss. Bottom of Lower Granite Gorge here is 1,115 feet below the top of the Tapeats Sandstone (Hamblin and Rigby, 1969). High on the right side, just above the Redwall Limestone, is a volcanic plug that fed a sill in the Watahomigi Formation. The sill was recently dated as 13.5 million years old (Paul E. Damon, personal communication, 1986).

Mile 228.2 Small north-trending fault crosses river; west side is down about 10 feet.

Mile 229.2 Stop #5. Camp at Travertine Grotto. Travertine Canyon enters from the south. Its water comes from a spring about 2 miles upcanyon in the Rampart Cave Member of the Muav. Most of the spring waters in western Grand Canyon are heavily laden with calcium carbonate that has resulted in heavy buildup of travertine, among the youngest deposits in Grand Canyon. Ages of large deposits of travertine elsewhere in Grand Canyon have been dated from 170,000 years to the present (Szabo and O'Malley, 1986), but the travertine here has undergone erosion for perhaps the last few thousand years or more.

Day Two

We will leave Travertine Grotto and continue down the Colorado River. Several rapids are still ahead. Estimated displacements on faults are in Paleozoic, not Precambrian rocks.

Mile 229.6 A small fault from the north crosses the river here and dies out a half mile south. At that point, a small reverse fault runs for a short distance to Travertine Spring, then becomes normal and trends southwest. Travertine Canyon is fault controlled; displacement is southwest side up and reaches a maximum of about 170 feet.

Mile 230. Beginning of the Travertine Falls pluton, which extends 0.9 mile downstream.

Mile 230.5 Travertine Falls on the south bank. The small spring above is from the Rampart Cave Member of the Muav and deposits the usual travertine. He will see several small springs and seeps in the next few miles. All on the south side of the river, because the regional dip (1°-2°) is toward the river from the southwest; subsurface water flows down dip, coming out where the Paleozoic strata are cut by the Grand Canyon. Almost all springs emerge from the Rampart Cave Member of the
Figure 8. Proposed Bridge Canyon Dam as envisaged by Reclamation project planners

A DAM IS PLANNED

Potential Bridge Canyon Dam as envisioned by the Bureau of Reclamation in 1946. (U.S. Department of the Interior, 1946, p. 169). View is to the northeast.

Muav and all deposit travertine. Only one spring is found on the north side of the Colorado. It is in upper Surprise Canyon, a northern tributary canyon farther downriver.

Mile 230.9 Beginning of 231-Mile Rapid and the Vishnu Group. Traverntine buildup on south wall is from a small active spring above the Tapeats Sandstone cliff.

Mile 232.0 232-Mile Rapid.

Mile 232.3 Beginning of highly contorted migmatitic gneisses of the 232-mile pluton that extends over the next 4.4 miles, the largest continuous exposure of migmatite in the Grand Canyon. These rocks are highly polished and fluted by the river for several feet above present water level and provide the best examples of river sculpture in Grand Canyon.

Mile 233.5 234-Mile Rapid; 234-Mile Canyon enters from the south and contains a small spring about a mile from the river.

Mile 233.6 Small northwest-trending fault crosses the river in about the middle of 234-Mile Rapid; west side is down about 10 feet.

Mile 233.8 Small northwest-trending fault crosses the river; east side is down about 15 feet.

Mile 235.2 Stop #6. Bridge Canyon Rapid. Bridge Canyon enters from the south. We will take a half-mile walk up-canyon to view the natural bridge. A small northwest-trending fault, east side down, crosses the river just above the rapid and ends a half mile up Bridge Canyon, where it passes just under the natural bridge. The bridge is composed of gravels that were cemented with travertine when Bridge Canyon was filled with gravel to the level of the bridge's top. Erosion has undermined the cemented gravels in the ravine, leaving the bridge behind.

Bridge Canyon is a small Colorado River tributary that has nearly captured the Tertiary paleovalley of Hindu Canyon to the south. The large supply of Tertiary gravels in the upper reaches of Bridge Canyon has clogged the relatively young canyon to heights represented by the natural bridge, and the canyon has now reeroded back down to its present depth. Water flows in the canyon, mostly under the thick gravel, but it can be seen where gneiss is exposed just upstream from the bridge.

The Bridge Canyon trail (that we will not follow) begins at the head of Bridge Canyon near Hindu Canyon in a narrow notch in the Redwall Limestone. The trail rapidly descends 1,800 feet of talus slope to the canyon floor. It continues westward along the Bright Angel-Tapeats contact to the proposed Bridge Canyon Dam site, a little more than 2 miles west from Bridge Canyon. From trailhead to the Colorado River via Bridge Canyon is a 4-mile walk and an elevation drop of 2,780 feet. The trail was constructed in the 1950’s to survey the proposed Bridge Canyon Dam site, but the dam has not been constructed because of environmental concerns and National Park protection on the north side of the river.

Mile 236.0 Gneiss Canyon Rapid. This rapid is the last one on the Colorado. All others downstream were drowned by Lake Mead. A small northeast-trending fault (west side down about 5 feet) crosses the river here. Gneiss Canyon enters from the south and a larger unnamed canyon enters from the north.
Mile 236.7 Beginning of the 237-mile granitic pluton, which extends 0.3 mile downstream.
Mile 237.5 Bridge Canyon Dam site, elevation 1,220 feet (Figure 8). The proposed dam would have been nearly 700 feet high, backing water as far as 81 miles to Havasu Canyon (U.S. Department of the Interior, 1946). Test tunnels are boarded-up on the north side in the wall of granite.
Mile 238.4 Prepare for a hike of about 3 1/2 miles and a climb of about 680 feet. Take water. Wild burros are still common on the south side of the Colorado River and we will follow their trails.

The remains of cabin foundations and other material left behind by workmen of the Bureau of Reclamation can be seen. We will climb up to the top of the Tapeats Sandstone and proceed northwest to south Separation Canyon. We will then walk down this canyon to the river, where the boats will meet us. The hike provides excellent views of the Lower Granite Gorge and the overlying 3,000 feet of Paleozoic rocks (Figures 9, 10). The upper 2,000 feet of Paleozoic rocks are eroded back for several miles to the north, and the skyline as viewed from the trail is formed of the middle of the Pennsylvanian part of the Supai Group. Separation Canyon, on both sides of the river, is a spectacular example of a fault-controlled canyon.

Figure 9. Aerial view looking north down south Separation Canyon to the Colorado River and up north Separation Canyon. The Colorado River crosses Separation Canyon from east to west in the gorge in about the middle of the picture.

Mile 239.6 Separation Canyon and fault. Meet boats here.

Separation Canyon is named for the separation of three men from Powell’s first expedition on the Colorado. Seneca Howland, O. G. Howland, and William H. Dunn, fearing disaster if the expedition continued on the river, hiked up north Separation Canyon on August 28, 1869. When they reached the Shiwits Plateau, they were mistaken for rowdy miners and killed by local Indians. Their memorial plaque is set in the bank of the Colorado River west of the south of north Separation Canyon.

The Separation fault controls north Separation Canyon for about 7 miles from the river and south Separation Canyon for about 4 miles from the river (Figure 9). It is a normal fault (west side up) having a maximum displacement of about 20 feet.
Mile 246.0  Spencer Canyon enters from the south. We will see a remnant of the Pleistocene basalt on the north bank. Before it was drowned by Lake Mead in 1935, this lava marked the beginning of Lava Cliff Rapid, considered the largest and roughest rapid in Grand Canyon. The stream flow in Spencer Canyon originates from various springs in Meriwhtica, Milkweed, and Hindu Canyons several miles to the south and southeast. All of these springs emerge from the Rampart Cave Member of the Muav, and produce travertine deposits from that stratigraphic level down to the Great Unconformity. (Permission to hike to these springs must be obtained from the Hualapai Tribe in Peach Springs.)

Mile 246.1  A small north-trending fault crosses the river; its east side is down about 5 feet. Beginning of the Surprise-Quartermaster granitic pluton.

Mile 247.0  Small north-trending fault crosses the Colorado and extends into side canyons on both sides of the river; its west side is down about 10 feet. Mile 248.3  Surprise Canyon, the largest tributary canyon of the Lower Granite Gorge, enters from the north. It is a 26-mile walk and nearly a 5,000-foot climb to the Shiwulits Plateau. About 12 miles up from the river, the canyon splits into Twin Springs and Green Spring Canyons. About 2 miles farther in Green Spring Canyon is Cottonwood Spring, the largest spring on the north side of the Colorado River in western Grand Canyon, flowing at an estimated rate of 120 gallons per minute. The spring emerges from the Mooney Falls Member of the Redwall Limestone and flows on the surface for a few miles before disappearing into the gravels of Surprise Canyon.

The type section for the newly named Surprise Canyon Formation (Billingsley and Beus, 1983) is about 14 miles northwest near the Bat Tower tramway. We will see the type section at mile 262.2. The formation (named for Surprise Canyon) is also exposed at five separate locations in the canyon between 7 and 13 miles from the river.

Mile 249.0  Clay Tank Canyon enters from the southwest (left). A little water flows near its mouth. A small remnant of the Pleistocene basalt flow is found in the protective reentrant. Mile 249.2  A small north-trending fault (west side down about 20 feet) crosses the river.

Mile 251.0  Very large slump blocks on both sides of the river resting on Bright Angel Shale.

Mile 251.6  Stop PB, south bank. Prepare for a 2-mile walk and climb of about 800 feet to the top of the Tapeats Sandstone. Once on top, we will have an excellent view north and south, along the trend of the Meriwhtica fault and monocline (Figure 12), of the Paleozoic stratigraphy and several large slump blocks. The canyon is 3,600 feet deep at this point.

Mile 252.2  Reference Point Creek enters from the southwest. The water flowing in the streambed gravels a mile upstream probably has its source in the Rampart Cave Member farther up the canyon. The Meriwhtica fault is a scissors fault that here has offset the Tapeats Sandstone about 290 feet down to the east. The fault plane dips to the west about 83° and thus this segment is a reverse fault. Beyond the hinge point of the fault 6 miles north, the fault becomes normal; farther north its maximum displacement is about 175 feet (Kenrich and others, 1986b). To the south, from Clay Tank Canyon to Milkweed Canyon, the fault is not exposed. Instead, Paleozoic rocks are bent over the fault, forming an east-dipping monocline with displacements of more than 1,000 feet (Huntoon, 1981; Billingsley and others, 1986b). The river follows the Meriwhtica fault for the next 3.3 miles.

Mile 254.2  On the north bank of the river, we will see the last remnant of the Pleistocene basalt flow, which has traveled at least 74 miles downstream the Colorado River. (The basalt actually may have gone a bit farther down the canyon.)

Mile 255.4  We will leave the Meriwhtica fault, which continues up Salt Creek to the north. Displacement here is about 185 feet, east side down. The Tapeats Sandstone is very close to the river on the right bank. A short distance up Salt Creek, the Tapeats is bent up sharply, dipping east along the fault, a good example of reverse-fault drag. Nearby, a small salt spring issues from the Tapeats. (All springs that emerge from the Tapeats Sandstone in Grand Canyon are salty.)

Mile 256.8  A fault (east side down about 100 feet) crosses the river. Jackson Canyon enters on the left (south).
Day Three

Mile 261.0 Tapeats Sandstone at lake level. (The Colorado River from here downstream is buried by Lake Mead sediments and flows only at low lake levels).

Mile 262.2 Directly west and high above the Redwall Limestone cliff is the type section of the Surprise Canyon Formation. Here the formation is thickest, most continuously exposed, and most accessible from the rim. (Look for reddish-brown rocks above the gray Redwall cliff.)

Mile 263.8 Tincanebits Canyon enters from the east. About 3 miles from the lake, this canyon makes a sharp bend to the north. Several Tertiary basalt dikes are found in the upper part of this canyon and in the next canyon to the north (Dry Canyon), which joins the river at mile 264.5.

Mile 265.0 The erosion of a small tributary canyon on the left (south) follows the trend of a Tertiary basalt dike. About 1,150 feet up the south wall in the cliffs of Muav Limestone, a conspicuous erosional surface can be seen between the Peach Springs and Kanab Canyon Members.

Mile 266.3 Bat Cave on the right about 700 feet up in the Kanab Canyon Member. The Bat Cave was discovered in the 1930’s and found to contain tons of bat guano, a valuable fertilizer. Merle Emery and Beal Masterson, the guano, but their barges sank in the river. They eventually sold the property to Kingman-Fiater, Co., who, in the late 1950’s, tried several methods of transporting the guano, including barge and aircraft. They built an air strip on a sand bar below the cave to fly the guano out to Kingman, Arizona, but a spring flood on the river eventually washed out the airstrip, ending that venture. King-Fiater sold the property to U.S. Guano Corporation of Calgary, Canada in 1958. A survey estimated that 100,000 tons of guano were present in the cave. U.S. Guano then spent nearly $3,500,000 constructing a tramway from the south rim to the cave. The span across the canyon was 7,500 feet and the vertical rise was more than 2,500 feet. The guano was sucked from the cave by a 10-inch vacuum hose and stored in a holding bin below the cave for the tram bucket. In one trip, the large bucket could haul simultaneously six men and 2,500 pounds of guano from the cave to the rim. The guano was trucked to Kingman and sold to markets on the West Coast starting in late 1959. During the height of operations, the bitter truth was realized. The cave did not contain 100,000 tons of guano as originally estimated, but only 1,000 tons. The rest of the deposit was decomposed limestone. The operation shut down in 1960 when all the guano was removed. Shortly afterward, a low-flying jet from Nellis Air Force Base was "hot-dogging it" down the canyon and hit the tram cable with its wing tip, severing the cable and sending it crashing into the canyon where it remains today. The jet made it back to Nellis with 6 to 8 inches of wing tip missing.

Miles 266.5 to 269.0 Several faults cross the lake and form a series of horsts and grabens on the Sanup (north) and Hualapai (south) Plateaus. Large deposits of travertine from springs in the Rampart Cave Member have accumulated on the north bank of the lake. The springs are now inactive and this travertine is not dated. On the south bank, several slump blocks have been partly reactivated since 1936 because the rising lake water has soaked their toes and caused them to slide farther. Fresh fault or slide scarps more than 30 feet high cut across some of the slumps.

Mile 272.9 Scorpion Spring on the south bank. Its water emerges from the Rampart Cave Member of the Muav and drips over the Meriwitica Member of the Bright Angel Shale, the next cliff below the slope. The Flour Sack, Meriwitica, and Tincanebits Members of the Bright Angel here are more typical of Muav lithology (and are, in fact, tongues of the Muav). For consistency in mapping throughout the western Grand Canyon, however, the base of the Muav is considered the base of the Rampart Cave Member. Mile 274.0 Weeping Spring on the south bank with travertine deposits. Water comes from the Rampart Cave Member above. Numerous caves in the Rampart Cave cliff on both sides of the lake in this area are filled with sandstone and travertine.

Mile 274.3 Cave Canyon drainage enters the lake from the south and forms Columbia Falls (originally Emery Falls). The water comes from a spring just above the Rampart Cave Member.

Mile 275.5 Rampart Cave is a half mile west of us at the top of the Muav member named for the cave. The Rampart Member is higher here than at Columbia Falls because of offset on the Rampart Cave fault. The fault begins near here and extends south about 30 miles into the Hualapai Plateau; its displacement (east side down) reaches a maximum of 600 feet. The cave contains Pleistocene giant-sloth dung deposits 11,000 to more then 40,000 years old (Long and others, 1974).

The large travertine bluff on our right (east) is the westernmost mass of travertine in the Grand Canyon. These deposits are from once-active springs that emerged from the Rampart Cave Member of the

Billingsley and others
Limestone. Progressively higher stratigraphic units of Mississippian rocks that dip northeastmost of the cliffs are in the Cambrian Muav dipping Paleozoic rocks (Figure 14). The part of the Paleozoic couplet (Kaibab and Toroweap over Hermit and Supai). The most prominent part of the upper Grand Wash Cliffs and the Grand Wash fault that formed them. In this segment, the upper and lower cliffs are not parallel, but diverge southward, probably as a result of increasing displacement along the fault in that direction and consequent greater erosional retreat of the upper cliffs (Figures 13, 14).

A structurally rotated block, once continuous with the upper Grand Wash Cliffs, and the Hualapai Plateau, can be seen along Wheeler Ridge, which is composed of east-tilted Paleozoic rocks (Figure 15). The reddish section and overlying gray section visible on Wheeler Ridge north of Lake Mead consist of the same upper Paleozoic rocks that form the upper cliffs (Supai and Hermit overlain by Toroweap and Kaibab). These rocks terminate abruptly just south of the lake, marking the location of the ancient scarp. Southward, Paleozoic rocks that are progressively lower stratigraphically are exposed along Wheeler Ridge almost to Meadview, where the featheredge of Paleozoic rocks in the Cambrian Bright Angel Shale and Tapeats Sandstone occurs.

The northeast dip of strata and the beveled surface on the Plateau, the upper Grand Wash Cliffs, and the featheredge of Paleozoic strata toward the south all reflect a belt of uplift of Laramide age (Mogollon Highlands) that existed southwest of the present Plateau margin before the onset of basin-range extension in Miocene time.

Structure

The Colorado Plateau block east of the Grand Wash Cliffs is structurally simple. Nearly horizontal Late Cretaceous-Miocene strata are cut by normal faults that have displacements measuring tens to several hundreds of feet and form horsts and grabens. Some of the faults are late Miocene. Others are as young as 7.5-6.5 Ma basalts (Lucchitta & Mckee, 1975); others can only be dated as post-Paleozoic. Going west towards the Grand Wash Cliffs, many of the faults are systematically up to the west, a sense of movement contrary to that of the Grand Wash fault (Lucchitta, in Goetz and others, 1975). An excellent example of such fault movement occurs in the westernmost Plateau block, which forms a conspicuous butte about 1 km north of the mouth of the Canyon. The persistent structural upwarping of the western edge of the Plateau may reflect a preexisting topographic uplift resulting in the Miocene to approaching a half dozen Miocene fault blocks.

Setting and Morphology

This viewpoint is an excellent stop for examining the regional aspects of three geologic subjects of considerable interest, namely (a) the Colorado Plateau Basin and Range structural transition; (b) the stratigraphy of a classic interior-basin deposit; and (c) the history of the Colorado River. Most of the information presented here is from Lucchitta 1966, 1967, and 1975.

AIRPORT POINT

The viewpoint is at the north end of Grapevine Mesa, near a WWII emergency landing strip. To the north is the Grand Wash trough, about 45 miles (70 km) long and rimmed to the east by the Grand Wash Cliffs, which are the western edge of the Colorado Plateau, and to the west by the south Virgin Mountains (Figure 13). The trough is closed to the north by the Virgin Mountains (8,064 feet, 2,460 m max.) and interrupted in the near and middle distance by Wheeler Ridge, which trends north-northeast.

Upper Lake Mead is conspicuous north of the viewpoint. Pierce Ferry, a popular put-in spot for fishermen and take-out point for river runners, is visible 2.5 miles (4 km) from and 1,700 feet (525 m) below the viewpoint. At its eastern end, Lake Mead enters the mouth of the Grand Canyon, although it is poorly visible from this angle.

South of the Grand Canyon, the Grand Wash Cliffs consist of a single step 2300 to 4000 feet (700 to 1200 m) high, cut chiefly in Cambrian, Devonian, and Mississippian rocks that dip northeast 2° to 4°. The caprock is the Mississippian Redwall Limestone, but most of the cliffs are in the Cambrian Muav Limestone. Progressively higher stratigraphic units are exposed along the cliffs toward the north. Owing to the divergence between the trend of the cliffs and the dip of the beds, the Hualapai Plateau, which is between the Grand Wash Cliffs and the Grand Canyon, is similarly underlain by progressively younger rocks toward the northeast. This is because an erosional surface that slopes northeast bevels the more steeply dipping Paleozoic rocks (Figure 14).

North of the Grand Canyon, the Grand Wash Cliffs consist of two steps—the lower and upper Grand Wash Cliffs, respectively. The upper Grand Wash Cliffs are an erosional scarp formed by the hard-over-soft upper Paleozoic couplet (Kaibab and Toroweap over Hermit and Supai). The most prominent part of the upper Grand Wash Cliffs trends northwest, parallel to the Grand Canyon, and was formed by scarp retreat northeastward down the structural slope (Figures 13, 14). The part best visible from the viewpoint, however, trends north-northeast, nearly parallel to the lower Grand Wash Cliffs and the Grand Wash fault that formed them. In this segment, the upper and lower cliffs are not parallel, but diverge southward, probably as a result of increasing displacement along the fault in that direction and consequent greater erosional retreat of the upper cliffs (Figures 13, 14).
junction that the Grand Wash fault has had its most recent movement. To the south, Wheeler fault again merges with the Grand Wash fault near the south end of Grapevine Mesa.

Iceberg fault crops out about 1.9 miles (3 km) west of Wheeler fault and follows Iceberg Canyon. Where now exposed, Iceberg fault dips 45° west; along most of its length, it is submerged by Lake Mead. However, according to Longwell, who mapped the area before the filling of Lake Mead (1936), the fault is concave upward and flattens out markedly to the west. The geometry is that of a listric fault along which the rocks of the Iceberg and south Virgin Mountain blocks have rotated to face steeply eastward. Similar relations are present in the block between Iceberg and Wheeler faults (which presumably has rotated along Wheeler fault) and on Wheeler Ridge (which is the exposed part of the block that has rotated along Grand Wash fault). The overall arrangement of rotated blocks strongly resembles the structure in Anderson's (1971) thin-skin tectonics, which in turn resembles the structure typical of the upper plates of core complexes. If these interpretations are correct, terrane of core-complex type would be present within a few kilometers of relatively undisturbed Colorado Plateau, an observation of much importance in attempting to understand the relation between the extended and the stable terranes. Another consequence of the interpretation is that the south Virgin Mountain block may have rotated together with the Paleozoic strata along Iceberg Canyon. If this is the case, the western end of the block would expose rocks originally 9 to 11 miles (14 to 17 km) deep in the crust.
The Muddy Creek Formation

The Muddy Creek Formation is a classic interior-basin deposit that is well exposed in three dimensions in the Pierce Ferry area because the Colorado River has cut to a depth of about 2,000 feet (600 m) beneath the original basin-fill surface. This dissection has revealed relations between facies in the vertical plane without obscuring them in the horizontal one.

The Muddy Creek was deposited in a basin formed by movement of the Grand Wash fault, which occurred after deposition of the 17-18 Ma Peach Springs Tuff and basalts on the Grand Wash Cliffs, and also of the 12-20 Ma Horse Spring Formation. The Muddy Creek includes basalts dated at 5 to 6 Ma (Anderson, 1978; Damon and others, 1978) and 10.9 Ma (Blair, 1978); it also includes tuffs about 8 Ma old (Blair, 1978; Bohannon, oral communication, 1982; 1984). Deposition ceased with establishment of through-flowing drainage at about 5 to 6 Ma. The formation, therefore, is middle to latest Miocene in age.

Paleogeography

The Muddy Creek Formation was deposited in an asymmetrical basin formed by movement on the Grand Wash fault. The axis of the basin trended north-northeast, parallel to the fault, and was near the eastern margin of the basin. The floor of the basin sloped gently from the north as well as from the south toward a low point located approximately where Pierce Ferry and the north edge of Grapevine Mesa are now. The surface of Grapevine Mesa is near the original stratigraphic top of the basin fill in this low spot. Filling of the trough was dominated from the west, as indicated by the predominance of igneous and metamorphic debris that includes clasts of the coarsely porphyritic rapakivi Gold Butte Granite of Longwell (1936). This granite crops out extensively in the south Virgin Mountains 4 to 12 miles (6 to 9 km) west of Wheeler Ridge (Volborth, 1962). Thirty-foot (9-m) boulders of the granite are present at Wheeler Ridge, and 20-foot (6-m) boulders are present at the foot of the lower Grand Wash Cliffs, a minimum transport distance of about 13 miles (20 km) from the nearest possible source area. Two prominent Muddy Creek fan lobes are visible to the north from Grapevine Mesa: one, the Pierce lobe, is south of the Colorado River and west of Pierce Ferry; the other, Tassi lobe, is north of the Colorado. Both lobes are east of Wheeler Ridge. During development of the fans, the country west of Wheeler Ridge consisted of a pediment cut on bedrock that functioned as a zone of transport rather than as one of deposition. This ended late in Muddy Creek time when movement on the Wheeler fault isolated the fans in the Pierce Ferry area from their source areas and created a separate basin of deposition in what is now the Greggs Basin area. Influx of material into the Grand Wash trough from the Grand Wash Cliffs to the east was minimal, partly because streams on the cliff face drain small areas, and partly because the carbonate rocks that form most of the cliffs erode slowly under arid conditions. Substantial fan lobes are present only at the foot of the Grand Wash Cliffs at Pierce, Snap, and Pigeon Canyons (Figure 13). The canyons that existed at these localities in Muddy Creek time were as deep and narrow as the present ones, but shorter and steeper.
All the fans are composed of poorly sorted breccia and conglomerate that contain chiefly angular to subangular clasts in pell-mell arrangement and with a high matrix-to-clast ratio. The tops of many depositional units show evidence of reworking by flowing water. These features, which suggest deposition by debris flows, can be studied conveniently in a prominent outcrop on the east side of the Pierce Ferry Road about 2.5 miles (4 km) north of the Sand Cove turnout. The areas between fan lobes are underlain by a fine-grained facies composed of freshwater limestone and dolomite, and gypsum. Silicic air-fall tuffs are common, many with delicate glass shards and bubbles still preserved. Twenty-three individual tuff layers are exposed about 1.5 miles (2.5 km) southeast of Pierce Ferry. Transition from the fan material to the fine-grained facies occurs partly through a progressive decrease in grain size, partly through abrupt interfingering. Rocks of the fine-grained facies are well bedded and well sorted. Deposition in quiet water is indicated by even bedding and the presence of tuff, gypsum, and carbonate. Many of these features are exposed near Pierce Ferry.

The fine-grained facies were deposited in intermittent playas and lakes. Initially, these playas and lakes were small and restricted to the lowest parts of the basin because the influx of clastics was high. As the basin filled and relief waned, playas and lakes occupied progressively greater areas. Finally, a lake occupied much of the basin in the lowest parts of the Pierce Ferry area because the influx of clastics was high. As the basin filled and relief waned, playas and lakes occupied progressively greater areas. Finally, a lake occupied much of the basin in the Pierce Ferry area. The Hualapai Limestone was deposited in the lake and eventually transgressed widely over other lithologies. These relations are well displayed on the north face of Grapevine Mesa (Figure 16). The transgressive stacking of limestone, fine-grained facies, and fanglomerate can be observed in road cuts and natural exposures on the east side of the Pierce Ferry road where it starts dropping off the north end of Grapevine Mesa, just north of the turnover to the viewpoint. The Hualapai Limestone therefore is not a sheetlike deposit that was laid down in a restricted interval at the end of Muddy Creek time, even though limestone deposition was most widespread at that time. Instead, it was deposited throughout the time represented by the 2000 feet (600 m) of Muddy Creek section exposed in the Pierce Ferry area as shown by the walls of Grapevine Canyon (cut into Grapevine Mesa), which are composed chiefly of Hualapai Limestone. At any time represented by the Pierce Ferry section, conglomerate, sandstone/siltstone, and limestone were being deposited simultaneously, but in different parts of the basin.

Inception of the Colorado River and Development of the Western Grand Canyon

Hypotheses

The Pierce Ferry area lies at the mouth of the Grand Canyon. Consequently, it provides important restrictions on interpretations of the history of the Colorado River and the development of the Grand Canyon. Longwell (1936, 1946) and Blackwelder (1934) long ago indicated that the present course of interior-basin deposits across the mouth of the Grand Canyon is the Grand Canyon. Consequently, it provides important restrictions on interpretations of the history of the Colorado River and the development of the Grand Canyon. Longwell (1936, 1946) and Blackwelder (1934) long ago indicated that the presence of interior-basin deposits across the mouth of the Grand Canyon in the Pierce Ferry area effectively precludes the existence of the Colorado River in its present course in Muddy Creek time. Others, however, do not agree with this view. Contrary hypotheses fall into two main groups:

1. The Grand Canyon existed in Muddy Creek time. The Colorado River emptied into the Grand Wash trough and ultimately exited by a course different from the present one (Lovejoy, 1980). Or: the Grand Canyon was a dry valley carrying neither water nor sediment (D. Elston, oral communication, 1983).

2. The Colorado River existed in Muddy Creek time, but its course near the southwestern edge of the Colorado Plateau was different from the present one. Eventually, the river was ponded near the western margin of the Plateau, whence it drained by subterranean piping to form springs and a lake in the
Grand Wash area. The Hualapai Limestone was deposited in this lake. The river eventually established its course in post-Muddy Creek time (Hunt, 1969).

Information Bearing on Hypotheses of Group 1.

There are no known deposits or structures in the Muddy Creek Formation that point to the existence of the Colorado River during Muddy Creek time. On the contrary, internal fabrics, facies distribution, as well as facies composition and relations, all point to a closed basin of interior deposition. It is unlikely that Colorado River sediments were deposited and then dispersed by wave action or currents, given the abundant evidence for quiet-water deposition in the Muddy Creek Formation. Even the oldest preserved gravels of the Colorado River, some at altitudes of several hundred feet above present grade, are immediately recognizable because of their coarse size (pebble-and-cobble gravel), excellent rounding, and exotic, far-traveled lithologies.

The facies distribution of the Muddy Creek Formation shows no evidence for an inlet or outlet, as would be required had the Colorado River existed. Specifically, such facies do not occur either at the mouth of the Grand Canyon, or at the south end of Grapevine Mesa, which is Lovejoy's (1980) postulated outlet. The latter locality is at an altitude of about 4,100 feet (1,250 m), or 1,150 feet (350 m) higher than the top of the basin fill near Pierce Ferry, and is underlain exclusively by locally derived fanglomerate and bedrock.

There is little reason to suppose that a drainage system nearby as large as the present Colorado River would have carried no sediment when drainage systems such as Pierce, Snap, and Pigeon Canyons, which are far smaller and only a few miles from the Grand Canyon, carried sediment in abundance. Nor can the absence of river sediments be attributed to a reduced base flow. The western United States is characterized by drainage systems (washes) that are dry most of the time, yet transport sediment in great quantities during rare floods. The amount of material transported by a drainage system is not closely related to its base flow.

Finally, a remnant of the Muddy Creek fan issuing from Pierce Canyon is present directly south of the mouth of the Grand Canyon. This material could not have been deposited in that position had the Grand Canyon existed at the time.

Information Bearing on Hypotheses of Group 2.

The evidence provided by the Muddy Creek Formation for conditions of interior drainage rather than through-flowing drainage near the mouth of the Grand Canyon cannot be bypassed by postulating another course for the Colorado River because locally derived interior-basin deposits of Miocene age are ubiquitous in the lower Colorado River region. Older deposits that predate formation of the basin of interior drainage indicate regional northeasterly drainage onto the Colorado Plateau instead of a westerly or southwesterly Colorado River drainage (see, for example, the distribution of the Peach Springs Tuff and the Rim gravels).

There is no evidence for springs feeding Hualapai Lake in the Grand Wash trough: the Hualapai Limestone was deposited during much of Muddy Creek time in scattered topographically low spots, which were frequently subject to playa conditions. Only at the end of the Muddy Creek time was limestone deposition widespread, when it occurred not only in the Grand Wash trough, but also tens of miles away and perhaps in separate basins.

Preferred Hypothesis

(Summarized from Lucchitta, 1966, 1972, 1979, and Mc Kee and others, 1967)

The Muddy Creek Formation of the Grand Wash trough shows that no Colorado River existed at the mouth of the Grand Canyon when the Muddy Creek was being laid down in middle to latest Miocene time. The youngest radiometric dates obtained from rocks correlative with the Muddy Creek Formation at the Grand Wash trough range from 5 to 8 Ma. The oldest date on rocks that reflect the existence of the Colorado River is 5.3 Ma, obtained from the Bouse Formation (Damon and others, 1978), an estuarine deposit that crops out widely along the lower Colorado River. The conclusion is that the lower Colorado River came into being after the end of Muddy Creek deposition and after the opening of the Gulf of California in latest Miocene time. The probable date for establishment of the lower Colorado River is about 5.5 Ma. The lower Colorado worked its way onto the Colorado Plateau by headward erosion and captured an older ancestral upper Colorado River, probably in the stretch between the Kaibab Plateau and the mouth of the Grand Canyon. In the process, the Canyon as we know today was formed. The western Grand Canyon was cut in 4 Ma at most, a remarkably short time.

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INTRODUCTION

The purpose of this field trip is to examine the structure, stratigraphy, and geomorphology of the eastern portion of Lake Mead and the lower portion of the Grand Canyon. This area straddles the boundary between the Colorado Plateau and the Basin and Range province, and Lake Mead provides a convenient access to this interesting transition. Because the western portion of the Grand Canyon is not normally visited by commercial raft trips, few geologists have had an opportunity to see it. This trip was conceived primarily as an introduction to this spectacular and geologically engaging region.

The trip originates in Las Vegas, Nevada with participants bused to Pierce Ferry, Arizona. At Pierce Ferry (Figure 1), participants will load into boats and journey up the Colorado River to Separation Canyon. From this point, we will slowly move downriver, discussing the structural evolution and stratigraphy of the Precambrian and Paleozoic section of the lower Grand Canyon. A number of stops and one long hike to Meriwitica monocline are planned. We will end day one on Lake Mead at Scorpion Island, near the entrance of Grand Canyon. As we leave the canyon, we cross the Grand Wash fault, the major structural element in the area. At this point we will be in the Basin and Range Province.

Day two will focus on the major structural features, stratigraphy, and geomorphology of the Transition Zone between the Colorado Plateau and the Basin and Range. Precambrian, Paleozoic, Tertiary, and Quaternary rock will be examined. We will look in some detail at the Wheeler fault on the north side of Lake Mead, and also at facies changes in the Paleozoic. A number of stops and short hikes are planned. The day ends at our campsite near Sandy Point in Gregg basin.

On day three we will continue to boat across Lake Mead, devoting our attention to Tertiary structures, sediments and geomorphic features of the southern portion of Gregg basin and Temple basin. Between Gregg and Temple basins we travel through Virgin Canyon, which cuts through Precambrian strata. The boat trip terminates early in the afternoon at Temple Bar, where we will be picked up for transport back to Las Vegas.

SUMMARY OF PALEOZOIC STRATIGRAPHY

The Lake Mead-western Grand Canyon region, which now lies athwart two very different geomorphic provinces (Colorado Plateau and Basin and Range), was covered by an epicontinental sea during most of the Paleozoic. North America was in an equatorial position oriented approximately 90 degrees clockwise from its present orientation. Warm, shallow seas repeatedly advanced southeastward (in terms of present...
coordinates) into northern Arizona from the Death Valley-Las Vegas direction, and then retreated again.

The Paleozoic section, therefore, is one of shallow-water carbonates punctuated by two siliciclastic intervals. The first siliciclastic interval is a transgressive one in the Cambrian, and the second is a regressive one in the Pennsylvanian and Permian. Because the Paleozoic sediments in the Grand Canyon-Lake Mead region lie directly on Precambrian basement (except, of course, for the Precambrian Unkar and Chuar Groups), this is referred to as a cratonic section. The westernmost cratonic section in the area is at Frenchman Mountain near Las Vegas. As shown in Figure 2, the cratonic Paleozoic section more than doubles in thickness between the eastern Grand Canyon and Frenchman Mountain. West of Frenchman Mountain the lower Paleozoic section lies conformably upon young Precambrian rocks and thickens dramatically. This section represents the rapidly subsiding miogeocline. Below is a brief summary of Paleozoic stratigraphy in the Lake Mead-Grand Canyon region, with an emphasis on thickness and facies trends between the Colorado Plateau and the Basin and Range provinces. For a more detailed discussion of the Grand Canyon Paleozoic section see McKee (1976) and Billingsley (1978). For Frenchman Mountain see Rowland (1987).

Figure 2. Cross section of upper Precambrian and lower Paleozoic rocks in the southern Great Basin and western Colorado Plateau (after Stewart and Suczek, 1977).

Cambrían

The Cambrian Tonto Group (Tapeats Sandstone, Bright Angel Shale, and Muav Limestone formations) records the Sauk transgression. Figure 3, from the classic study of McKee and Rosser (1945), shows the Cambrian stratigraphic nomenclature of the region. A later photograph will identify the various members of the Bright Angel and Muav as seen from river level in the Grand Canyon. We will have ample opportunities to examine portions of the Tonto Group on this field trip. Cambrian stratigraphy in the Grand Canyon is further discussed in the river log portion of this field guide. The Muav Limestone and overlying unclassified dolomites account for a large portion of the westward thickening of the cratonic section. The members of the Muav generally lose their distinctiveness westward, and the section becomes predominantly dolomite. At Frenchman Mountain these Cambrian carbonates have thickened to more than 2000 feet of dolomite and are usually referred to as the Bonanza King Formation (Figure 4) of the miogeocline.

Ordovician and Silurian

There are no known Ordovician or Silurian strata in the cratonic section. Although Figure 2 indicates the...
Figure 3. Diagrammatic section of Cambrian deposits in the Grand Canyon (from McKee and Reeser, 1945).

Figure 4. Stratigraphic columns of Paleozoic rocks in northern Arizona and southern Nevada, taken from various sources.
presence of Lower Ordovician limestones at Frenchman Mountain, they are in fact not present there (Rowland, 1987). Ordovician and Silurian units do occur in the miogeoclinal sections to the west.

Devonian

In the western Grand Canyon and Lake Mead region, the light-gray Cambrian dolomites are overlain by distinctively dark Devonian dolomite. Although this unconformity represents a stratigraphic hiatus of at least 100 million years, it is difficult to put one's finger on the contact. Clearly, the early Paleozoic was an interval of quiet cratonal subsidence in this region.

The thin, unobtrusive Temple Butte Limestone of the central Grand Canyon grades westward into a dark cliff-former. Stratigraphic tradition has been to carry the name Temple Butte all the way to the Grand Wash Cliffs, where most Grand Canyon studies end. Basin and Range stratigraphers have evolved a separate nomenclature for most of the Paleozoic, and in the Lake Mead region this unit is usually referred to the Sultan Formation (Figure 4). The Sultan typically contains conspicuous silicified stromatoporoids, which are present in western Grand Canyon sections as well as at Frenchman Mountain.

Mississippian

Mississippian units maintain nearly constant thickness and lithologic characteristics throughout the Grand Canyon and into the southern Great Basin, except that they are sometimes altered to dolomite in the Lake Mead-Las Vegas area. The Redwall Limestone has four members (Figure 4), and the same four members are distinguishable as far west as the Spring Mountains west of Las Vegas. Owing to the provincial development of stratigraphic terminology, however, west of the Grand Wash Cliffs the Redwall Limestone equivalents are usually referred to the Monte Cristo Formation. In both the Redwall and the Monte Cristo, the second member from the bottom is a very distinctive cherty interval (Thunder Springs Member of the Redwall and Anchor Member of the Monte Cristo).

The Upper Mississippian Surprise Canyon Formation, which fills erosional valleys cut into the top of the Redwall, is represented in western cratonal sections by a nonresistant horizon of red beds directly overlying the Monte Cristo Formation. In miogeoclinal sections this interval is represented by the siliciclastic Indian Springs Formation, which is composed of terrigenous sediments eroded off the Antler orogenic belt of central Nevada.

Pennsylvanian

The Pennsylvanian System contains the most conspicuous facies changes seen in the Paleozoic of this region. The lower three formations of the Supai Group (Watahomigi, Manakacha, and Wescogame) grade from fine- to medium-grained siliciclastic facies in the eastern Grand Canyon to carbonate-dominated facies of the Calville Formation in eastern Lake Mead (Figure 4). We will have an opportunity to examine the Calville Formation in Iceberg Canyon.

The Upper Pennsylvanian Pakoon Formation, which first appears in the section in the western Grand Canyon, thickens westward into a dolomite and evaporite sequence at Frenchman Mountain (Figure 4).

Permian

The basal Permian Esplanade Sandstone is recognizable westward across the craton, but is known by the name Quantoweap Sandstone in western sections where it is underlain by the Pakoon. The Hermit Shale, which in spite of its name is predominantly a sandstone, thickens westward and becomes a prominent red-bench forming unit in southern Nevada. The Hermit is often called merely "Permian redbeds" in the southern Basin and Range. The Toroweap and Kaibab formations retain their basic characteristics from the Colorado Plateau into the southern Great Basin, except that the cliff-forming members (Kaibab Canyon Member of the Toroweap and Fossil Mountain Member of the Kaibab) become conspicuously cherty when they leave the plateau. The stratigraphic relationships between Permian units in northeastern Arizona, the Grand Canyon, and southern Nevada are shown in Figure 5. We will not see the Kaibab, Toroweap, or Hermit while we are in the Grand Canyon; however, we will see all three units at lake level as we pass the Wheeler Ridge fault block on Lake Mead. The three units also form the erosional scarp of the Upper Grand Wash Cliffs.

![Figure 5. Cross section from miogeocline (west) to craton of Permian rocks. Inset shows marine transgressions and regressions schematically. Triassic rocks rest unconformably across these formations (from Peterson and others, 1980).](attachment:image)

**ROAD LOG - DAY ONE**

Two stops will be made en route to Pierce Ferry. The mileage indicated is measured from the Highway 93 and Dolan Springs, Arizona turnoff.

**Stop 1**

**36.6 Meadview Overlook.** Pull over on left at the Ranger Station. Overview of Gregg Basin and Virgin Canyon (day three of boat trip). Across the lake and to the west and north are the South Virgin Mountains composed of Precambrian granites. The brown, rugged hills to the west and on this side of the lake are composed of Cambrian carbonate. The tilted carbonates are capped by relatively flat-lying Tertiary Muddy Creek gravels. We are standing on a Muddy Creek fanglomerate consisting of granite clasts derived from the South Virgin Mountains. Since the Colorado River now lies between the Virgin Mountains and the fanglomerate, the river must postdate deposition of this member of the Muddy Creek Formation.

**Stop 2**

**40.8 Airport Overlook.** Turn right to landing strip. Viewpoint is on Grapevine Mesa, which is capped by the...
Hualapai Limestone Member of the Muddy Creek Formation. To the north and in the foreground eroded coarse-grained facies of the Muddy Creek are visible. The eastern panorama is the Colorado Plateau marked by the Grand Wash fault-line scarp. In this area the fault-line scarp (Lower Grand Wash Cliff) is capped by the Redwall Limestone. At the entrance to the Grand Canyon (NE but obscured), the Bright Angel Shale occurs at water level. In the distance to the northeast the Upper Grand Wash Cliffs, capped by the Kaibab Limestone, are visible. To the north and trending southwest the east-dipping Paleozoic section of Wheeler Ridge is in view. The portion of Wheeler Ridge on the north side of Lake Mead is capped by Kaibab/Toroweap and the underlying bright red unit is Hermit Shale. Further to the west is the Callville Limestone. Wheeler Ridge is the easternmost tilted fault block of the Basin and Range Province (for a more complete description of the geology from this viewpoint, see Lucchitta and Young, 1986).

48.0 Pierce Ferry. Leave bus and load into boats. Pierce Ferry is now an uncommercialized boat-launching and camping area. Prior to construction of Hoover Dam, a ferry provided transportation across the Colorado River. This was the first crossing-point below Lee's Ferry, 280 miles up river (Hamblin and Rigby, 1969). From Pierce Ferry we will boat up river approximately 40 miles to Separation Canyon.

COLORADO RIVER LOG - DAY ONE

The boat trip from Pierce Ferry to Separation Canyon will be relatively rapid and with minimal geologic discussion. The river log begins at Separation Canyon, with indicated mileage representing distance from Lee's Ferry, Arizona. This is standard procedure for river runners. Geographically, we will be on Lake Mead throughout the water portion of the field trip. For this guidebook, however, everything "upriver" from the entrance to the Grand Canyon will be considered to be river. The problem of lake/river occurs because of dramatic and sometimes rapid changes in Lake Mead water level. In July 1983 Lake Mead water level reached a record elevation of 1225.83 feet. This would place Lake Mead at Bridge Canyon Rapid (mile 235.5), or 4 miles above Separation Canyon. In April 1956 the all-time low elevation of 1083.23 feet was recorded. At this time large banks of river silt were exposed along the sides of the Colorado River from about Separation Canyon well into Lake Mead proper. The lake itself would have extended to about Surprise Canyon (mile 248.5). From record low to record high, a 143-foot increase in lake level results in a 13-mile up-canyon expansion of Lake Mead.

Stop 3

239.5 Separation Canyon. The bedrock at water level is mapped as a Precambrian foliated granitic pluton (Huntoon and others, 1982). The prominent dark-brown cliff overlying the granite is the Tapeats Sandstone. The Great Unconformity, representing perhaps 1.2 billion years of erosion, occurs below the Tapeats. Similar to most tributary canyons in the Grand Canyon, Separation Canyon is fault controlled (Figure 6). The Separation fault, east down a few tens of feet at the Colorado River, is a rather minor northeast-trending normal fault remarkable more for its linearity than any other characteristic. The displacement along the fault attenuates upward in the Paleozoic section, which is typical of normal faults that propagate upward from the basement in the region. The upward attenuation of displacements readily explains why fault densities usually increase with depth of exposure on Grand Canyon geologic maps. The Separation fault is one of a series of late Tertiary normal faults resulting from general east-west extension across this part of the Colorado Plateau. Faults in this group began to develop in Miocene time and activity along them continues, although no recent deformation can be documented along the Separation fault.

![Aerial view of Separation Canyon (mile 239.5) showing the fault-controlled nature of the canyon. The Separation fault trends southwestward and continues into the canyon on the opposite side of the Colorado River (bottom, left). East side is down.](image)

It was at this location that Seneca Howland, O.G. Howland and William H. Dunn mutinied on August 28, 1859, during the famous Powell expedition through the Grand Canyon. Fearful of the rapid that lay ahead (now gone because of the influence of Lake Mead), they left the boat party and hiked northward up Separation Canyon eight miles to the rim of the Shivwits Plateau. Once on the Shivwits Plateau, they continued northward through the forests, starving and exhausted, until they were murdered by Indians. The Indians mistook them for renegade white miners from the north who had murdered a squaw during a drunken brawl days before. Their defense based on a voyage through the Grand Canyon on the Colorado River was simply too fantastic to explain their presence in this remote part of the world in 1869 (Powell, 1895, p. 280, 323).

From Separation Canyon downriver to mile 254, Precambrian basement occurs continuously at water level. The basement complex, as mapped by Huntoon and others (1982), consists of foliated and nonfoliated granitic plutons, foliated tonalite plutons, mica schist, mica schist and amphibolites, and paragneiss.

246.0 Lava Cliff Rapid and Spencer Canyon. Submerged under Lake Mead at this point is the most treacherous
rapid in the Grand Canyon. It, like all other Grand Canyon rapids, owed its origin to debris washed in from tributary canyons, in this case the extensive Spencer Canyon drainage. The basalt outcrop along the north bank is a remnant of an inter-canyon lava flow that probably entered the canyon upstream at mile 180. The site of the former village of the Hualapai Indians is found in Meriwitica Canyon, three miles upstream from the mouth of Spencer Canyon. The Hualapai lived in isolation in adobe-type dwellings on a 1/2-mile-wide alluvial terrace developed behind a 600-foot-high travertine dam below Meriwitica Spring. The spring, which discharges from the Rampart Cave Member of the Muav Limestone, supplied ample drinking and irrigation water. The flat alluviated canyon floor and spring combined to provide a setting similar to the currently occupied Havasu Canyon 51 miles to the northeast. A meteorite slammed into the north wall of Milkweed Canyon after dusk one evening in 1924 (Grant Tapije, personal recollection, 1986), leaving a scar on the face of the Redwall cliff above the village (Young, 1978, p. 288). The scar is apparently obscure now, although one photo of it is known (Richard Young, personal communication, 1986). The Indians took this terrifying event as a bad omen and fled the canyon the next day, never to return permanently. The Indians occasionally, though rarely, venture into the forbidden place and once attempted to establish a small farm in the valley after World War II (Richard Young, personal communication, 1986). Time has obliterated the village.

Stop 4

252.0-255.5 Maxson Canyon, Meriwitica Monocline and Fault. Weather permitting, we will stop at Maxson Canyon and undertake a 4-to 5-hour hike. The hike will be approximately 5 miles round trip with a vertical ascent of about 2000 feet. The purpose of the hike is to view the Meriwitica monocline. Following the hike we will continue downstream, where superb exposures of the Laramide Meriwitica monocline and underlying basement fault occur along the west side of the canyon. The Tapeats Sandstone bench is displaced 450 feet down to the east by the west-dipping, high-angle, reverse Meriwitica fault (Figure 7). This single fault is the basement fault underlying and coring the Meriwitica monocline visible along strike to the south. It appears to be a reactivated Precambrian normal fault identical to the documented reactivated Precambrian normal faults which core the Hurricane, West Kaibab, and East Kaibab monoclines to the east. The fault attenuates upward in the Paleozoic section, the monocline was restricted to the Laramide period, and folding has not been recurrent in subsequent time. The excellent outcrops exposing the underlying basement rocks along the Meriwitica monocline in Milkweed Canyon provide definitive evidence for horizontal compression as the causative mechanism for emplacement of Laramide monoclines in the Grand Canyon region. The west-trending segment of the Meriwitica monocline lying between Meriwitica and Milkweed Canyons links two high-angle, west-dipping, north-striking, reactivated Precambrian faults. The eastern end of this linking segment was not underlain by a preexisting Precambrian fault. Consequently the Laramide fault that developed under this segment caused deformation of reasonably isotropic basement rocks. The result was development of a 30-degree west-dipping thrust fault in the basement rocks during Laramide compression (Figure 8). The dip of this unusual fault, coupled with numerous small-scale conjugate thrust faults in both the Precambrian and Paleozoic sections, was used by Huntoon (1981) to deduce a horizontal orientation for the maximum principal stress tensor during monocline emplacement.

All early Tertiary sediments and younger volcanic and sedimentary rocks are undeformed over the Meriwitica monocline. Therefore, the emplacement of the monocline was restricted to the Laramide period, and folding has not been recurrent in subsequent time. The Meriwitica monocline has not been downfaulted to the west by late Tertiary extension. This is in

Bachhuber and others
contrast to many other Grand Canyon monoclines, which have been displaced by normal Tertiary motion along their coring Precambrian faults (Figure 8). Oddly the Meriwitica monoclone did not develop in the area north of Lake Mead, although the Precambrian basement fault appears to extend a considerable distance northward. However, the late Tertiary extension has caused down-to-the-west normal reactivation along the preexisting basement fault, resulting in displacement of the Paleozoic section in the region north of the lake. What is unusual about the Meriwitica structure is that east-down Laramide monoclinal deformation and west-down late Tertiary normal faulting are spatially separated along the controlling basement fault, rather than being superimposed as is typical of the principal Grand Canyon monoclines such as the Hurricane, Toroweap, and West Kaibab structures to the east.

A. LARAMIDE FOLDING OVER REACTivated PRECAMBRIAN FAULT; ORIGINAL FAULT WAS NORMAL.

B. LATE TERTIARY NORMAL FAULTING.

C. LATE TERTIARY CONFIGURATION AFTER CONTINUED EXTENSION.

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Figure 8. Diagrams showing typical Grand Canyon monoclines. A - Profile typical of the Meriwitica monoclone where Tertiary faulting has not taken place. B - Profile across the West Kaibab monoclone and fault where Tertiary extension has resulted in faulting of the monoclone. C - Profile representative of the Hurricane monoclone and fault zone where Tertiary extension has progressed to the point that basins are beginning to form along the structure.

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253.2 Lava Remnant. The furthest exposed downriver remnant of an inner-canyon lava flow occurs along the east side of the river. The flow is believed to have entered the canyon 7/4 miles upstream. If Lake Mead water level is high, we may not be able to see the outcrop.

In this area the Tapeats Sandstone occurs at water level along the east side as the river follows the trace of the Meriwitica fault. A Precambrian foliated tonalite pluton crops out along the western bank. About one mile ahead, the river turns west but the Meriwitica fault continues northward into Salt Creek Canyon.

255.5 Salt Creek Canyon. The Meriwitica fault continues northward as the river bends to the west. As we cross the fault trace at Salt Creek Canyon, Precambrian basement crops out at the same elevation along both sides of the river. From this point the Great Unconformity descends progressively towards water level until it disappears downriver from Quartermaster Canyon.

Stop 5

256.5 Quartermaster Canyon and Spring. A beautiful exposure of the Great Unconformity occurs across the river from Quartermaster Canyon. The microtopography of the Precambrian landscape is clearly evident. On the Quartermaster Canyon side of the river (south side), the Great Unconformity is 20 to 30 feet higher than on the north side. This may be due to the northward regional dip (Hamblin and Rigby, 1969), original Precambrian topography, or a minor fault.

A large travertine deposit crops out along the east side of Quartermaster Canyon. A similar deposit about 1/4 mile up canyon is the site of Quartermaster Spring, which discharges from the base of the travertine. The water originates from solution tubes developed in the Rampart Cave Member of the Muav Limestone, which subcrops in the canyon wall under the top of the travertine deposit 850 feet above the river. Most of the water flows through the travertine to the floor of Quartermaster Canyon; however, a small spring discharges on top of the travertine deposit during high-flow periods. Discharge from the Muav Limestone is localized in the topographically lowest outcrops of the Rampart Cave Member in the vicinity and has no other structural control. The water within the Rampart Cave Member apparently circulates through caves that are localized on joints that are favorably oriented parallel to the hydraulic gradient within the unit. The Rampart Cave Member of the Muav Limestone is the most productive aquifer in the region, and the water is perched within it above the underlying thick Bright Angel Shale member.

The temperature of the water issuing from Quartermaster Spring is 75 degrees F, a value very characteristic of numerous Muav springs in the area west of Kanab Canyon (Huntoon, 1977b, p. 26). In contrast, ground-water discharge from the Muav Limestone to the east of Kanab Canyon is commonly between 55 and 65 degrees F. The difference is herein attributed in part to higher geothermal gradients in the western Grand Canyon district resulting from thinned crust in the region west of the West Kaibab fault zone as revealed by geophysical evidence (Keller and others, 1975).

Downriver from Quartermaster Canyon numerous opportunities exist for viewing the Paleozoic section (Figure 9) of the Grand Canyon. The canyon rim is usually capped by the Redwall Limestone, but in places the lower members of the Supai Group can be seen. If lighting is good (and it won’t be), the Surprise Canyon
Consequently you can climb through a series of rocks, Shale, and Muav Limestone, on the basis of each having shales members of the Bright Angel Shale, and the section in the western Grand Canyon, all the formations mappable. As you view the Cambrian ascending formations, Tapeats Sandstone, Bright Angel intertonguing entities over scores of miles, a fact Cambrian formational nomenclature. Noble defined three and Muav formations. When viewed in plan, the repeats red-brown sandstones members of the Tapeats Sandstone. Noble's lithology-based nomenclature, McKee and Hesser for the superimposed sequence that he found along the classic when they described a common lithology, a perfectly appropriate framework impeccable detail. The work was flawed, however, by take on the appearance of thin imbricate thrust that alternate back and forth between the Bright Angel and Muav formations. When viewed in plan, the repeats take on the appearance of thin imbricate thrust sheets! This quirk in the nomenclature led to one serious series of water-well prospecting failures on the Hualapai Plateau documented by Huntoon (1977a). Huntoon and Billingsley were forced to arbitrarily use the base of the Rampart Cave Member of the Muav Limestone as their mappable Muav-Bright Angel contact on their three western Grand Canyon sheets. Unfortunately the Rampart Cave Member does not extend into the eastern Grand Canyon, so the base of the Peach Springs and Kanab Canyon members of the Muav Limestone had to be used for the contact on the eastern Grand Canyon sheet, producing a discontinuity between the sets of maps. Work is progressing on redefinition of the Cambrian stratigraphic nomenclature, an effort that will honor McKee and Resser's finding that the entire Cambrian section represents a common depositional environment, giving it legitimate group status. It could be internally subdivided into formations wherein the formations more closely honor the criterion of mappability.

264.0 Tincanebits Canyon. Tincanebits Canyon enters from the east (right). In this area the lower portion of the Bright Angel shale occurs at water level and exhibits many of the stratigraphic problems discussed in the previous section. Focus on the two orange-brown dolomitic units, the Heriwitica Tongue (upper) and the Tincanebits (spelling from McKee and Resser, 1945) Tongue, in the upper portion of the Bright Angel Shale. These tongues can be traced to the entrance of the Grand Canyon and to the Devil's Cove section (day two) of Lake Mead. The tongues thicken to the west and become the Lyndon Limestone at Frenchman Mountain (Figure 8), east of Las Vegas. The Tapeats Sandstone disappears below water level 1 to 2 miles upstream, depending on the level of Lake Mead.

265.9 Bat Cave. As we approach Bat Cave, note the towers on the southern skyline. The highest tower is on a lower member of the Supai Group. The towers were constructed in conjunction with a guano-mining operation. Following a few previous attempts, U.S. Guano Corporation began developing a guano mine in 1956 that produced bat guano from the prominent Muav Limestone cave 800 feet above the lake on the north wall of the canyon (Billingsley, 1974). The guano deposit was mistakenly estimated to contain 100,000 tons (New York Times, 1957). Actual production reached about 1,000 tons at a cost of approximately $3.5 million (George Billingsley, personal communication). The deposit, which met or exceeded 6 percent nitrate, extended 2,000 feet into the cave mined by manually loosening it with hoes, whereupon it was moved through a 10-inch vacuum hose to the loading bin at the entrance (Beatty, 1962). From there it was transported by tram via a cable tramway for a 9,820-foot-long, 3,000-foot climb to the tower visible on the south rim. Next it was trucked to Kingman for sacks and distribution. The operation ceased in the early 1960's as a result of depletion of reserves. In 1962, several months after the mine was closed, a wayward jet from Nellis Air Force Base collided with the tram cable, causing the cable to collapse into the canyon. The pilot managed to limp back to base with 6 to 8 inches of his wing tip missing. Local legend has it that a Federal damage payment to the mine operator finally placed the mining venture in the black.

265.9 Bat Extensional Fault Zone. Numerous north-trending normal faults trend through this area, producing a horst-graben complex in the adjacent plateaus. These faults are probably late Tertiary in age. Maximum displacements on individual faults attain 200 feet, but the displacements diminish with depth, indicating that the faults are propagating downward in
the section over a deforming substrate. As extension
is taking place, the Pennsylvanian and Mississippian
rocks capping the plateau are deforming mostly through
brittle failure, whereas the underlying Cambrian units
are accommodating the strain in part by ductile
deformation. Travertine deposits along the north side
of the river are not faulted, indicating that
extension has not taken place within the fault zone
during Quaternary time.

Stop 6

271.0 Garden of Eden. Informally named by river
runners, the Garden of Eden canyon enters from the
south (left) through the Bright Angel Shale. We will
take a 1/4-mile hike to a spring-fed, fern-covered
glen, typical of many of the standard tourist stops
along the Colorado River. As we hike to the spring,
note the Quaternary landslide deposit on the left
canyon wall. Also note the modern travertine deposits
at the spring.

273.0 Grand Pipe. The Grand Pipe is one of the finest
exposed and largest dissolution collapse structures
known in the Grand Canyon region. It occurs 1.5 miles
north of the river on the Esplanade surface but is out
of view from the river (Figure 11). The pipe is 0.5
mile in diameter and its center contains a few
hundred feet of infolded and infaul ted Hermit Shale,
which otherwise has been stripped from the surface in
the vicinity. The collapsed core of sediments in the
pipe is surrounded by ring faults and the deformed
sediments are altered (Hoffman, 1977, p. 12-21).

Breccia pipes and circular collapse structures are
very common in the Grand Canyon region and crop out at
all stratigraphic levels above the Redwall Limestone.
Some of the pipes contain economic uranium and copper
ores. For example, the Orphan Pipe near Grand Canyon
Village has produced hand samples containing 55
percent \( U_3O_8 \) (Gornitz and Kerr, 1970). More than
1000 breccia pipes and collapse features have been
discovered on the Hualapai Indian Reservation through
an exploration project led by Wenrich (1985). A few of
the Hualapai pipes appear to be economic. The pipes
are characterized by various combinations of the
following: (1) inwardly dipping country rock, (2) ring
fractures and faults, (3) downward displaced cores,
(4) breccias derived from overlying strata, (5)
bleached and limonite-stained rocks, and (6)
mineralized rock. Uranium mineralization is Triassic
in age (Ludwig and others, 1986), thus predating even
the Laramide monoclines in the region. Most of the
pipes stoped upward from Mississippian paleokarst
cavities in the upper Redwall Limestone. Space was
progressively created within the upward stoping pipes
through dissolution of carbonates as ground water
circulated through the pipes. Mineralizing solutions
appear to have circulated upward through the pipes as
the pipes pierced confining layers above the Redwall
proposed an alternate hypothesis involving downward
circulation of fluids within the pipes.

274.5 Emery (Columbine) Falls, Columbine Spring, and
the Rampart Fault Zone. Columbine Spring discharges
from the Rampart Cave Member of the Muav Limestone.
The spring is localized on extended joints in the
floor of Cave Canyon a half mile east of the Rampart
fault on the downdropped side. The Rampart fault has
as much as 600 feet of east-down displacement on the
Hualapai Plateau and extends southward more than
12 miles from the canyon rim. Dissolution-enhanced
fracture permeability in the lower Paleozoic
carbonates allows this fault zone to serve as a drain
for the northwestern part of the Hualapai Plateau.

275.0 Rampart Cave. The Rampart Cave Member of the
Muav Limestone is named for the cave that occurs in
the member in the alcove on the south side of the
canyon. The cave, at an elevation of 1750 feet or
Figure 11. The Grand Pipe, a solution collapse structure that is 1/2 mile in diameter. Infolded strata consist of the Hermit Shale. Surrounding rocks are the Esplanade Sandstone. View is toward the west across the edge of the Colorado Plateau.

about 550 feet above water level, contains a thick section of dung (more than 4 feet in areas) left by Pleistocene giant ground sloths (Nothrotherium shastense). Based on radiocarbon chronology, the ground sloths occupied the cave from 35,000 to 10,000 yr B.P. (Martin and others, 1961). The dung deposits were badly damaged by a fire in 1976 that burned for about 6 months. The fire was apparently started by explorers. The entrance to the cave is now fenced and locked.

278.0 Lower Grand Wash Cliffs. The Lower Grand Wash Cliffs mark the western edge of the Colorado Plateau and for this guidebook, the beginning of Lake Mead. Prior to construction of Hoover Dam, which was completed in the early 1930's, the channel of the Colorado River at this point had an elevation of about 900 feet. If the Lake Mead water level today is 1209 feet (projection made 1/87, Bureau of Reclamation, personal communication), 300 feet of additional stratigraphic section would be visible. This would place us in the Precambrian basement complex. Therefore, at the time of Powell's expedition and through the early 1930's, the Lower Granite Gorge (beginning at mile 216) extended all the way to the Grand Wash Cliffs.

Stop 7

279.0 Scorpion Island. End of day one and campsite. Scorpion Island consists of Quaternary-age Colorado River gravel and sand.

LAKE MEAD LOG - DAY TWO

279.0 Scorpion Island. Standing on the highest part of Scorpion Island, we have a 360 degree panorama of Colorado Plateau/Basin and Range geology. To the east the Lower Grand Wash Cliffs mark the edge of the Colorado Plateau. The Lower Grand Wash Cliffs compose a fault-line scarp that has retreated 2 to 3 miles eastward. The trace of the Grand Wash fault, the physiographic boundary of the Colorado Plateau, is buried here by the youngest beds of the late Miocene-early Pliocene Muddy Creek Formation. The displacement on the Grand Wash fault at the mouth of the Grand Canyon could exceed 15,000 feet (Lucchitta and Young, 1986). The northern exposed trace of the fault terminates in the Virgin Mountains of southern Utah, where displacement is only a few hundred feet, whereas the southern portion can be traced to west-central Arizona (Lucchitta and Young, 1986). The actual geometry of the Grand Wash fault has yet to be determined and a healthy debate on this trip is expected.

The age of the Grand Wash fault is somewhat less speculative. Certainly the fault predates the basin fill that surrounds us. Lucchitta (1967) presents evidence for early Miocene through Pliocene timing for most of the displacement along the fault north of the Grand Canyon. Moore (1972, p. 45) describes evidence for late Pliocene- Pleistocene(?) displacements along the fault near the Arizona-Utah State line. Most of the displacement along the fault occurred after the emplacement of the Peach Springs Tuff (17 Ma) but prior to deposition of the upper part of the Muddy Creek section in the area west of the Hualapai Plateau and south of the Colorado River. An additional 1,000 feet of posttuff offset has occurred in the Truxton Valley to the south (Young and Brennan, 1974, p. 86). The tectonic trough in which we are standing appears to be completely filled with Muddy Creek sediments near the mouth of the Grand Canyon and a thick salt section at Red Lake to the south (Peirce, 1972).
The Muddy Creek Formation, which surrounds us, can be subdivided into three main facies: (1) the upper Hualapai Limestone, (2) a siltstone-sandstone facies, and (3) a fanglomeritic facies. Lucchitta and Young (1986) subdivide the fanglomerate into sedimentary-clast-bearing and Gold-Butte-granite-bearing units. The stratigraphically higher siltstone-sandstone facies fills the central portion of the basin and was deposited when topography was reduced and tectonism decreased. Lucchitta and Young (1986) believe that the siltstone-sandstone facies was deposited under playa conditions with the playas initially being small but progressively expanding through time as the basin filled with sediment. During late Muddy Creek time the Pierce Ferry basin was flooded with the resultant deposition of the lacustrine Hualapai Limestone. The Hualapai Limestone caps Grapevine Mesa to the south of us, the location of Airport Overlook of Stop 2.

The Muddy Creek Formation contains numerous air-fall tuffs and lava flows. Flows dated by various workers indicate ages from 5 to 11 Ma (Anderson, 1978; Blair, 1976; Damon and others, 1978). Radiometric dating, therefore, suggests a Miocene age for the Muddy Creek with the uppermost portion possibly being Eocene.

To the west of us we can see Wheeler Ridge, the first tilted fault block of the basin and Range. The remainder of the field trip will be spent in the Basin and Range province.

283.0 Wheeler Ridge. At this point the Colorado River/Lake Mead bisects Wheeler Ridge. The exposed Paleozoic section is folded into grand east-dipping hogbacks on the upthrown side of the Wheeler fault, which is still out of view. The most striking view of Wheeler Ridge occurs on the north side of the lake. Here the ridge is capped by Kaibab/Toroweap with a thick section of the Hermit Shale forming the slope. The small scenic cove indenting the Hermit is God's Pocket. The bedded carbonates of the Callville Formation form the resistant upthrown side. The base of the Hermit consists of numerous normal faults, and repeats a number of times. In this small cove we have an excellent opportunity to view the Callville in all aspects.

Stop 8

285.0 Wheeler Fault. We will stop on the north shore of the lake and hike into the Wheeler fault zone. The hike will take approximately 3 hours. The Wheeler fault is a major down-to-the-west Basin and Range extensional structure. The fault has in excess of 4,000 feet of displacement and has not been active in this area since the youngest Muddy Creek beds were deposited over the fault. In the Gregg basin we will see the southwest-trending fault displace the Hualapai Limestones by up to 1000 feet (day three). The Wheeler fault joins the Grand Wash fault to the north and to the south.

288.5 Grand Wash Bay. The entrance to Grand Wash Bay trends to the north. We will boat a short distance into the bay to view the Muddy Creek, Miocene basalt flows and Quaternary gravels. A spectacular outcrop of tilted Paleozoic rock, flat-lying Muddy Creek, and capping lava flow occurs at the southwest corner of Grand Wash Bay.

288.0 Driftwood Cove. Now on the west side of Wheeler fault, the Paleozoic section is repeated. The red rock in the center of the cove is Hermit Shale capped by Kaibab and Toroweap. As we enter the cove we cross the trace of the Iceberg fault. The fault can be seen on the east side of the cove with the right block up. Turning 180 degrees we look down the trace of the fault into Iceberg Canyon.

288.1 Entrance to Iceberg Canyon. Except for the exposure in Driftwood Cove, the trace of the Iceberg fault (Figure 12) is covered by Lake Mead. Longwell (1936) mapped the fault prior to the development of the lake. He reports that the fault is concave upward and flattens out to the west, having a geometry similar to a listric fault (Lucchitta and Young, 1986).

The coockscomb bedding of the Callville Formation occurs on the right side (northwest) of Iceberg Canyon. On the left side we pass down section from the Callville to Cambrian carbonates. The left-up block across the channel exhibits numerous normal faults, repeated sections, and slump blocks.

Time permitting, we will take an excursion into a small canyon located near the entrance to Iceberg Canyon on the northwest side. This intriguing canyon enters the Calville perpendicular to strike, turns along strike, and repeats a number of times. In this canyon we have an excellent opportunity to view the Callville in all aspects.

289.5 North Howland Cove. North Howland Cove enters Iceberg Canyon on the left (southeast). As we enter the cove we pass through Cambrian carbonates. At the end of the cove we will see the Mississippian section and the easily identifiable Thunder Springs Member of the Redwall Limestone (Figure 4).

Stop 9

291.0 Entrance to Devil's Cove. Devil's Cove is the first large cove on the right as we leave Iceberg Canyon. The cove contains the complete, but faulted, Paleozoic section from the Callville down through the Tapeats Sandstone. The Tapeats is in contact with Precambrian granite. A short hike with a vertical ascent of several hundred feet is planned.

Stop 10

294.0 Sandy Point Area. End of day two and campsite. We will camp on Quaternary-age gravels and sands similar to our day-one campsite. Sandy Point is the prominent promontory to the south of us.

L A K E M E A D L O G - D A Y T H R E E

Stop 11

295.0 Sandy Point. After breaking camp we will travel the short distance to Sandy Point. The sand-and-gravel deposits of Sandy Point are protected by an interbedded lava flow. At the tip of the promontory a sequence of lower petrolastic gravel, basalt, and upper gravel and sand is visible. The upper gravel is a Quaternary deposit, but the age of the lower gravel is unknown. Most likely, however, it is of Muddy Creek age.

When we leave Sandy Point we will travel across the Gregg basin. Since the Miocene the Gregg basin (as a tectonic basin) has been geographically separated from the Pierce Ferry basin. The two tectonic troughs had separate but similar depositional histories. Fanglomerates poured into the basins following the formation of the Grand Wash and related faults. But as the basins filled, the topographic sill between the
basins was "breached." By upper Miocene, the Hualapai Limestone extended from the Pierce Ferry basin into the Gregg basin. The flat-lying, resistant unit to the southeast is the Hualapai Limestone. Note, however, that the unit is displaced by the Wheeler fault (Figure 13). In places up to 1000 feet of displacement occurs. Since the Wheeler fault cuts across the Hualapai Limestone, much of the movement occurred in late Miocene, much later than most of the movement on the Grand Wash fault.

300.0 Entrance to Hualapai Bay. Hualapai Bay occurs on the southeast shore of Gregg basin. The trace of the Wheeler fault extends up Hualapai Wash and eventually merges with Grand Wash fault about 10 miles to the south. The Muddy Creek Formation crops out along the shores of Hualapai Bay. A series of landslides and rotated slump blocks occur on both sides of the entrance to the bay. The fresh-appearing scarps indicate that mass movement was relatively recent. Most likely sliding and slumping was initiated during the record high-water stand of 1983.

301.0 Entrance to Virgin Canyon. The narrow Virgin Canyon is cut into Precambrian granite and metamorphic rock.

301.5 Spring Canyon. Spring Canyon is the small canyon on the left (east). It extends into the Precambrian basement complex. Spring Canyon is a proposed site for a pump reservoir. A dam will be built across the canyon mouth and water pumped into the reservoir during the night when electrical demand is low. During times of peak electrical usage, water will be released from the reservoir for the production of electricity. Unless one is a die-hard environmentalist, this appears to be a rather innocuous construction project. As we pass through Virgin Canyon, however, note the large-scale slump features on both sides of the canyon wall. Similar slump features could be activated in Spring Canyon.

305.0 Entrance to Temple Basin. Although structurally isolated from the Pierce Ferry and Gregg basins, the Temple basin was formed at about the same time and records a similar depositional history. The sand-and-gravel deposits forming the steep cliffs are assigned to the Muddy Creek Formation. Little mapping of these deposits has been done, but no Hualapai Limestone equivalent is recognized. A series of interbedded basalt flows occurs along the east side of the basin.

Stop 12

313.0 Temple Bar. End of field trip, early afternoon. We have traveled approximately 74 miles from Bachhuber and others
REFERENCES CITED


—1981, Grand Canyon monoclines, vertical uplift or horizontal compression?: Contributions to Geology, v. 19, p. 127-134.


Bachhuber and others
INTRODUCTION

The Grand Canyon of the Colorado River in northwestern Arizona is an awesome natural spectacle that to this day defies complete scientific understanding, literary interpretation, and human exploration. After the opening of the Gulf of California in latest Miocene time (5.3 Ma), the Colorado River dissected the southwest portion of the Colorado Plateau to create the Grand Canyon. Lavas found just 50 feet (15 m) above the present river level in western Grand Canyon indicate that the canyon existed at least by 1.8 Ma. As remarkable as it may seem, forces of headward erosion formed the Grand Canyon, within a 4-million-year period! In the Grand Canyon, like few places on earth, rock layers are exposed that span nearly one-third of geologic time. Such exposure requires that geologists should, at least once in their lives, take a walk back through time to attempt to comprehend all that the Grand Canyon reveals.

OBJECTIVES

As part of the 1987 GSA Annual Meeting, this field trip provides the opportunity for a limited group of adventurous geologists to explore the many wonders of the Grand Canyon. By taking a fairly rigorous but rewarding 3-day rim-to-river backpack trek, participants will be introduced to the stratigraphy, geologic evolution, and natural history of the Grand Canyon. It is our hope that this trip will reinforce what we each know about earth history and geologic concepts, while at the same time, humble us with the realization that there is much we do not yet fully comprehend.

TRIP ROUTE

The trip begins on the South Rim of the Grand Canyon near Yaki Point at an elevation of 7200 feet (2200 m). Starting in the marine sedimentary rocks of the Permian Kaibab Formation, we will hike the South Kaibab Trail as it descends through successively older formations, crossing numerous major unconformities, 6.5 miles (10.5 km) to the Colorado River. At river level within the Inner Gorge, we will have traversed the entire Paleozoic section, walked across tilted slivers of middle Proterozoic sedimentary rocks, and hiked into early Proterozoic metamorphic rocks of the Vishnu Group (the oldest rocks exposed in the Grand Canyon). While our legs will feel the drop of more than 4700 vertical feet (1400 m), our minds will be weary from passing through roughly 1.5 billion years of geologic time. There is no need to worry because steaks and cold beverages call us from across the river to Bright Angel Campground and Phantom Ranch.

After a lay-over day spent exploring any number of features in and around Bright Angel Creek, the third and final day will be occupied hiking 9.5 miles (15 km) back up the South Rim via the Bright Angel Trail. Depending on one's lung capacity, there will be ample time to ponder the topographic expression of the different rock layers, the concept of geologic time, and oneself.

TRAIL GUIDES

Trail guides for both the South Kaibab and the Bright Angel Trails have been recently published (Breed and others, 1986; Beus, 1987) and are provided as part of the trip package. Trip leaders will summarize points of interest as we hike down the South Kaibab Trail to the river on the first day. A guided hike (optional) from Phantom Ranch will be offered on the second day. Trip participants will be on their own to follow the geology along the Bright Angel Trail on the hike out the third day.

ACKNOWLEDGMENTS

The authors thank Susan Sabala Foreman for typing the manuscript and the Department of Geology at Northern Arizona University for providing support for its preparation. Among the many famous geologists who have studied rocks and features in the Grand Canyon, Clarence E. Dutton is known for his ability to convey his feelings, thoughts, and ideas about what his experiences in the canyon taught him scientifically, as well as spiritually. The leaders dedicate this field trip to him.

REFERENCES

Upper Holocene Alluvium
of the Southern Colorado Plateau: A Field Guide

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INTRODUCTION

This field trip has three purposes: 1) to demonstrate the widespread distribution of upper Holocene alluvium in several southern Colorado Plateau valleys, 2) to examine and interpret the deposits, and 3) to discuss environmental factors affecting deposition. Two mappable units are present in most of these valleys: an older unit, the "pre-modern alluvium," deposited about A.D. 1400-1880, and a younger unit, the "modern alluvium," deposited about 1940-1980.

The field-trip area is the Little Colorado and Paria River valleys (Figure 1). These are typical Colorado Plateau streams having high-sediment yield, ephemeral discharge, and sand-bed channels that adjust quickly to hydrologic changes. Unlike other streams in the region, the Little Colorado and Paria Rivers have long, continuous discharge records. This makes these streams particularly useful to study because hydrology can be correlated with changes in alluvial-valley morphology.

The geomorphology, sedimentology, and stratigraphy of this upper Holocene alluvium will be examined in the field. This information provides a basis for understanding the timing and processes of regional stream entrenchment and alluviation. In addition, this information has implications for semiarid region land management, geomorphology, archeology, and surface-water hydrology.

The cause of late Holocene fluvial erosion and aggradation in the Southwest is the topic of considerable debate (Cooke and Reeves, 1976), a debate referred to by geomorphologists as "the arroyo problem" (Graf, 1983). Three general hypotheses have been formulated to explain late Holocene aggradation and erosion: 1) poor land use, such as overgrazing; 2) changes in climate; and 3) random variations in the fluvial system that are unrelated to climate.

Stratigraphic correlation of upper Holocene deposits is important to understanding the arroyo problem, but there are no stratigraphic studies of late Holocene valley-fill alluvium. A major objective of this field trip is to demonstrate that various alluvial deposits of specific age are correlative along a particular stream as well as between basins. This regional correlation suggests a common causal mechanism for alluviation, although the specific cause is not well understood for pre-modern alluvium.

DATING METHODS

Upper Holocene alluvium examined on this field trip is dated by tree-ring methods. This dating method has several advantages over conventional radiocarbon dating: it is inexpensive, is not affected by contamination, and yields dates directly comparable with calendar years.

Sample contamination by Cretaceous-age coal and other impurities is a serious problem in Colorado Plateau streams that head in coal-bearing strata. The radiocarbon time scale, furthermore, is not calibrated with calendar years for material younger than about 450 years (Bradley, 1985, p. 68); indeed, a single radiocarbon date gives multiple calendar dates.

Living juniper trees partially buried in alluvium occur in many Colorado Plateau valleys (Figure 2). These trees provide an excellent means of dating alluvium. The field method consists of manual excavation of the tree to its root collar and removal of a transverse cross section of the trunk. A ring count gives the approximate germination date of the tree and the maximum age of the overlying deposit. Ring counts were done by Dennis Boden at the Laboratory of Tree-Ring Research, University of Arizona, Tucson. Juniper trees produce an annual growth ring, although only one specimen collected for this study was cross-dateable with a regional tree-ring chronology.

Modern (post-1940) alluvial deposits were dated in a similar manner except that ring counts were made of partially buried saltcedar and sagebrush, which also produce annual growth rings (Ferguson, 1964; Hereford, 1984).

GEOMORPHOLOGY

Modern Alluvium

Modern and pre-modern alluvium are mappable in the Paria and Little Colorado River valleys. Modern alluvium, deposited mainly since about 1940, includes the active stream channel and floodplain. This alluvium partially fills the entrenched channels cut in the late 1800s during the widely recognized period of arroyo cutting in the Southwest (Bryan, 1925). Deposition of this alluvium produced substantial geomorphic changes in many Colorado Plateau valleys that have been noted by several workers (Eames, 1974; Leopold, 1976; Dunne and Leopold, 1978, p. 692-693; Love, 1983). On this field trip, modern changes in the channels of the Little Colorado and Paria Rivers will be evaluated through comparison of pre-1940 channel photographs with present channel conditions (Stops 1 and 4, Figure 1).

Modern alluvium is much less extensive than pre-modern valley fill. The modern deposits,
Figure 1. Map of field trip showing stops.

nevertheless, are mappable at scales of 1:250,000 in the Little Colorado River valley (Hereford, 1979) and 1:100,000 in the Paria River drainage basin. For example, mapping in the Paria River basin shows that modern alluvium has an average thickness of 2 m, a surface area of about 20 km², and an estimated weight of about 50 million tons. Moreover, during the 40-year history of alluviation, the sediment-delivery ratio of the Paria was about 70 percent, based on measured sediment yield of the river.

Active stream channels in the southern Colorado Plateau typically have either discontinuous or continuous longitudinal profiles. Graf (1985)

Figure 2. Living juniper tree partially buried in alluvium in a tributary of Kaibab Gulch at the Cockscomb (Stop 8, Figure 1). Scale is 1.4 m long.

suggests that the continuous type is predominant. Discontinuous channels have a knickpoint or headcut in older valley-fill alluvium that migrates headward during runoff. Sediment in these streams is derived mainly from erosion of the alluvium at the knickpoint. Discontinuous channels are probably typical of the shorter streams. These channels probably contribute only a small portion of total sediment yield and runoff.

Continuous-channel systems, in contrast, lack knickpoints. Sediment in these channels is derived mainly from lightly weathered bedrock on hillslopes at the head of the stream. Continuous channels consist of two segments having differing transport capacities (Figure 3). Upstream segments have high transport capacities because of steep gradients and narrow channels. Downstream segments, however, have

Figure 3. Generalized distribution of stream power and sediment-transport capacity as a function of location in the drainage network (modified from Graf, 1982, Figure 9.5).
low transport capacities because of low gradients and wide channels. The upstream segments, therefore, are characterized by net sediment transport, whereas downstream segments are characterized by sediment storage. This explains the typical absence of floodplains in upper stream reaches and the presence of floodplains in lower stream reaches. A field-trip stop at Kitchen Corral Wash (Stop 11, Figure 1) illustrates the two types of channel systems and their respective sediment source areas.

Pre-Modern Alluvium

The topographically highest and oldest mappable upper Holocene terrace in the Paria River basin and Little Colorado River valley is underlain by pre-modern alluvium. This terrace is 1 to 5 m above the modern floodplain. The terrace is characterized by large, senescent cottonwood trees (Hereford, 1984), although the trees are not present along every stream. This surface was utilized locally for construction, grazing, and farming by pioneer settlers during the late 1860s to 1880s, suggesting that the surface was at or near its present elevation by that time.

Deposition of the pre-modern alluvium began after about A.D. 1300-1400 in the Paria River basin, and at about the same time elsewhere in the southern Colorado Plateau (Hack, 1942; Dunne and Leopold, 1978). In the Paria River basin and most of the Little Colorado River valley, the alluvium covers older Holocene deposits. Thus, although thin locally, the pre-modern surface is widespread in the alluvial valleys of most streams. Within a decade or two of 1880, streams in southern Utah and most of the plateau widened and deepened rapidly, producing the pre-modern terrace. This stream entrenchment caused problems with irrigation ditches and loss of valuable farmland at pioneer settlements.

In the Paria River basin and probably elsewhere, stream entrenchment was rapid. According to early workers (Gregory and Moore, 1931; Bailey, 1935), channels attained their maximum width and depth between 1883 and 1890. Channels remained in this widened state until the early 1940s, when deposition of modern alluvium began (Hereford, 1986).

Two coincident geomorphic surfaces and corresponding stratigraphic facies are present in pre-modern alluvial valley fill. The valley-axis surface is near the present stream and is approximately parallel with its longitudinal profile (Figure 4). This axial surface grades into the valley-margin surface located in small tributary valleys adjacent to the waning slope of the hillslope system (Figure 5). Alluvium beneath the surfaces interfingers, giving rise to valley-margin and valley-axis stratigraphic facies.

SEDIMENTOLOGY AND STRATIGRAPHY

Modern Alluvium

Modern alluvium is present as floodplain deposits in numerous southern Colorado Plateau streams (Figure 6). These floodplains have physical characteristics in common with other river floodplains (Lewin, 1978), particularly levees and overbank channels (Hereford, 1984). Riparian vegetation such as saltcedar and cottonwood, and nonriparian plants such as juniper, Russian olive, and big sage, are abundant. Many of these plants are partially buried in alluvium; ring counts of saltcedar, cottonwood, and big sage were used by Hereford (1984; 1986) to establish a depositional chronology.

Grain size of floodplain alluvium ranges from sand to clay with minor gravel. Median grain size is very fine to fine sand that is poorly to very poorly sorted. Thickness of the alluvium ranges from about 1.5 to 5 m.

Three stratigraphic units are widespread in southern Colorado Plateau streams. The units are 1) a basal gravel (absent locally), 2) an intermediate thin-bedded unit, and 3) an upper thick-bedded interval (Figure 7). These units typically overlie a fine to medium sand that is coarser grained than the floodplain alluvium. The unit is rarely exposed and is mappable only at very large scales (Hereford, 1986, Figure 7). This alluvium, termed the older channel alluvium (Hereford, 1984; 1986), is probably a channel bar or other transient deposit present
before floodplain aggradation.

Near the top of the upper unit is a marker interval (Figure 7). The interval is distinctive in the stratigraphic column because of its fine grain size compared with beds above and below it. At the Paria and Little Colorado River sections, the interval is a single bed. At the Chinle Creek, Comb Wash, and Lime Creek sections the marker consists of several thin beds of silty, clayey sand interbedded with silty sand.

**Stratigraphic Correlation**

Stratigraphic correlations shown in Figure 7 indicate that deposition of the basal and intermediate units began between 1939 and 1946. This time interval, therefore, is the approximate beginning of regional floodplain aggradation. Evidence from aerial photographs and ground-based photography support this conclusion. The upper interval was deposited after 1954-1956 (Figure 7) until about 1980. Floodplains have not been extensively flooded since 1980 at the localities in Figure 5.

The marker interval, deposited between 1972 and 1974 (Figure 7), demonstrates a high level of depositional synchronicity between drainage basins. This probably resulted from distinctive weather conditions that affected runoff over a large area. Indeed, 1972–1973 was a time of unusual global weather patterns (Kukla and Kukla, 1974) that affected the Colorado Plateau. October 1972, for example, was the wettest October at Flagstaff in 92 years (Hereford, 1984).

Floodplain alluviation was broadly synchronous throughout a substantial portion of the Colorado Plateau, as illustrated in Figure 7. This implies that alluviation was controlled by external factors—probably climate—rather than by internal geomorphic controls (Schumm, 1981). Furthermore,

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Figure 5. Modern alluvium study sites, southern Colorado Plateau.

Figure 6. Modern alluvium study sites, southern Colorado Plateau.

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Figure 7. Cross section showing stratigraphic correlation of floodplain alluvium at four localities in the southern Colorado Plateau. Dated horizons on left side of each column. Parenthetical dates at the Paria River gage section are based on correlation with other sections in Paria River basin (Hereford, 1986, Figure 11). Estimated grain size shown by dashed lines. Note unequal vertical and grain-size scales.
alluviation was progressive—erosion and deposition evidently did not alternate in a random manner. Thus, streams evidently responded in a direct and simple manner to sediment input from the source area; otherwise, correlation among drainage basins would not be possible.

Depositional Processes and Sediment Storage

Floodplains are generally thought to form through either lateral (Wolman and Leopold, 1970) or vertical accretion (Schumm and Lichty, 1963); or through some combination of the two processes (Stene, 1980). Vertical accretion was probably the dominant mechanism of floodplain construction of southern Colorado Plateau streams. This is indicated by the vertical arrangement of beds (Hereford, 1986, Figure 8), the lateral continuity of stratification in transverse floodplain sections, and the absence of epsilon surfaces, which are distinctive cross-cutting contacts suggestive of lateral accretion (Allen, 1963). Moreover, floodplain alluvium overlaps the older channel deposits (Hereford, 1984, Figure 3), which also indicates vertical accretion. These relationships will be observed at field-trip stops 1, 4, 9, and 10 (Figure 1).

Vertical accretion implies that sediment was stored rather than merely redistributed across the channel. Furthermore, because the channel remains approximately fixed during vertical accretion, alluvium deposited on the floodplains is preserved. Channel stability and sediment storage were not typical of the period 1880 through the mid-1940s. Therefore, the poorly preserved, discontinuous deposits of the older channel alluvium probably record erosion or sediment transport, whereas the floodplain deposits record progressive sediment accumulation.

Possible Causes of Modern Sediment Storage

Three causes of floodplain aggradation and sediment storage are possible: 1) land use and human activity, 2) geomorphic controls inherent to the fluvial system, and 3) change in flood discharge due to climatic variation. Land use alone probably does not account for floodplain aggradation and sediment storage. Conservation measures such as construction of dams and stock tanks were implemented on plateau streams after enactment of the Taylor Grazing Act of 1934. These structures were designed to reduce sediment load and peak-flood discharge. Their effect on the larger streams, however, is probably small as suggested by the discharge histories of the two studied streams with long-term hydrologic records. These records do not show a gradual decline in discharge as might result from increased water use and impoundment (Hereford, 1984, Figures 2B and 15B; 1986, Figures 13A and 15B). Furthermore, many plateau streams with floodplains are free flowing through most of their length and lack significant impoundments.

Introduction of nonnative riparian plants, specifically saltcedar, has changed channel systems (Graf, 1978). Saltcedar was brought into the Colorado Plateau in the late 1800s; it spread through the channel systems in the 1930s and 1940s (Christensen 1962; Robinson, 1965) and it presently is the dominant vegetation on the floodplains of many streams.

These plants trapped sediment (Hadley, 1961) and undoubtedly affected sedimentation, but it seems unlikely that they were the single cause of aggradation. First, floodplains are present where saltcedar is not abundant or absent. Second, the plant was under cultivation in several riverside settlements, but photographs of streams taken near the settlements in the early 1900s do not have saltcedar (Hereford, 1984). Finally, although saltcedar spread rapidly through the master streams (Colorado and Green Rivers; Graf, 1978), it did not appear in tributary streams until the late 1930s and early 1940s, after a decline in peak-flood discharge. The spread of saltcedar, therefore, was dependent on a decline in flood discharge—the same decline associated with sediment storage beginning in the early 1940s.

Geomorphic factors inherent to the fluvial system cause simultaneous erosion and aggradation in the channel system, as shown by a number of experiments and field observations (Schumm, 1977). This type of stream behavior is known as the "semiarid cycle of erosion" (Schumm and Hadley, 1957). Basically, sediment accumulates in a widened channel downstream from a migrating knickpoint. This phase of the cycle begins to operate after sediment accumulates on the channel floor to a threshold slope. Above this slope the stream becomes unstable and the system reverses operation, leading to knickpoint migration and erosion. A characteristic of the semiarid cycle of erosion is that erosional and depositional events do not correlate within or between drainage basins (Patton and Schumm, 1981); each basin has a different alluvial history. The simultaneous initiation of aggradation as well as the regional correlation of floodplain alluvium in numerous southern Colorado Plateau streams does not support the hypothesis that the semiarid erosional cycle was the principal depositional process.

A change in peak-flood discharge emerges as the most likely and immediate cause of sediment storage. Although discharge records are not maintained for many streams, the few continuously measured streams show a decline in the frequency of large floods after the early 1940s (Figure 8). The sediment yield of the Colorado River in the Grand Canyon (Figure 9) suggests that a major hydrologic change occurred. Sediment yield at this station declined after 1942 or 1943. This decline resulted partly from improved land use and conservation.

![Figure 8. Cumulative frequency distributions for two periods of Little Colorado and Paria River partial duration flood series. LCR = Little Colorado River 1927-1942 (uppermost curve) and 1943-1980 (second curve from top). PR = Paria River 1924-1942 (second curve from bottom) and 1943-1980 (lowest curve).](image-url)
measures initiated in the 1930s (Hadley, 1974; 1977), but the abrupt decline shown by the sediment-yield time series (Figure 9) suggests that sediment accumulation in tributary-channel floodplains also contributed to the decline. Also, it should be noted that in 1944 the suspended sediment sampler used on the Colorado River was changed. However, comparative field and laboratory tests (available from St. Anthony Falls Hydraulic Laboratory, Minneapolis) between the earlier sampler and its replacement suggest that the samplers yield consistent results, indicating that the change in sampling device does not account for reduced sediment yield.

A change in the climate of the Northern Hemisphere was approximately concurrent with the change in flood discharge and in fluvial process from erosion to aggradation. The hemispheric mean surface air temperature (MSAT) rose by slightly more than 0.5°C between 1880 and 1940; thereafter, MSAT declined by a similar amount (Bryson, 1974; Kukla and others, 1977; Douglas and others, 1982; Jones and others, 1982, Figure 2). These changes were probably caused by a shift in atmospheric circulation (Kalnicky, 1974) that may have altered rainfall patterns (Bryson, 1974; Reitan 1980· Zishka and Smith, 1980). Hence, the fluctuation in

![Figure 9. Time series of Colorado River sediment yield, Grand Canyon National Park, 1926-1962, showing the 62-percent decline in sediment yield beginning about 1942. Horizontal lines are medians for the periods 1926-1941 and 1942-1962.](image)

MSAT is linked through atmospheric circulation to surface runoff and fluvial processes.

Pre-Modern Alluvium

Pre-modern alluvium forms a terrace in valleys of many Colorado Plateau streams. These deposits are identified through geologic mapping in southwestern Utah (work in progress) and the Little Colorado River valley (Hereford, 1979; Ulrich and others, 1984). This field trip will examine dated pre-modern alluvial deposits at several localities in the Paria River basin (Figure 1). Median grain size of pre-modern alluvium ranges from coarse silt to fine sand, although gravel is present locally, particularly near the base of the deposit. The sand-size alluvium is typically very poorly sorted. Thickness depends on position in the valley. Valley-axis alluvium is thickest, ranging from 5-10 m. Thickness of valley-margin alluvium ranges from 1-2 m.

Pre-modern alluvium rests disconformably on older Holocene alluvium. This relationship is well exposed in Kitchen Corral Wash (Figure 10). Three units are present: 1) an older unit of possible late Pleistocene age, 2) an intermediate unit that mostly predates A.D. 1075-1275, and 3) a younger unit of pre-modern alluvium. The late Pleistocene date of the older unit seems unlikely because deposits of known late Pleistocene age in most of the southern Colorado Plateau are gravelly, whereas the older unit is sandy. Possibly the dated carbonaceous material was contaminated with Cretaceous coal. The upper part of the intermediate unit is dated at A.D. 1075-1275, based on archeological materials (Douglas McFadden, written commun., 1985). These materials include two human burials and ceramic sherds of PII-III periods (A.D. 1075-1275) of Kayenta Anasazi affiliation.

Deposition of the younger unit (Figure 10), the pre-modern alluvium of this field guide, followed stream entrenchment sometime after A.D. 1075-1275. The younger unit overlaps the two older alluviums, which explains the general absence at the surface of older units. This exposure (Figure 10) and the archæological site will be examined at Stop 10 (Figure 1).

Valley-margin and valley-axis pre-modern alluvium are distinguished by grain size as well as their previously discussed locations in the valley. Valley-axis alluvium is substantially coarser grained than marginal alluvium. Valley-margin alluvium is typically coarse silt with about 30 percent clay, whereas valley-axis alluvium is very fine sand with only about 8 percent clay. The

![Figure 10. Stratigraphic cross section of early Holocene and early Pleistocene(?) alluvium at Kitchen Corral Wash (Figure 1).](image)
Figure 11. Interfingering of valley-margin and valley-axis facies of pre-modern alluvium at Coyote Wash (Stop 7, Figure 1).

two types of alluvium interfinger as shown in Figure 11. The downstream coarsening of alluvium suggests that fine-grained material is transported through the channel system.

Three stratigraphic units, recognized on the basis of grain size, bedding thickness, and stratification type, are present in pre-modern alluvium (Figure 12). Unit 1 is typically fine to very fine sand that averages about 10 percent clay, although sandy gravel is locally present. Unit 2 typically lacks gravel, is coarse silt to very fine sand that averages about 20 percent clay, and is significantly finer grained than unit 1. At the Kitchen Corral Wash sections (Figure 12), unit 2 is characterized by parallel, continuous stratification with beds of variable thickness. At the remaining sections, stacked sequences of fining-upward cycles are typical of the unit. The basal layer of the fining upward cycle is about one phi. size coarser grained than the overlying finer layer. A dark, sparingly carbonaceous bed at or near the base of the unit is distinctive. Unit 3 is typically a thick-bedded interval that contains from one to

Figure 12. Stratigraphic correlation of pre-modern alluvium in Kitchen Corral Wash area. KCW = Kitchen Corral Wash; VC = Vermilion Cliffs; WST = White Sage Wash tributary; FMV = Five Mile Valley; CKGT = Cockscomb at Kaibab Gulch tributary. KCW1 and KCW2 are in valley-axis alluvium. Cross-hachures show extensively bioturbated intervals. Note scale change.
several beds. The top of this unit is the pre-modern terrace.

Age and Regional Correlation

Figure 12 shows the depositional chronology of pre-modern alluvium in the field-trip area. From the dated horizons (rounded to the nearest century), I conclude that unit 1 was deposited after 1300 until 1700, that unit 2 was deposited from 1400 until 1800, and that unit 3 was deposited between 1800 and 1900. In short, pre-modern alluvium post-dates 1300; most of the exposed deposits, moreover, post-date 1400. This pre-modern alluvium of southern Utah is probably equivalent to the late Holocene Naha Formation in northeast Arizona (Hack, 1942, Karlstrom and Karlstrom, 1986) and southeast Utah (Christenson and others, 1985).

Depositional Processes

Large quantities of sediment were stored in many valleys during the 500 to 600 years of pre-modern aggradation. This deposition filled a pre-existing, entrenched channel system that was up to 10 m deep (Figure 10) and at least as wide as the modern channel. Alluviation extended far into tributary valleys (Figures 4 and 5), increasing their elevation by 1-2 m.

In the broadest sense, pre-modern alluvium constitutes a major fining-upward cycle. Unit 2, the fine-grained component of the cycle, is characterized by a number of minor fining-upward cycles. At present, the origin of these minor and major cycles is not clear. The minor cycles in unit 2, however, probably did not develop in meandering or braided streams, the typical fining-upward cycle depositional environments; they are too thin and too fine grained, and lack the rich variety of sedimentary structures found in better known fluvial fining-upward cycles.

ROAD LOG

The road log begins in Flagstaff, Arizona at the lobby pull-in of the Little America Hotel, located near the Butler Avenue interchange of Interstate 40.

0.0 Exit Little America parking lot and enter I-40 eastbound on ramp.

3.0 Turn right at Exit 201 to Grand Canyon and Page on U.S. 89 north.

51.6, MP 466.8 Cameron Trading Post near Stop 1. Turn left at the gas station then turn right and proceed 0.1 mi to old highway bridge.

Built in 1911, this bridge was a vital transportation link that made automobile travel practical north of the Little Colorado River. Before 1911, the river was crossed 1.5 mi upstream at Tanner’s Crossing where there is a break in the canyon walls. Shifting channels and quicksand made this a dangerous crossing that claimed several lives.

Stop 1

Little Colorado River

The purpose of this stop is to examine modern changes in the Little Colorado River channel.

Figures 13 and 14 show the condition of the channel in the early 1900s. Since 1937, channel width has decreased about 50 percent through deposition of 2 m of sediment on both sides of the channel. These deposits are present as a floodplain along a 100-mi reach of the river from Cameron to near Winslow, Arizona (Figure 1).

Changes in the Little Colorado River channel were evidently related to a decrease in the frequency of large floods that began in the early 1940s.
Hereford was River on north side of highway. 122.6. 123.

Moenkopi Wash Modern and modern alluvium is typical of most modern floodplain and the pre-modern terrace base of Echo Cliffs are a rotational landslide, or spaned Marble Canyon 467 ft above the Colorado

78.1 0 MP about of highway is a Pleistocene debris overlie incompetent shale formations.

51.9, MP 466.8 Torn north on U.S. 89. 62.3, MP 477.2 Moenkopi Wash and Stop 2.

Stop 2

Moenkopi Wash

At this stop the inset relationship between the modern floodplain and the pre-modern terrace will be examined. This relationship is best viewed a short distance upstream or downstream from the highway bridge.

Modern deposits have a moderately dense to light growth of saltcedar and are inset several meters below pre-modern alluvium. Pre-modern alluvium probably post-dates A.D. 1400, although the alluvium has not been dated at this locality. Large, senescent cottonwood trees are present locally at or slightly below the pre-modern terrace. According to Gregory (1917, p. 131), Moenkopi Wash was entrenched sometime after 1878.

This cut-and-fill relationship between pre-modern and modern alluvium is typical of most southern Colorado Plateau streams. Modern alluvium partially fills the arroyos cut in the late 1880s.

76.1, MP 491 Echo Cliffs and Hamblin Wash on east side of highway.

78.1 MP 493 Early (?) Pleistocene alluvium, terrace, and pediment on both sides of road for next 9 mi.

83.1 MP 498 To the east is The Gap (a wind gap in the Echo Cliffs) and The Gap Trading Post and service station.

107.1 MP 522 Steeply dipping sandstone ledges at base of Echo Cliffs are a rotational landslde, or toreva block. This type of landslide is common in the Colorado Plateau where competent sandstone beds overlie incompetent shale formations.

111.1, MP 526 Blocky surficial deposit on east side of highway is a Pleistocene debris flow. Vermilion Cliffs and Paria Plateau to northwest.

122.6, MP 537.5 Pleistocene gravel of the Colorado River on north side of highway.

123, MP 537.9 Navajo Bridge, Marble Canyon, and the Colorado River. Completed in 1929, Navajo Bridge spans Marble Canyon 467 ft above the Colorado River. This is the only crossing for automobiles between Glen Canyon and Hoover Dam, a distance of about 360 mi as measured along the Colorado River. Before 1929, the river was crossed by ferry at Lees Ferry near the head of Marble Canyon. This crossing was dangerous and several people drowned in the 57-year history of the ferry.

123.3, MP 538.2 Junction with Lees Ferry road; turn right. Five Mile Point ahead, a landmark for early southbound travellers, signalling only 5 mi to Lees Ferry and the dangerous Colorado River crossing.

123.9, MP 0.6 Cathedral Rock and Stop 3.

Cathedral Rock

This stop illustrates the slow rate of change typical of Colorado Plateau hillslopes. This slow change contrasts with the channel system that has the potential for rapid change. Figure 15A is a reproduction of a photograph taken sometime between 1871-1873 (Wheeler, 1875) of Cathedral Rock, an erosional outlier of slope-forming Moenkopi Formation (Triassic). Compare this photograph with Figure 15B, an approximate relocation of the earlier photograph. Little change is apparent on this easily eroded hillslope. This apparent lack of change is widely noted in Colorado River canyons (Baars and Molenaar, 1971, p. 90-99). Yet these same hillslopes are the principal source of the abundant sediment transported by Colorado Plateau streams. Actually, hillslope erosion is evident. The fresh, light-colored scree slope near the center of Figure 15A has been removed through erosion. The two light-colored boulders on the slope above and to the right of the soldiers now rest on an erosional pedestal several centimeters above the slope.

Hillslopes such as Cathedral Butte are the principal source of sediment for Colorado Plateau streams. Although the rate of hillslope erosion is low at a single locality, the vast exposures of unvegetated hillslope material provide abundant sediment to the channel system. Moreover, it is apparent that even a slight change in hillslope-weathering rate would have a large effect on quantity of sediment supplied to the channel system.

125.5, MP 2.1 Pedestal rocks on left indicate up to 2 m of surface lowering since they fell from the overlying sandstone cliff. This probably occurred sometime in the middle to late Pleistocene, as this type of colluvium interfingers with Pleistocene gravel.

128.6, MP 5.1 Paria River and Lees Ferry (Lonely Dell) Ranch road; turn left.

128.8 Lonely Dell Ranch Historic District parking area, south of Paria River Canyon, and Stop 4. Cross the fence and walk north about 0.5 mi to Lonely Dell Cemetery, then walk northeast along the Paria River 0.2 mi to the gaging station.

Lees Ferry and the Lonely Dell Ranch area have a long history that is well described in Rusho and Crampton (1981). Located between two virtually uncrossable canyons, Glen Canyon upstream and Marble Canyon downstream, this area has been a Colorado River crossing since prehistoric times, even though crossing requires swimming at most river levels.

The first written account of this area is in the Dominguez-Escalante Journal (Chavez, 1976, p. 93-95), an account of an expedition through the Southwest in 1776. The expedition camped on the banks of the Paria River near the Colorado River just east of Lonely Dell Ranch from October 27 to November 1, 1776. Unable to ford the Colorado
River, the expedition climbed the east wall of Paria River Canyon about 2 mi north of Lonely Dell and crossed the Colorado River at Paiute Crossing, or Crossing of the Fathers, about 25 mi north of here. Lonely Dell and Lees Ferry were settled by J. D. Lee in late 1871. Lee was an infamous southern Utah pioneer and religious zealot forced to resettle in this area because of religious and legal difficulties. In the 1920s, Lees Ferry was designated the "compact point," a geopolitical boundary between the lower and upper Colorado River basins. Lees Ferry is now a recreational area for fishermen in Glen Canyon and river runners embarking for the Grand Canyon.

Stop 4
Paria River Gaging Station

At this stop, modern deposits are particularly well exposed. A series of photographs of the channel near the gaging station taken by U.S. Geological Survey personnel from 1931 to 1964 document sediment storage beginning after 1939. Figure 16 illustrates the channel changes immediately upstream from the gaging station.

A stratigraphic section of modern alluvium just upstream from the gage (near scale Figure 16B) is

Figure 16. A) Downstream view of the Paria River channel taken in October 1939. B) Similar view in January 1985 (from Hereford, 1986, Figure 5).
illustrated in Figure 17. Regional correlation of the alluvium in the Paria River basin is shown in Hereford (1986, Figure 11). At this stop, it is important to note that the alluvium at the gage correlates with deposits in the Paria River basin and with modern alluvium in other southern Colorado Plateau streams (Figure 7). A substantial change in the frequency of large floods occurred at this station in the early 1940s (Figure 8), and this change corresponds closely in time with the beginning of sediment storage.

134.2, MP 538.2 Junction with U.S. 89A; turn right.

137, MP 541 Pleistocene alluvial fan both sides of road for next 3 mi; deposits grade upslope into debris flows.

140, MP 544 Two early (?) Pleistocene debris-flow lobes at base of lowest cliff.

143.8, MP 547.8 Late Pleistocene alluvial fan or pediment exposed almost continuously, except in washes, for next 15 mi.

150.7, MP 554.7 Debris flows and toreva blocks on north side of highway.

153.7, MP 557.7 Dominguez-Escalante historical marker; turn right to Stop 5.

Stop 5

Dominguez-Escalante Exhibit

This historical exhibit (administered by Bureau of Land Management) shows the route of the 1776 Dominguez-Escalante expedition and illustrates the Mesozoic stratigraphy exposed in the Vermilion Cliffs. This stop is useful because the Dominguez-Escalante Journal describes, often in detail, environmental conditions along the expedition route. Mesozoic formations exposed in the cliffs are the principal sediment-producing units in this part of the Colorado Plateau.

153.9, MP 557.7 Turn right on U.S. 89A.

161.8, MP 565.6 House Rock Valley road and House Rock Ranch (abandoned); turn right (north). This road is maintained as necessary and is usually passable with low-clearance vehicles. Use caution, however, in wet weather because the road can be muddy and crosses several washes.

170.6 Drainage divide between House Rock Valley and Coyote Valley in Paria River basin. Pink Cliffs at distant horizon are the headwaters of the Paria River. Coyote Buttes to the northeast.

178.5 Junction with Coyote Ranch road and Stop 6. An option at this point is to drive to Coyote Ranch for a closer inspection of Coyote Spring and the topic of Stop 6 discussion.

Stop 6

Coyote Spring

This stop is at or very near the route of the Dominguez-Escalante expedition, and a discussion of the environmental conditions they described is relevant. The expedition camped at Coyote Springs on the east side of Coyote Wash October 22-24, 1776 (Chavez, 1976, p. 88-92) after descending Buckskin Mountain to the west. In the vicinity of Coyote Spring, the journal mentions two water holes, a small (shallow) arroyo, and a spring of permanent water that apparently was not Coyote Spring. East and west of Coyote Spring they found pasturage for their horses. I infer from the journal that Coyote Wash was not entrenched in 1776, whereas now it is entrenched for its entire length. A tentative conclusion is that in the 1770s there was more surface water in this area, and streams were not entrenched. This conclusion is supported by the journal entry for October 20, 1776 (Chavez, 1976, p. 87-88). Camped at Bulrush Wash southwest of Fredonia (Figure 1), Escalante states that the area near the campsite had "... a great supply of water and good pasturage." Today there is no surface water at this locality and Bulrush Wash is deeply entrenched.

179.5 Road begins short, steep descent into Coyote Wash.

179.8 Coyote Wash crossing. Valley fill on either side of wash is pre-modern alluvium with modern alluvium inset beneath.

181.9 Coyote Wash study site and Stop 7. Site is located in sharp channel bend that undercuts road. Three conspicuous red buttes with buff sandstone ledges located on west side of valley fill. Excavated juniper tree is located on east side of wash about 50 ft downstream from large, light-colored rectangular block of Shinarump Sandstone (Triassic).

Stop 7

Coyote Wash

The purpose of this stop is to examine the interfingering of valley-axis and valley-margin facies of pre-modern Holocene alluvium in a typical alluvial valley. Figure 11 illustrates interfingering of the two facies exposed on the east side of Coyote Wash. Figure 4 is a photograph showing the extensive valley-axis surface as viewed downstream from the butte west of the road. Two sediment sources are evident: Hillslopes on both sides of
the valley supply fine-grained, very poorly sorted sediment. Upstream sources supply relatively coarse grained, somewhat better sorted sediment.

Stratigraphic units exposed here are traceable upstream for several miles beyond Coyote Spring. Thus deposition of some of these deposits was in un entrenched channels under conditions suggestive of more surface water than at present.

182.6 State line. Road continues on pre-modern alluvium for next 1.5 mi.

183.1 Coyote Wash crossing downstream from junction with Pine Hollow Wash.

183.7 Road in Coyote Wash for next 0.1 mi. Modern alluvium inset beneath pre-modern valley fill on both sides of wash.

184.1 Wire Pass trailhead (Paria Canyon-Vermilion Cliffs Wilderness Area) and northeast bend in Coyote Wash. Not far downstream from here, Coyote Wash enters a narrow slotlike canyon. Valley-fill deposits are not present in this type of canyon, although local slawater deposits in such settings preserve flood history under appropriate conditions (Webb, 1985).

184.3 Road crosses bedrock of Moenave Formation (upper Triassic? and lower Jurassic) at or near contact with Chinle Formation for next 1.6 mi.

185.9 Drainage divide; road enters a small tributary of Kaibab Gulch.

187.2 Road crosses pre-modern valley fill for next 2 mi with many dead and living, partially buried juniper trees. Note that active channel is narrow and lacks a floodplain or other evidence of sediment storage. Stream power and sediment transport capacity of this small tributary probably are large and increase downstream, as illustrated in Figure 3.

188 Kaibab Gulch trailhead turnoff; continue north. This is a wide, alluviated area between two narrow canyons, upper and lower Kaibab Gulch. Prominent red cliffs to east are The Cockscomb; Buckskin Mountain is to the west.

188.1 Kaibab Gulch crossing.

189.2 Kaibab Gulch tributary study area and Stop 8. Measured section in wash 75 ft east of road. Road crosses bedrock for next 1.7 mi.

Stop 8
Kaibab Gulch Tributary

At this stop a measured section of valley-margin alluvium (Figure 3) will be examined. The excavated juniper tree is dated at A.D. 1442; its stratigraphic context is shown in Figure 12 (section C8GT). Most sediment at this locality is derived from nearby hillslopes. The alluvial surface traces continuously downstream to valley-axis surface at Kaibab Gulch.

190 Junction with Five Mile Mountain road; continue north.

192.5, MP 25.7 Junction with U.S. 89; turn left (north). Highway crosses west side of Five Mile Valley. The Cockscomb is to the east and Five Mile Mountain is to the west.

193.8, MP 27 Pre-modern alluvium on both sides of highway.

195.4, MP 28.6 Five Mile Valley study-site turnoff; continue north. Stratigraphic section shown on Figure 12 (section FMV). Sand Gulch to west.

197.6, MP 30.8 Turnoff for Paria (also spelled Pahreah) towns site and Stop 9; turn right. This road is treacherous when wet.

201.3 Colorful strata at base of The Cockscomb is the Petrified Forest Member of the Chinle Formation.

201.7 Road very dangerous beyond here when wet. View of Paria River valley and Paria towns site to north.

202.3 Paria movie set and campground. Continue to Paria River on poorly maintained road.

202.8 Paria cemetery.

203.4 Paria River, Paria town site study area, and Stop 9. Road ends on the floodplain at the west bank of the Paria River channel. Paria towns site is on the opposite side of the river on the pre-modern terrace.

The broad alluvial flats in this area appeared favorable to the pioneers for irrigation and settlement. Paria was settled in 1874 and reached a maximum population of 107 in 1884 (Gregory and Moore, 1931, p. 30). Floods converted the formerly narrow channel into a wash that in places extended across the valley and forced abandonment of the town in 1885. About 1910-15, the area southeast of the townsite was mined hydraulically for gold in the Chinle Formation.

Stop 9
Paria Townsite

Modern and pre-modern alluvium are well exposed at this study area. Photographs of this area taken in 1918 (Gregory and Moore, 1931, plt. 27A) show that the floodplain was not present and that the channel was then wider, sandy, and lacked vegetation. Pre-modern alluvium has not been dated here due to the lack of dateable material. The alluvium is similar, however, to pre-modern valley-axis alluvium elsewhere in having large, old cottonwood trees and cultural structures dating from the late 1800s at or just below the terrace surface.

This stop has three purposes: 1) to examine the erosional contact between modern and pre-modern deposits, 2) to examine the sedimentology of modern alluvium, and 3) to point out evidence for channel adjustment since 1980. The unconformity between pre-modern and modern alluvium is exposed on the east-side terrace rise near the large cottonwoods. Floodplain alluvium is well exposed on the west bank just upstream from this locality (Hereford, 1986, Figure 11, PR 4) and the deposits here correlate with modern alluvium in other drainage basins. Recent channel adjustment is indicated by the cutbanks and scoured floodplain surface visible from the low butte on the west bank upstream from this locality.

The conclusion is that the modern alluvium has been deposited since slightly before 1950 (Hereford,
1986, Figure 11, PR 4). Modern flood deposits here, moreover, correlate with those in other basins suggesting a broadly similar flood history for the entire Paria River basin.

209.2, MP 30.8 Junction with U.S. 89; turn right.

215.4, MP 37 Junction with Kitchen Corral Wash road, turn right (north). This road is usually well maintained. Pleistocene gravel and terrace east side of road.

215.9 Road crosses pre-modern valley fill.

216.6 Well-exposed pre-modern alluvium and Kitchen Corral Wash floodplain. Channel meander has migrated east from line of saltcedars on floodplain since 1983. Further channel migration threatens road.

217 Kitchen Corral Wash study area and Stop 10. Study area is reached by a short walk along sandy, unimproved road that ends at the wash. Cross section in Figure 10 was made from exposures in east arroyo wall near fence line downstream from end of road. Previously mentioned P-II archeological site is on west arroyo wall about 300 ft upstream from end of road.

**Stop 10**

**Kitchen Corral Wash**

The principal purpose of this stop is to examine the stratigraphic context of Holocene alluvium shown in Figure 10. This site demonstrates that post-1400 A.D. alluvium was deposited in and eventually overtopped a preexisting arroyo. This arroyo was entrenched shortly after human occupation of the area, about A.D. 1075-1275.

A second purpose of this stop is to examine the sedimentology of modern alluvium in a Paria River tributary. Sedimentology and regional stratigraphic correlation of modern alluvium at this site are shown in Hereford, 1986 (Figure 11, KCW). Apparently, Kitchen Corral Wash has a flood history similar to other streams in the Paria River basin.

218.1 Junction with unimproved road. Walk (or drive if you have a high-clearance vehicle) 0.5 mi to Stop 11.

218.6 Small reservoir and Stop 11.

**Stop 11**

"Conservation Corps" Reservoir

This stop addresses three aspects of hillslope sediment yield that affect fluvial processes: average runoff recurrence interval, sediment yield of individual runoff events, and sediment-yield history. These sediment-yield characteristics are inferred from the deposits in this reservoir, which are mapped in Figure 18. Drainage-basin area upstream of the reservoir is 2.76 km², and relief in the basin is large, which is typical of the region. Built by the U.S. Civilian Conservation Corps in 1937, the reservoir operated with high trap efficiency until 1976 when the spillway washed out.

Sediment in the reservoir comes mainly from erosion of the steep, barren hillslopes of the Vermilion Cliffs. This can be verified by tracing active-channel deposits upstream to the base of the cliffs. An easy 1.5-mi walk in the wash at the northeast reservoir inlet (Figure 18) brings one to the base of the cliffs.

Stratigraphy and sedimentology of the reservoir fill are shown in Figure 19. Two types of deposits are distinguishable by grain size. Sediment just upstream from the dam is mostly clay with minor silt; these are the "ponded deposits" (Figures 18 and 19). The bulk of the sediment is clayey sand; these are the "delta deposits" in the delta area of the reservoir.

The number of runoff events was inferred from the ponded deposits (Figure 19). Stratification in these deposits shows that sediment entered the reservoir fill in a single event.
reservoir only 21 times in 38 years, giving an average runoff recurrence interval of 1.8 years. Thus, on the average, the particular combination of rainfall intensity, duration, and antecedent moisture conditions producing runoff did not recur often. Certainly not every rainfall produced runoff, which is contrary to most sediment-yield models.

Sediment yield of individual runoff events was inferred from the estimated volume of correlative beds shown in Figure 19. Average sediment yield was 2,500 m³/km² (5.3 ac ft/ft²) with standard deviation of 1,300 m³/km² (2.7 ac ft/ft²). Variation in sediment yield was not great, being slightly less than one order of magnitude (Figure 20). This small variation is not expected of small basins, and suggests that sediment yield is limited by the availability of freshly weathered material in the hillslope system.

Sediment-yield history was evaluated using runoff recurrence interval in a random-walk sedimentation model. This model simulates a time-stratigraphic section of reservoir deposits resulting from randomly spaced runoff events. The sediment-yield history derived from the model is illustrated in Figure 21. The methods used to generate this Figure have practical application. Using this method, land-use specialists can estimate the useful lifespan of existing reservoirs. In addition, these methods will be useful in designing new reservoirs if runoff recurrence interval and sediment yield per event are known for a nearby reservoir.

221.8, MP 37 Junction with U.S. 89; turn left to return to Flagstaff.

233.1, MP 25.7 Junction with House Rock valley road; continue on U.S. 89.

238.1 Paria River. PRL section (Hereford, 1986, Figure 11) was measured on west bank of river. END OF ROAD LOG. Flagstaff is 168 mi; gasoline available at Page, Arizona.
Late Pleistocene Alluvium and Megafauna Dung Deposits of the Central Colorado Plateau

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ABSTRACT
Recently discovered (post-1983) dung deposits, derived from extinct Pleistocene megafauna and interbedded with cave sediments or capping alluvial deposits, have presented new unduplicated records of the paleoenvironment of the central Colorado Plateau. The pre- and post-dung-deposit alluvium provides the medium in which records of paleohydrology, paleontology, palynology, and a C-14-controlled chronology are preserved. The alluvial record, combined with the botanical and faunal records from the dung beds, provides an opportunity to examine the late Pleistocene environment. These data provide an examination of the cause of Pleistocene megafaunal extinction from "the far side," i.e., ±40,000 BP to the Pleistocene-Holocene boundary (±10,000 BP). They also provide the oldest, nonduplicated paleoenvironmental record of the central Colorado Plateau.

INTRODUCTION
The arid Southwest has been the focus of paleoenvironmental studies for approximately half a century (Antevs, 1939). Early pollen work proved instrumental in developing a framework in which to begin a systematic reconstruction of the environments of the latest Pleistocene. Martin (1963a,b; Martin and Mehringer, 1965) illustrated that certain cienega bogs and pluvial-lake regions in southwestern Arizona contained a late Pleistocene pollen record. Mehringer (1967a,b; Mehringer and others, 1967) found that certain limited pollen information could be recovered from spring mounds and from occasional alluvial sections.

The advent of pack-rat (Neotoma spp.) midden analysis (Wells and Jorgensen, 1964) provided a major thrust into the understanding of the late Pleistocene paleoenvironments of the Southwest. Arid climate coupled with the episodic fluctuating water tables proved detrimental to the preservation of most exposed fossil-pollen locations in the Southwest. However, the same xeric conditions, when coupled with a stable rock shelter, provided a unique situation - dry preservation. Such xeric locations provided not only the preservation of pollen and macrobotanical remains, but also the soft-tissue and other usually biodegradable remains of animals (such as skin, hair, keratinous tissues, and dung; Wilson, 1940). An entirely new area of research was opened to understanding the latest Pleistocene.

The study of pack-rat middens in the Southwest provided a reconstruction of the Wisconsinan plant communities never before observable in such detail (Wells and Berger, 1967; Mehringer and Ferguson, 1969; Van Devender, 1973; Phillips and Van Devender, 1974; reviewed in Spaulding and others, 1983). The dry cave deposits were quickly seen as a warehouse of late Pleistocene information. Gypsum Cave (near Las Vegas, Nevada) and Rampart Cave (Grand Canyon, Arizona) were the scenes of the first paleoecological studies utilizing dry preserved dung of an extinct animal. Laudermilk and Munz (1934, 1938) found a wealth of information preserved in the dung of the extinct Shasta ground sloth (Nothrotheriops shastensis). Later studies expanded on the data available from the sloth dung (Martin and others, 1961; Hansen, 1976; Spaulding and Martin, 1979; Thompson and others, 1980). Dry cave deposits were found to contain the occasional limited record of pollen (Schwartz and others, 1958; Hevly, 1964; Spaulding and Petersen, 1980). Similar studies have begun on the dry preserved remains of the extinct mountain goat, Oreamnos Harringtoni (Mead and others, 1986; Mead and others, in press).

The study of the faunal remains of the latest Pleistocene in the Southwest has never received the full attention of paleontologists as it has in other regions of North America (Kurtén and Anderson, 1980). Lindsay and Tessman (1974) provided the first checklist of Pleistocene age localities in Arizona, and Harris (1977) provided the same for southern New Mexico/TransPecos Texas. Most fossil locations in studies during the early years were the records from alluvial locations, but there was the occasional report of animals from the dry cave (Harrington, 1933; Wilson, 1940). Pack-rat middens provided a wealth of information about the herpetofaunas of the arid Southwest that was not available in the alluvial records (Van Devender and others, 1977; Van Devender and Mead, 1978; Mead and Phillips, 1981; Mead, ms. subm.).

The most recent reviews of the faunal record of the arid Southwest illustrate that al-
though important and pertinent information is known, this information is spotty. There is a physical gap in the record - the majority of the Colorado Plateau (Lundelius and others, 1983; Harris, 1985; Agenbroad and Mead, in press). Some of the most complete information on animals, which was recovered in direct association with plant remains, comes from dry cave deposits in the Grand Canyon, which is ecologically atypical of the Colorado Plateau. The review of the vegetational history of the Southwest also points to a significant gap in the overall record in the central Colorado Plateau (Spaulding and others, 1983).

The alluvial record in the Southwest has received fair attention (Bryan, 1928, 1950; Antevs, 1939, 1952, 1954, 1955, 1962). Many of the alluvial stratigraphic studies have centered in the Basin and Range Province, those locations that also have the remains of paleo-Indians (Sayles and Antevs, 1941; Haury, 1950; Haury and others, 1959; Haynes, 1966, 1967, 1968, 1980; Waters, 1985; to name a few). The alluvial record of the Colorado Plateau has received some attention (reviewed in Christenson, 1985), although much of this record has centered on Holocene deposits (Hack, 1942; Lance, 1963; Agenbroad, 1975; Lipe and Matson, 1975; Lipe and others, 1975). The most recent research on the central Colorado Plateau has occurred in the Escalante River drainage (Boisson and Patton, 1985; Webb, 1985).

OBJECTIVES, METHODS, AND BACKGROUND OF THE CURRENT RESEARCH ON THE CENTRAL COLORADO PLATEAU

Objectives

Our research focus is a reconstruction of the late Pleistocene and early Holocene paleoenvironments of the central Colorado Plateau. Of necessity, this reconstruction will include the fauna, flora, paleoclimatic, paleohydrologic, and geologic parameters.

Methods

Our approach entails the utilization and integration of a set of data derived from a variety of disciplines and subdisciplines, i.e., an integrated, multidisciplinary approach. To date, we have combined archaeological, geological, macrobotanical, palynological, paleohydrological, and paleoentological evidence to construct our working model. We have initiated the use of hair analysis of extinct fauna to supplement our osteological/dental faunal evidence. We have taken fecal analysis of extinct fauna to greater detail and results than prior investigations in this geographic area. We are attempting to initiate biochemical analyses that should produce new support for the paleontological data, as well as shed insight into the problems of dietary stress/nutritional aspects of the late Pleistocene paleoenvironment. The working model we have generated will be tested and revised as we acquire added information and investigate newly discovered and undiscovered sites.

DUNG AND ALLUVIUM

Pleistocene Dung Deposits

Cowboy Cave in south-central Utah (Figure 1) produced a large dung bed underlying the Holocene archaeological deposits (Jennings, 1980). It contained the most complete chronological and paleoenvironmental data from a single locality for the late Pleistocene and Holocene of the Colorado Plateau. The dung bed was composed primarily of bison dung; however, mammoth, horse, ground-sloth, big-horn-sheep, deer, elk, and entmastic "camel-elk" dung or dung fragments were also identified (Hansen, 1980). Hairs attributed to bison, canids, and humans were collected from the dung bed; some hairs are still unidentified. Size-fraction analysis of the dung components, plus macrobotanical studies, produced floral information that would not otherwise be evident.

Investigations of a second dry-cave dung bed began in 1983 at Bechan Cave, (Figure 2) in the southern Utah canyonlands (Davis and others, 1984). Beneath the Holocene-age geological and cultural deposits, an extensive dung bed was encountered (Figure 3), which was composed primarily of trampled mammoth dung, but contained boluses and pellets of other animals as well.
Analysis of the dung, hair, and skeletal components of the deposit provided a bestiary that would have been incomplete from one line of evidence alone. Micro- and macrobotanical samples from the dung bed produced an environmental reconstruction much different than the modern vegetation community.

October 1985 and 1986 produced the discovery and cursory surficial sampling of new dung beds overlying alluvial sequences in alcoves of a nearby canyon. An additional locality, in an alcove in a canyon of Natural Bridges National Monument, produced dung deposits of the extinct mountain goat.

Figure 2. Bechan Cave entrance.

October 1985 and 1986 produced the discovery and cursory surficial sampling of new dung beds overlying alluvial sequences in alcoves of a nearby canyon. An additional locality, in an alcove in a canyon of Natural Bridges National Monument, produced dung deposits of the extinct mountain goat.

Figure 3. Contours of equal thickness of the dung bed, Bechan Cave (C.I. = 10 cm).

Chronology of Pleistocene Dung Deposits

Cowboy Cave (Jennings, 1980) provided the first Utah example of a large-volume deposit of Pleistocene megafauna dung dating from 12,000-11,000 BP (Table 1). The discovery and investigation of Bechan Cave in 1983 (Agenbroad and others, 1984; Davis and others, 1984; Mead and others, 1986), provided another central Colorado Plateau deposit also dating ±12,000 BP. Newly discovered deposits capping Pleistocene alluvial sequences were located in alcoves within nearby canyons. Initial dates on the fecal material indicate a doubling of the recorded antiquity of such deposits on the central Colorado Plateau (back to ±28,000 BP). These deposits also provide an unduplicated source of information for late Pleistocene megafauna - material critical to the investigations into the cause of extinction of these species.

Natural Bridges National Monument also provided new fecal, dietary, and paleoenvironmental information. The oldest layer of these deposits of extinct mountain-goat dung exceeds the limits of radiocarbon dating. For the first time, we can approach the late Pleistocene extinction controversy from "the far side," i.e., moving upward in time from the ±40,000 BP limit of C-14 dating, to the extinction event at ±11,000 BP, via the analysis of biological by-products of the extinct animals - the dung.

Alluvium

Bechan Cave occurs in a box-canyon system containing up to 20 m of alluvium. The recent erosion has exposed sections of the alluvium containing organic bands signifying high watertable conditions favoring plant growth. It was apparent that a C-14-controlled analysis of the alluvium would produce a chronologic sequence of geologic, hydrologic, paleoclimatic events that might correlate to the cave-fill record. This type of correlation was not possible in the canyon in which Cowboy Cave occurs owing to the lack of significant alluvial deposits near the cave.

The alcove deposits are in the proximity of Bechan Cave and overlie alluvial sequences of up to 73-m thickness. Information to be derived from detailed studies of these sediments promises to extend our paleoenvironmental reconstruction of the central Colorado Plateau to unprecedented antiquity.

Alluvial Chronologic Framework

Discovery, investigation, and analysis of the unique dung deposits of Pleistocene megafauna in Bechan Cave (Agenbroad and others, 1984; Davis and others, 1984; Mead and others, 1986) initiated a research effort to establish an alluvial chronology and paleoenvironmental sequence that could be correlated to the cave deposits. Research of the alluvial deposits in the canyon system containing Bechan Cave was begun with support from the National Park Service and the National Geographic Society. The preliminary conclusions are as yet unpublished, except in reports to the funding agencies. The results of the research include an alluvial paleohydrologic model that extends beyond 37,000 BP. Fig-

Agenbroad and Mead
TABLE 1
C-14 Chronology of Pleistocene megafauna dung: Colorado Plateau

<table>
<thead>
<tr>
<th>Locality</th>
<th>Mt1 Dated</th>
<th>C-14 Age Span and Lab. Numbers</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cowboy Cave</td>
<td>Artiodactyl dung</td>
<td>11,020±180 (A-1660)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>13,040±440 (A-1654)</td>
</tr>
<tr>
<td>Bechan Cave</td>
<td>Mammuthus dung</td>
<td>11,670±300 (A-3212)</td>
</tr>
<tr>
<td></td>
<td>cf. Eueratherium</td>
<td>13,505±680 (Gx-9371)</td>
</tr>
<tr>
<td></td>
<td>dung</td>
<td>11,630±150 (Beta-18269)</td>
</tr>
<tr>
<td>Stanton Cave</td>
<td>Oreamnos dung and</td>
<td>10,870±200 (A-1155)</td>
</tr>
<tr>
<td></td>
<td>horn sheaths</td>
<td>19,690±1200 (A-3083)</td>
</tr>
<tr>
<td>Natural Bridges</td>
<td>Oreamnos dung</td>
<td>23,350±1740, (Gx-11313)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>&gt;39,800 (Gx-11312)</td>
</tr>
<tr>
<td>Rampart Cave</td>
<td>Oreamnos dung</td>
<td>18,430±300 (A-1278)</td>
</tr>
<tr>
<td></td>
<td>Nothrotheriops</td>
<td>10,780±200 (A-1067)</td>
</tr>
<tr>
<td></td>
<td>dung</td>
<td>&gt;40,000 (A-1042)</td>
</tr>
<tr>
<td>Muav Cave</td>
<td>Nothrotheriops</td>
<td>11,140±160 (A-1212)</td>
</tr>
<tr>
<td></td>
<td>dung</td>
<td>11,290±170 (A-1213)</td>
</tr>
<tr>
<td>Tse'an Bida Cave</td>
<td>Oreamnos dung</td>
<td>11,850±750 (RL-1134)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>24,190±4300 (A-2373)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>2800</td>
</tr>
<tr>
<td>Tse'an Kaetan Cave</td>
<td>Oreamnos dung</td>
<td>14,220±320 (A-2835)</td>
</tr>
<tr>
<td></td>
<td></td>
<td>30,600±1800 (A-2722)</td>
</tr>
<tr>
<td>Grobot Grotto</td>
<td>Mammuthus dung</td>
<td>28,920±2100 (Beta-14422)</td>
</tr>
<tr>
<td></td>
<td>Artiodactyl dung</td>
<td>20,930±400 (Beta-14420)</td>
</tr>
<tr>
<td>BF Alcove</td>
<td>Pseudotsuga and</td>
<td>11,790±190 (Beta-14727)</td>
</tr>
<tr>
<td></td>
<td>Oeer needles and</td>
<td></td>
</tr>
<tr>
<td></td>
<td>seeds in assoc. w/cf. Came~ops dung</td>
<td></td>
</tr>
</tbody>
</table>

ure 4 presents the absolute chronology derived from Holocene and Pleistocene alluvial deposits on the Colorado Plateau. Even cursory inspection of Figure 4 indicates that most of the dates are Holocene in age; in fact, 80% of these dates are late Holocene (<5,000 BP). Inspection of the Pleistocene alluvial chronology (+10,000 BP) indicates that data are sparse, as compared to the Holocene, but that a record does exist to near the limit of radiocarbon dating.

We have demonstrated that we have at least five dung deposits (including the newly discovered Hoopers Hollow deposit) that date to greater than 10,000 BP. All but one of these (Bechan Cave) is intimately related to older underlying alluvial sequences. Even the alluvium outside Bechan Cave, but in the local canyon system, has yielded absolute dates of 10,020±270 (Gx-11145), 32,800±1,300 (AA-1511), and >37,500±7,000 (Alpha-3034) yr BP. The alluvial chronologic framework will be supplemented by fecal chronology and the pack-rat midden analysis. The pack-rat chronology has also been extended (2x-3x) temporally beyond those data from the local study of Allen and Fishmouth caves (Betancourt, 1984).

Palynological data from the C-14-controlled alluvial sequence will be cross-correlated to the paleobotanical records supplied by the dung deposits and the pack-rat middens. The integration of all lines of evidence will provide a more comprehensive synthesis of paleoecologic change, tied to an absolute chronology.

PALEONTOLOGY

Bechan Cave

Osteological/Dental Remains. Osteological and dental remains were sparse in the test excavations of 1983 and 1984 in Bechan Cave. Only one Pleistocene species, Eueratherium collinum, was positively identified from dental material.

Dung Analysis. The majority of the Bechan Cave dung bed was derived from Mammuthus dung. The mass of the deposit is represented by trampled, mixed mammoth dung, although 17 complete or nearly complete boluses were recovered in the limited testing of the deposit (Figure 5). Incorporated within the mass of mammoth dung were the fecal pellets and boluses of additional species (Table 2).

Agenbroad and Mead
Mammuthus sp. - The size, shape, and bulk of the largest boluses plus the size-fraction analysis and composition of the material eliminated all but the proboscideans as the depositional agent from the Pleistocene megafauna. The composition (+95% grass, sedge, and rushes; Davis and others, 1985) serves to segregate mammoth-derived material from that of the mastodon. (The mammoth was adapted to grazing, whereas the mastodon was a browser). The paucity of mastodon material from the region as contrasted to mammoth remains, plus the dietary components of the dung and direct comparison with both zoo and wild African- elephant (Loxodonta africana) fecal material, allows us to identify Mammuthus as the major contributor to the deposit (Mead and others, 1986). This conclusion is substantiated by hair analysis.

Size fraction analyses of mammoth boluses from Bechan Cave and Loxodonida dung are strongly similar (Figure 6), as contrasted to analyses of horse and ground-sloth dung. A similar analysis serves to differentiate among bighorn sheep (Ovis canadensis), extinct mountain goat, extinct musk ox (Ovibos moschatus), elk (Cervus), and deer (Odocoileus) (Mead and others, 1986).

Bacatherium collinum - Analytic criteria applied to certain pellet-formed dung within the Bechan Cave deposit allowed us to identify a form of dung which is very similar to modern musk ox (Ovibos moschatus) winter pellets. The dental material, noted above, permitted a preliminary assignment to Bacatherium collinum.

Nothrotheriops shastensis - Shasta-ground-sloth dung boluses were recovered during the test excavations. Their coarser constituents and smaller fragments (or particles), plus sloth-dung-selective insect activity, did not allow the state of preservation shown in the mammoth boluses or animal pellets. We assign the material to Nothrotheriops shastensis, based on comparison with the Rampart Cave material.
TABLE 2
Fauna present at Bechan Cave, Utah; designations based on dung identification

<table>
<thead>
<tr>
<th>Species</th>
<th>Common Name</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Mammuthus</em> sp.</td>
<td>Mammoth</td>
</tr>
<tr>
<td>cf. <em>Euceratherium</em> sp.</td>
<td>mylodont sloth</td>
</tr>
<tr>
<td><em>Nothrotheriops</em></td>
<td>Shasta ground sloth</td>
</tr>
<tr>
<td><em>Oreamnos harringtoni</em></td>
<td>shrub ox</td>
</tr>
<tr>
<td>? sp.</td>
<td>horse</td>
</tr>
<tr>
<td><em>Equus</em> sp.</td>
<td>bighorn sheep</td>
</tr>
<tr>
<td><em>Bison</em> sp.</td>
<td>deer</td>
</tr>
<tr>
<td><em>Ovis</em> sp.</td>
<td>bear</td>
</tr>
<tr>
<td><em>Canis latrans</em></td>
<td>coyote</td>
</tr>
</tbody>
</table>

Hair Analysis. The extremely dry conditions within Bechan Cave promoted the preservation of pelage within the dung bed. Techniques for the identification of modern species of fur-bearing animals are known. Using similar techniques and reference specimens, Bolen has identified both modern and extinct animals from Bechan Cave (Table 3; Bolen, 1983).

Microfauna

<table>
<thead>
<tr>
<th>Species</th>
<th>Common Name</th>
</tr>
</thead>
<tbody>
<tr>
<td><em>Sorex</em> sp.</td>
<td>shrew</td>
</tr>
<tr>
<td>Neotoma sp.</td>
<td>pack rat</td>
</tr>
<tr>
<td><em>Myotis</em> cf. leibii</td>
<td>small-footed bat</td>
</tr>
<tr>
<td><em>Perognathus</em> cf. mantaiulus</td>
<td>deer mouse</td>
</tr>
<tr>
<td><em>Solinomys</em> cf. aberti</td>
<td>Abert's squirrel</td>
</tr>
<tr>
<td><em>Perognathus</em> cf. fasciatus</td>
<td>olive-backed mouse</td>
</tr>
</tbody>
</table>

The alluvial deposits in the canyon containing Bechan Cave are thick (20 m) and extensive. It was realized early in our research in Bechan Cave that the alluvial fill might contain paleoenvironmental and paleoecological information that could be correlated to the cave record at both pre- and postdung depositional intervals.

Alluvium is well exposed due to the recent (<100 yrs) erosion cycle, which has incised the entire alluvial sequence to the bedrock floor of the canyon. These exposures of the alluvium reveal multiple organic bands (presumed to represent soil horizons), as well as datable materials at numerous localities.

Alluvial study in these canyons began in the latter half of 1984 and is ongoing (Agenbroad and Elder, 1985; 1986). A sequence of five terraces (T4-T0) was recognized and mapped (Figure 7). The terrace profiles produced a sequence of alluvial events that could be tied to a radiocarbon-controlled chronology. From these data, a preliminary

geologic-paleohydrologic model was constructed for the canyon alluvium (Figure 8). At present, we have 22 radiocarbon dates and 1 thermoluminescence (TL) date (Table 4) from the alluvial and eolian deposits, which provide the chronologic framework for the alluvial events. Of those, five dates are from buried archaeological sites extending back to 5300 BP.

Eolian deposits were probably accumulating during most of the depositional history of the canyon. Remaining climbing and falling dunes, currently stabilized, attest to periods of more rapid accumulation than attested to by the small active dunes. Dates on charcoal from within the dunes and from dune modification attest to the antiquity of these deposits. There is geomorphic evidence that the entrance to Bechan Cave, during the use of the cave by the prehistoric fauna, was accessible via a climbing dune that has been partially destroyed by runoff over the canyon wall since that time.

Based on the currently available dates from the alluvium, the oldest unit in the canyon is T4, as indicated by a TL date of >37,500 to <54,300 BP (Alpha-3034). This older alluvium is preserved as remnants beneath talus blocks, or as high terrace remnants against portions of the canyon wall. Repeated searching of the exposures failed to reveal material suitable for standard radiocarbon dating. Samples of a small bivalve, *Platidium*, from the upper portion of T4 produced a TAMS (tandem accelerator mass spectrometer) radiocarbon age of 32,000 ±1300 BP (AA-1511). This TAMS date refines the age of T4.
Quaternary Alluvial Geology
Adjacent to
Bechan Cave

Figure 7. Quaternary alluvial geology adjacent to Bechan Cave.

analyzed on TL and may well displace the later date.

Terrace 4 filled the canyon floor to as much as 20 m above the bedrock. This aggradation interval provides evidence of cyclic deposits indicative of high-energy flow (sand) separated by low-energy deposits (silt and clays). Sometimes after the aggradation of T4, erosional events initiated degradation at ±11,000 BP causing most of the T4 sediments to be removed from the canyon during an arroyo-cutting episode. Between 11,000 and 10,500 BP an aggradation cycle began, filling the canyon with alluvium again (T3a). Terrace 3 aggraded to, or nearly to, the level of T4 and was then subjected to an erosion interval between 9000 and 6000 BP. Aggradation of this arroyo system created T3b which filled to ±15 m above the canyon floor and comprises most of the alluvial terraces exposed in the canyon today (Figure 8). Between 6000 and 4500 BP, T3 was eroded to the bedrock of the canyon floor. By approximately 4500 BP this episode reversed, causing deposition of T2, which appears to contain at least three units, one of which (T2b) is older than 4000 BP, but younger than ±4500 BP. T2b represents a cut-and-fill sequence that is younger than 4000 BP, but older than 900 BP. At approximately 900 BP, T2a was eroded, to be filled before 300 BP. Based on the radiocarbon dates available, T2 and other units were subjected to erosion at 100 to 75 BP, judging from the age of the oak (Quercus) tree stand that covered T2 and T3 surfaces prior to that time. As the erosional interval continued, the water table was lowered, causing the death of the oak stand. This erosion interval again cut to the bedrock floor of the canyon and began deposition of T1 between 100 and 50 BP. Erosion of T1 began at approximately 50 BP, leaving low T1 terraces and reaching the bedrock floor of the canyon. Bedrock comprises most of the stream floor at the present time.

Soils

Soils preserved in the alluvium of the canyons in the vicinity of Bechan Cave have been studied by Anderson and Karlstrom (1986) and Anderson (1985, 1987). Thirteen soil profiles have provided 198 samples from in and on the alluvial terraces. Laboratory analyses for particle size, percentage of calcium carbonate, percentage of organics, percentage of iron, manganese, and aluminum, pH, and clay mineralogy were utilized to determine the source of the material, stage of soil development, amount of soil zonation, and paleoenvironmental conditions of soil formation.

Results of the analyses indicate that all the soils in the canyon alluvium can be assigned to two soil orders, Entisols and Mollisols. Entisols are incipient soils with simple A/C horizons. They form the majority of soils in the canyon and are common in arid regions with perennial or ephemeral water sources. The incipient A horizons are repeatedly buried by accretion of the flood plain. The Entisols in the vicinity of Bechan Cave did not develop into more advanced soils because of rapid burial, a
Figure 8. Schematic cross section of alluvial fill(s) in the canyon adjacent to Bechan Cave.

<table>
<thead>
<tr>
<th>Date (BP)</th>
<th>Lab No.</th>
<th>Material Dated</th>
<th>Strat Position</th>
<th>Field No.</th>
</tr>
</thead>
<tbody>
<tr>
<td>0 ±75</td>
<td>Gx-10759</td>
<td>oak log</td>
<td>Top T2</td>
<td>85-2</td>
</tr>
<tr>
<td>215 ±80</td>
<td>Gx-11144</td>
<td>root</td>
<td>T2(c)</td>
<td>85-10</td>
</tr>
<tr>
<td>940 ±200</td>
<td>Gx-11431</td>
<td>charcoal</td>
<td>T2(b)</td>
<td>85-9</td>
</tr>
<tr>
<td>950 ±160</td>
<td>Gx-11018</td>
<td>charcoal (hearth?)</td>
<td>T2(b)</td>
<td>85-8</td>
</tr>
<tr>
<td>1290 ±75</td>
<td>Gx-11338</td>
<td>charcoal (hearth)</td>
<td>dune on T3</td>
<td>85-15</td>
</tr>
<tr>
<td>1315 ±75</td>
<td>Gx-11340</td>
<td>charcoal (hearth)</td>
<td>T2(b)</td>
<td>85-25</td>
</tr>
<tr>
<td>1525 ±130</td>
<td>Gx-11336</td>
<td>charcoal (hearth)</td>
<td>dune on T2</td>
<td>85-13</td>
</tr>
<tr>
<td>1600 ±60</td>
<td>Beta-15640</td>
<td>charcoal</td>
<td>T2(b)</td>
<td>86-1</td>
</tr>
<tr>
<td>1750 ±90</td>
<td>Beta-16587</td>
<td>charcoal</td>
<td>T2(b)</td>
<td>86-3</td>
</tr>
<tr>
<td>2510 ±80</td>
<td>Gx-11562</td>
<td>wood</td>
<td>T2(b)</td>
<td>85-19</td>
</tr>
<tr>
<td>2550 ±80</td>
<td>Gx-11561</td>
<td>wood</td>
<td>T2(b)</td>
<td>85-18</td>
</tr>
<tr>
<td>3000 ±145</td>
<td>Gx-11339</td>
<td>charcoal (hearth)</td>
<td>T2(b)</td>
<td>85-16</td>
</tr>
<tr>
<td>3750 ±165</td>
<td>Gx-11563</td>
<td>charcoal</td>
<td>T2(b)</td>
<td>85-20</td>
</tr>
<tr>
<td>4025 ±95</td>
<td>Gx-10939</td>
<td>root</td>
<td>T2(a)</td>
<td>85-4</td>
</tr>
<tr>
<td>4090 ±155</td>
<td>Gx-10758</td>
<td>charcoal &amp; plant debris</td>
<td>T2(a)</td>
<td>85-1</td>
</tr>
<tr>
<td>4120 ±170</td>
<td>Gx-11564</td>
<td>charcoal</td>
<td>T2(a)</td>
<td>85-21</td>
</tr>
<tr>
<td>4165 ±95</td>
<td>Gx-11019</td>
<td>wood</td>
<td>T2(a)</td>
<td>85-3</td>
</tr>
<tr>
<td>4260 ±170</td>
<td>Gx-11337</td>
<td>root</td>
<td>T2(a)</td>
<td>85-14</td>
</tr>
<tr>
<td>5300 ±235</td>
<td>Gx-11146</td>
<td>charcoal (hearth)</td>
<td>dune on T3</td>
<td>85-12</td>
</tr>
<tr>
<td>10,020 ±270</td>
<td>Gx-11145</td>
<td>charcoal</td>
<td>T3(a)</td>
<td>85-11</td>
</tr>
<tr>
<td>32,800 ±1300</td>
<td>AA-1511</td>
<td>TAMS on Pisidium</td>
<td>T4</td>
<td>85-6-5</td>
</tr>
<tr>
<td>&gt;37,500 ±7000</td>
<td>Alpha-3034</td>
<td>TL date on sediment</td>
<td>T4</td>
<td>86-4</td>
</tr>
<tr>
<td>&lt;54,300 ±10500</td>
<td>Alpha-3034</td>
<td>TL date on sediment</td>
<td>T4</td>
<td>86-4</td>
</tr>
</tbody>
</table>
parent (source) material that is nearly pure quartz sand, small amounts of organic material, and lack of water. Minor Mollisols, reflecting higher amounts of moisture with concomitant vegetational response provides more organic material. This prairie or grassland soil type is found in the desert Southwest in areas of higher soil moisture, as in riparian habitats.

Both soil types are present in the modern environment within the canyon. Periods of 50 to 100 years of stability could produce the modern and ancient soils. Rapid aggradation would bury such soils, thus providing the multiple organic horizons observed in the canyon alluvium.

The oldest soil sampled is dated at just over 4000 BP. The ±37,000-year-old sediments of the T 4 remnants did not yield geosol information. On the basis of soil formation, soil chemistry, and mineralogy, it is inferred that the past 4000 years of soil-forming conditions have changed little; they reflect conditions similar to the present, with intervals of wetter and drier conditions, as evidenced by the cut-and-fill cycles and terrace formation.

One soil in lower Bechan Canyon was identified as the "altithermal" soil during initial field inspection. The laboratory analysis combined with a TAMS (tandem accelerator mass spectrometer) C-14 date of 1750 ±90 BP (Beta-16587), indicates that this soil reflects a larger influx of Kayenta Formation parent material, as determined by the clay mineralogy and the low percentages of iron, manganese, and aluminum present in the soil.

ALLUVIAL POLLEN

Palynological analysis of the alluvial fill of the canyons has been undertaken by R. Hevly and S. ClayPoole (1985, 1986). The results presented here are preliminary, and analysis is still underway.

Terrace 4, dating to greater than 37,000 BP, yielded pollen indicative of a plant community that included ponderosa pine, birch, juniper, ChenoAms, low-spine Asteraceae (including Artemisia), and grasses. The older portion (+10,000 BP) of T 5 contains only grass pollen. These samples support the botanical record derived from the Bechan Cave dung bed, which brackets an interval of 1670-13,505 BP. Early Holocene portions of T 3 , as dated by interpolation of sedimentation rates (Hevly and ClayPoole, 1986), reflect an environment in which ponderosa pine, Artemisia, grass, ChenoAms, low- and high-spine Asteraceae, and Ephedra were common, but which also contained some brief episodes of spruce, white fir, Douglas fir, and juniper. Upper T 3 and T 2 sediments, dating 2500-4100 BP, indicate decreased proportions of Chenopod pollen, increasing grass pollen, and very little arboreal pollen.

Alluvial and cave pollen records indicate that the late Holocene pollen profiles demonstrate paleoenvironmental conditions similar to the present.

Canyon Alcoves

Deposits. The canyon deposits consist of a series of alcoves within the entrenched meanders of the canyon (Figure 9) that have preserved alluvial sediments capped by Pleistocene dung deposits and associated macrobotanical remains. Two dung samples, one of the deposits dated 20,930 and 28,290 BP (Table 1), Douglas-fir needles and maple seeds associated with possible camel dung from another alcove date 11,790 BP.

Paleontology. The canyon-alcove deposits contain osteological, dental, and fecal remains. It is expected that hair will also be recovered, as excavations proceed. Thus far in our field investigations, identifiable osteological remains have been recovered only from Hoopers Hollow. Grobot Grotto has osteological material that will require stabilization prior to removal and identification. All the alcoves contain dung deposits. Mammoth and bison dung boluses have already been recovered from the alcove dung mats, as have the large pellets assigned tentatively to camel and the smaller pellets assigned to the shrub ox and Harrington's mountain goat. A series of controlled sampling excavations is underway in the alcoves. The faunal and botanical deposits overlie up to 23 m (75 ft) of stratified alluvium, preserved as remnants where sheltered by alcove overhangs, at up to 73 m (238 ft) above modern base level (Figure 10).

Alluvium. The five alcoves containing the elevated ancient sediments are being studied and compared to younger alluvial units within and upstream from the alcoves. The hydrologic-geologic history of the drainage is being interpreted from the evidence remaining in the deposits. It has become apparent that a younger, lower-elevation sequence of alluvial units is also preserved within the canyon. In alluvial studies on the Colorado Plateau, as discussed previously in this paper, post-10,000 BP data are sparse. The canyon alcoves provide multiple exposure of a thick sequence of alluvium that has to predate the dung beds that cap it. The potential for retrieval of paleoenvironmental data from these deposits is greater than for other deposits from the Plateau.

DUNG, DIET, AND EXTINCTIONS

An estimated 300 m³ of Pleistocene dung forms the dung blanket at Bechan Cave (Davis and others, 1984). The majority of this dung was deposited by mammoth. Analysis of the mammoth dung (Davis and others, 1984, 1985; Mead and others, 1986) reveals that more than 95% of the mammoth boluses are composed of a graminoid matrix. This reflects the analysis of 25 fragments of mammoth dung from complete, or nearly complete, boluses collected in 1984 test excavations.

The graminoid matrix is composed of grasses, sedges, and rushes indicative of riparian plants. Nonmatrix sections of the dung include saltbush, cactus, and sagebrush. Plant macrofossils from the dung bed include birch, rose, saltbush, sagebrush, blue spruce, wolfberry, and red osier dogwood (Davis and others, 1985).

The paleoenvironment at Bechan Cave has been interpreted as a sagebrush-steppe upland with a riparian community near the streams. The Bechan Cave botanical community can be nearly duplicated in the nearby uplands of the Henry Mountains, or the Aquarius Plateau, 1200 m
(±4000 ft) higher in elevation than the cave. The upward migration (elevation) of the plant community can be roughly correlated with warming and seasonal precipitation changes near the cave during the past 11,000 years.

TABLE 5
Fauna from the Canyon Alcoves

<table>
<thead>
<tr>
<th>Osteological/Dental Remains:</th>
<th>Fauna</th>
</tr>
</thead>
<tbody>
<tr>
<td>Locality</td>
<td></td>
</tr>
<tr>
<td>Hoopers Hollow</td>
<td>Metapodial Oreamnos harringtoni</td>
</tr>
<tr>
<td>Hoopers Hollow</td>
<td>Tooth Equus</td>
</tr>
<tr>
<td>Hoopers Hollow</td>
<td>Mandible Marmota</td>
</tr>
<tr>
<td>Fecal Remains:</td>
<td></td>
</tr>
<tr>
<td>Grobot Grotto</td>
<td>Bolus Mammuthus</td>
</tr>
<tr>
<td></td>
<td>Bolus Oreamnos</td>
</tr>
<tr>
<td></td>
<td>Pellet cf. Baeoaeratherium</td>
</tr>
<tr>
<td>BF Alcove</td>
<td>Pellet cf. Camelops</td>
</tr>
<tr>
<td>Hoopers Hollow</td>
<td>Pellet Oreamnos</td>
</tr>
<tr>
<td></td>
<td>Pellet cf. Baeoaeratherium</td>
</tr>
</tbody>
</table>

Comparison of the alluvial record from the canyons near Bechan Cave, the cave records from Bechan Cave and Cowboy Cave, an extended botanical record from pack-rat middens of Natural Bridges National Monument (Mead and others, in press), and a developing chronology from nearby drainages has led to the development of a paleoenvironmental model tied to a C-14 chronologic sequence. Bechan Cave provides a large repository of mammoth dung dating just 670 years prior to the extinction. Grobot Grotto has supplied mammoth dung dating 28,290 yr BP, which allows us to approach the extinction event from "the far side," with data from dietary samples dating prior to extinction, rather than work back in time from the Holocene record. If the chemical analyses fail or prove to be nondiagnostic, we will have from three repositories within the central Colorado Plateau the botanical record from dung deposits and pack-rat middens, as well as pollen analyses, for cross-correlation of paleobotanical records.

ACKNOWLEDGMENTS

Numerous persons and agencies have provided funds and accumulated data. The National Park Service, Glen Canyon National Recreation Area, has been most cooperative and has provided the majority of funding for our research efforts. Initial logistical support was followed by a multiyear contract (NPS: CX-1200-4-A062) to study the Quaternary deposits of Glen Canyon National Recreation Area and Canyonlands National Park. Natural Bridges National Monument has provided assistance as well. The National Geographic Society has provided funding for stages of our research under three grants: 2711-83; 3016-85; and 3280-86. National Geographic Society support provided data and details not called for in the NPS contract.

The NPS administrative staff of Glen Canyon National Recreation Area, Canyonlands National Park, and Natural Bridges National Monument have been enthusiastic supporters of our research. A list of individuals by name would be lengthy, indeed. The Rocky Mountain Region office (NPS) in Denver, Colorado has been instrumental in support of our efforts. In particular, A. Anderson, A. Johnson, and K. Warner have provided outstanding support.

Northern Arizona University has provided numerous salary, logistic, and equipment requests. Colleagues and research assistants in-
clude K. Anderson, S. ClayPoole, D. Elder, R. Hevly, E. Mead, D. Meir, and R. Ryan. P. Martin and O. Davis of the University of Arizona were coresearchers of the Becham Cave deposits.

FIELD TRIP

Regional Geology

The Glen Canyon National Recreation Area is situated in a picturesque setting that originates from diverse geologic exposures. The oldest units exposed during our up-lake field excursion will be the Triassic Moenkopi Formation, the Shinarump Conglomerate, and the Chinle Formation (Figure 11). Much of the close-at-hand geology for this trip will be the Triassic-Jurassic Glen Canyon Group. This sequence of sandstones accounts for most of the spectacular canyon-wall scenery and houses the caves and alcoves that have preserved the dung beds and alluvial deposits that are the object of this field examination.

The Glen Canyon Group is composed of, in ascending order, the Wingate Sandstone, the Kayenta Formation (siltstone sandstone), and the Navajo Sandstone. The Wingate and Navajo Sandstones are easily recognized as prominent cliff-forming units. The Kayenta Formation is usually a sloping shelf that separates the sandstones. The Triassic-Jurassic boundary is thought to occur in the upper Kayenta Formation (Hintze, 1975). Most of our travel by boat and on foot will be within the Kayenta and Navajo exposures.

The Kaiparowits Plateau provides the upland topography north of the river and provides distant exposures of Jurassic, Cretaceous, and Cenozoic-Tertiary formations (Figures 12, 13, 14, and 15). To the south of the river, Navajo Mountain, a laccolithic intrusion, dominates the topography and causes upwarping of the older deposits (Figure 15).

Modern Climate and Vegetation

Elevation varies greatly within the area of this field trip, from a low of 1348 m (4400 ft) at Page to a high of 3166 m (10,300 ft) at Navajo Mountain. Precipitation ranges from 35 cm in the highlands to 14 cm at the lower elevations.

The vegetation near Becham Cave occupies a variety of habitat types that range from bare walls and ridge tops of Navajo Sandstone, vegetated only along cracks or in pockets, to rich riparian growth along a semipermanent stream. The region is also noted for its hanging "gardens" (Welsh and Taft, 1981), face and foot wall gardens, and plunge basins fed by seepage that may harbor primrose (Primula eapaula), maidenhair fern (Adiantum), rock spirea (Petrosphygium), cardinal flower (lobelia), death camas (Zigadenus), orchids (Epipacta), monkey flowers (Mimulus), columbia (Aquilegia), and other aquatic or mesic species that are unusual in this arid region. Above the canyon bottoms waterpockets (deep plunge pools) that support willows (Salix etigma) maintain a permanent water supply during the summer season. The riparian habitat near Becham Cave includes Fremont cottonwood (Populus fremontii), birch-leaf buckthorn (Rhamna betulifolia), sedge (Carex spp.), poison ivy (Toxicodendron radicans), and other aquatic or mesic species. Single-leaf ash (Fraxinus anomala) occupies cracks in the bedrock.

Scattered trees along the canyon bottom include Utah junipers (Juniperus osteosperma), Gambel oak (Quercus gambelii), and Utah service berry (Amelanchier utahensis). Shrubs include black brush (Coleogyne ramosissima), Mormon tea (Ephedra spp.), four-winged saltbush (Atriplex canescens), spiny hop sage (Coelogyna spinosa), and prickly pear (Opuntia phaeacantha). Common grasses include Indian rice grass (Oryzopsis hymenoides), needle grass (Stipa comata), and Sacaton (Sporobolus sp.). Above the canyon and away from bare cliffs, the vegetation is dominated by Coleogyne ramosissima, Juniperus osteosperma, and Oryzopsis hymenoides. (Davis and others, 1984).

FIELD-TRIP LOG

This field excursion will be somewhat different than the "normal" field trip, in that nearly 300 miles of highway travel is necessary just to reach the point of assembly for the field portion of the trip. The trip will, of necessity, require travel by boat (and hikng) for approximately 200 additional miles. The trip from Phoenix/Tempel will provide a southward cross section of the state of Arizona; from the Basin and Range Province through the Transition Zone (Mogollon Rim) to the Colorado Plateau.

Rather than attempt a road log for the highway portion of the trip, the authors refer the participants to the roadside geology logs for Interstate 17 and Highway 89 prepared by Chronic (1983). The logs for Interstate 17 from...
Phoenix to Camp Verde (p. 143-146), Camp Verde to Flagstaff (p. 191-193), Flagstaff to Cameron (p. 217-222), and Cameron to Page (p. 223-228) provide good descriptions of the route to the point of assembly. The remainder of the excursion is our field trip, to be accompanied by a boat log for travel on Lake Powell.

All of the formal stops are located within the Glen Canyon National Recreation Area, which is administered by the National Park Service. A collecting permit is required for the entire area to be visited. In compliance with a National Park Service request, the exact locations of the alcoves and Bechan Cave are withheld from this publication.

We will visit the Bechan Cave deposits with the associated alluvium during the first field day. On the second day, we will visit the alcove deposits. Our interpretations and model of the paleoenvironment and its chronology of change will be presented. We hope to initiate a discussion about the significance of the deposits with respect to the generation of a local, chronologically controlled, paleoenvironmental reconstruction. These interpretations and descriptions will be related to a more regional perspective, as known from published data, and the research that is currently being conducted by the field-trip leaders and their colleagues.

Figure 12. Landsat image of field-trip area.

Figure 13. Sketch map of Landsat imagery showing physiographic features.

FIRST DAY

Highway travel from Phoenix/Tempe to the point of assembly at the group camping area of the National Park Service campground at Wahweap (5 miles north of Glen Canyon Dam). Steak fry and overnight camp. Welcoming comments by the National Park Service.

SECOND DAY

Departure 0700 hrs. from the Wahweap boat ramp. Each participant will be provided with a Glen Canyon National Recreation Area map. Be aware that the up-lake mileage, as depicted on the map, begins at Glen Canyon Dam. Mileage buoys will be encountered as we proceed up-lake.

We will travel between canyon walls of Navajo Sandstone of Jurassic age. The Navajo Sandstone is world famous for its large-scale crossbeds, indicative of eolian origin. The paleowind direction has been determined to have been from the north or northwest (Hintze, 1975). The Navajo Indian Reservation begins at the south shoreline.

Mile 42: Fuel stop at Dangling Rope marina. The north wall provides exposures of the Kaiparowits Plateau, dominated by a Jurassic through Cretaceous to Cretaceous/Tertiary se-
sequence. Navajo Point is the prominent southern extension of the Kaiparowits Plateau. The south wall of the canyon is capped by Cummings Mesa and the spectacular upwarping of the Glen Canyon Group is displayed as we approach Navajo Mountain. Navajo Mountain is one of several mid-Tertiary laccolithic intrusions located in southeast Utah.

Figure 14. Generalized geologic map of field-trip area. P = Permian Cedar Mesa SS; R = Triassic Moenkopi Fm to Kayenta Fm; J = Navajo SS to Morrison Fm; Kj = Dakota SS; Ku = Mancos Sh-Mesa Verde SS-Kaiparowits Fm; T = Tertiary intrusives. A-A' and B-B' = lines of cross sections.

Mile 49: We will make a short side trip to Rainbow Bridge National Monument. As we continue up-lake, rounded gravel deposits are visible capping islands in the lake and as terraces along the shoreline. These gravels are collectively known as the San Juan River Gravels. Their constituent lithology reflects source areas in southwestern Colorado and transport of these terrace deposits by the ancestral San Juan River.

Figure 15. Schematic cross section A-A'; vertical exaggeration = 10x.

Mile 53: The confluence of the San Juan River arm of Lake Powell. Up-lake from the confluence the gravel-capped terraces are derived solely from the ancestral Colorado River.

Mile 64: This is the location of "Hole-in-the-Rock," where a wagon train of Mormons en route to Bluff, Utah descended the canyon wall. Cottonwood Canyon (on the east shore) was the ascent route from the river. The riprap road bed is still visible a short walk up the canyon from lake level.

Mile 68.5: The confluence of the Escalante River arm with the Colorado River arm of Lake Powell. The Straight Cliffs escarpment (NE) of the Kaiparowits Plateau dominates the southwest skyline.

Miles 77-78: An abandoned meander bend, "The Rincon" is visible on the south shore. Within the Rincon, the Kayenta Formation, Wingate Sandstone, Chinle Formation, and the Moenkopi Formation are exposed. On the north shore, the Glen Canyon Group is upwarped by the Waterpocket Fold, an anticlinal structure that continues northward to the Capitol Reef National Monument, a distance of ±150 mi. The Henry Mountains, a laccolithic mass to the north, are visible from several locations on the river and from the walls of The Rincon.

Mile 96: Halls Crossing marina and Bullfrog Bay. At this point, a commercial ferryboat can take vehicles from the east bank to the west bank of the lake across Bullfrog Bay.

Figure 16. Schematic cross section B-B'; vertical exaggeration = 30x.
Stop 1

(Hike up the canyon toward Bechan Cave.)

Stop 1(a): Examination of the soil profiles exposed in the alluvial exposures such as Figure 17. Several profiles will be examined and discussed as we proceed up the canyon.

Stop 1(b): Bechan Cave. Here we will describe the dung deposit, the fauna that excreted it, and the dung, hair, osteological, and dental analyses of it. Approximately 5% of the known deposit was disturbed by scientific sampling. A brief discussion of the archaeological remains encountered in the sample excavations will be presented, as will the paleobotanical evidence derived from the dung bed. This model can be dramatically contrasted with the present vegetation of the canyon. The change has taken place within the past 11,000 years.

Stop 1(c): After leaving the cave, we will hike up-canyon to view several alluvial and eolian exposures whose chronology, palynology, etc., will be discussed in a reconstruction of the alluvial events. The terrace system will be studied, as will the postdepositional modification of some of the eolian deposits. Cultural sites within the alluvial units will be described, as will their relationship to the alluvial processes, and a cross-correlation of the alluvial and postdung cultural activity in the cave will be presented.

Return to the boats and camp.

THIRD DAY

Stop 2

After a relatively short boat trip, we will set up camp for the second night and then proceed to the "jumping-off-place" for the hike to the canyon alcoves.

Stop 2(a): Hoopers Hollow alcove. A scramble up the slick rock slope of the inner canyon will bring us to the ancient alluvial deposits and their dung deposits. Discussion of the fecal, dental, and osteological evidence will take place here. The botanical evidence and its significance will be presented.

Figure 18. Stratified alluvium in BF Alcove; arrow locates fossil dung deposit.

Figure 17. Exposed alluvium in canyon adjacent to Bechan Cave.

Figure 19. Stratified alluvium in Grobot Grotto; dashed line indicates basal limits of dung deposit; arrow indicates field personnel for scale.
Stop 2(b): BF Alcove (Figure 18). An examination of the thick sequence of alluvium, its age, and its geologic-hydrologic significance will be addressed.

Stop 2(c): Grobot Grotto (Figure 19): An examination of the sedimentary sequence, the chronology, and the facal and osteological aspects of the deposit will be presented. A discussion of the causal relationship among the topography, geomorphology, and depositional and erosional histories will be discussed. Evidence for a second, much younger cycle of deposition and erosion will also be addressed.

Return to camp.

FOURTH DAY

Return by boat to Wahweap. Drive to Phoenix/Tempe.

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Late Cenozoic Volcanism in the San Francisco and Mormon Volcanic Fields, Southern Colorado Plateau, Arizona

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INTRODUCTION

The objective of this trip is to study the volcanic history, volcanology, petrology, and structural geology of two late Cenozoic volcanic fields in north-central Arizona. In the northern and eastern San Francisco field, late Miocene to Holocene basaltic flows, cones, and maars are well displayed, and their relationships to tectonic structures and erosion surfaces are observable. Silicic and andesitic lavas and pyroclastic deposits of the San Francisco Mountain composite volcano and five peripheral silicic centers occur in the central part of the field. The Mormon field is composed of late Miocene to Pliocene(?) basaltic sheet lavas and shields, scoria cones and small associated flows, and scattered silicic and andesitic domes, in addition to the major silicic centers of Mormon Mountain and Hackberry Mountain. Older basaltic lavas of the field accumulated against an ancestral Colorado Plateau marginal escarpment, but younger lavas cascaded over the escarpment and formed a westward-sloping ramp from the Plateau into the Transition Zone. Unless referenced otherwise, K-Ar ages are by P.E. Damon (written comm., 1977) and E.H. Mckee (written comm., 1973).

LATE CENOZOIC VOLCANISM ON THE SOUTHERN COLORADO PLATEAU

Geologic provinces in Arizona include the Colorado Plateau in the north and the Basin and Range in the south (Figure 1). Between these provinces is a structurally intermediate region known as the Transition Zone; in most places its boundary with the Colorado Plateau is paralleled by the Mogollon Rim (Escarpment; Peirce, 1985).

Late Cenozoic volcanism (younger than 16 Ma) was widely distributed in Arizona, but the eruptive products are most voluminous in volcanic fields situated on the southern margin of the Colorado Plateau and in the adjacent Transition Zone. The major volcanic fields, which are dominated by basalt, form a zone that trends southeast across the state from its northwest corner (Luedke and Smith, 1978). The San Francisco and Mormon volcanic fields are in the middle of this zone; both fields are predominantly on the southern Colorado Plateau margin, but they extend across the Mogollon Rim into the Transition Zone.

SAN FRANCISCO VOLCANIC FIELD

Late Miocene to Holocene lava flows and pyroclastic deposits of the San Francisco volcanic field cover more than 5000 km² of the southern Colorado Plateau and Transition Zone (Figure 1). About 500 km³ of volcanic rocks were deposited on erosion surfaces above nearly horizontal strata (Wolfe and others, 1983). In general, the oldest lavas were extruded on the highest erosion surfaces, which in many places were developed on the Triassic Moenkopi Formation, whereas younger lavas rest on the erosionally stripped Kaibab Formation of Permian age or on the older lavas (Cooley, 1962; Ulrich and others, 1984).

Major structural features in the San Francisco field are high-angle faults and monoclines (Figure 1). Faults of small displacements are dominantly on northwest trends, but north- and northeast-trending faults occur. Although normal faulting and volcanism in the field are broadly coeval, most of the faults displace the older lavas and only a few lavas and deposits of Brunhes age (<0.73 Ma) are faulted (Wolfe and others, 1983; Tanaka and others, 1986). The major silicic centers and many of the scoria cones are aligned or elongated on trends subparallel to the principal strike lines of the faults. In the San Francisco Mountain volcanic system, several of the major vents, the conduit system of the central volcano, and a large valley that breaches its northeast side are collinear on a northeast trend parallel with the three major silicic centers in the western part of the field (Figure 1).

Basaltic lava flows and tephra deposits dominate the eruptive products in the volcanic field and constitute most of the nearly 600 volcanoes that have been identified. The less abundant intermediate to silicic rocks occur in, or peripheral to, a few major volcanic centers (Robinson, 1913; Figure 1). The basaltic volcanism was broadly contemporaneous with the development of the intermediate to silicic centers, and K-Ar ages reveal a general progression of volcanism from the Transition Zone, beginning more than 14 Ma ago, to the central and eastern parts of the San Francisco field by latest Miocene to Holocene time (Luedke and Smith, 1978).

Mafic rocks of the San Francisco field are dominantly alkali basalts, but compositions range from strongly undersaturated basanitoids to olivine tholeiites, and hawaiitic to mugear-
Figure 1. Map showing field-trip route and distribution of late Cenozoic volcanic rocks in the San Francisco (SF) and Mormon (M) volcanic fields; BH, Black Hills. Volcanic geology modified from Luedke and Smith (1978) and Lewis (1983); structure from Ulrich and others (1984). AM, Anderson Mesa; BH, Bill Williams Mountain; CR, Colorado River; LCR, Little Colorado River; E, Eldon Mountain; HM, Hackberry Mountain; K, Kendrick Peak; MM, Mormon Mountain; O'L, O'Leary Peak; S, Sitgreaves Mountain; SFM, San Francisco Mountain; WCC, West Clear Creek. Dashed line shows approximate location of Mogollon Rim. Cross section X-X' across Eldon Mountain is shown in Figure 8. K-Ar ages (Ma) from Wolfe and others (1983), Luedke and Smith (1978), and McKee and Elston (1980).
ian varieties occur. Volcanic rocks of the San Francisco Mountain composite volcano and five smaller silicic centers peripheral to it constitute a coherent and continuous suite of lithologies, from low-silica andesite to alkali rhyolite (comendite), that forms a compositional continuum with the mafic rocks of the field (Figure 2).

In the central and eastern parts of the San Francisco field the lavas and cones have been classified in five age groups primarily on the basis of stratigraphic and physiographic relationships, degree of weathering and erosion, and K-Ar ages (Moore, 1974). These age groups are: Cedar Ranch, 6.5-0 Ma; Woodhouse, 3.0-0.8 Ma; Tappan, 0.7-0.2 Ma; Merrim, <150,000 years; Sunset, 1064 A.D.

The silicic rocks, however, form two distinctive suites, a northern suite at and near Mormon Mountain, and a southern suite at the Hackberry Mountain center and West Clear Creek (Figures 1 and 2). The northern suite is markedly lower in K2O than the San Francisco Mountain suite, and this is reflected in the sparse occurrence of biotite in the former. Silicic rocks of the southern suite are intermediate in K2O content between the Mormon Mountain and San Francisco Mountain suites and commonly contain biotite.

DESCRIPTION OF FIELD TRIP

First Day: Stop 1.

Cherry Street, Flagstaff (30 minutes)

A small cut immediately north of the Coconino County Courthouse exposes a meso- capping, late Miocene (5.82+0.34 Ma) basalt lava flow that rests unconformably on Lower Triassic redbed sandstone of the Moenkopi Formation. The basalt is in the oldest age group (Cedar Ranch) in the San Francisco volcanic field. The purpose of this stop is to observe the unconformity and the structural and petrographic features of the lava flow. A structureless regolith about 1 m thick separates the sandstone and the basalt; this weathered surface may be correlative with the region Zuni erosion surface (Cooley, 1962). The top 20 cm of the regolith has been baked red by the lava, which displays a smooth, generally breccia-free bottom. Spherical, smooth-wall vesicles are concentrated in the lower part of the lava; above this zone, large vesicles are concentrated in vesicle cylinders. The olivine tholeiite basalt carries phenocrysts of olivine (Fo81) and sparse clinopyroxene (Moore, 1974).

The purpose of this stop is to examine the structure and xenoliths of Crater 160, a composite cinder, tuff, and spatter cone of Tappan age (Figure 3). Its growth began with the buildup of welded basalt spatter and rootless flows, which form the layers exposed in the crater's wall. A dikelike body of spatter fragments in the northeastern wall was the source of the lava flow to the north. A phreatomagmatic event late in the history deepened the crater, widened the rim, and deposited a mantle of palagonitic tuff that contains an abundance of xenoliths for which the locality is noted. The most abundant xenoliths display cumulus textures and include clinopyroxene and wehrlite, but websterite, gabbro, anorthosite, granite, granulite, and Paleozoic sedimentary rocks are also present (Cummings, 1972; Stoesser, 1973). The last events in the cone's history included a series of fire- fountain eruptions, ending with the 35-m-high red cinder cone on the floor and scattered bomb fragments around the rim. The floor of the crater is about 80 m lower than the average surface outside.
Figure 3. High-altitude (U-2) oblique aerial photograph by U.S. Air Force of the San Francisco volcanic field. View is to southwest.

Stop 3. SP Cone and Flow (20 minutes)

The purpose of this stop is to see the cone and lava flow of SP Mountain, which are examples of the basaltic andesites of the San Francisco volcanic field. The cone's sharp-rimmed profile, radial symmetry, and steep flanks mark it as the youngest volcano in the northern part of the field (Figure 3); its age was determined as 71,000±4,000 yrs (Baksi, 1974; revised for new constants) and it is typical of the Merriam age group. The cone is 250 m high and 1200 m in diameter at its base; its summit crater is 400 m across and about 120 m deep. The slopes of the cone are covered with lapilli and bombs; ash is minor. Welded spatter forms a ruff around the crater's rim. The blocky lava flow, 15 m thick in this location, extruded early in the vent's history and moved down a multiple graben for 7 km; it is 55 m thick near its terminus. Spatter from the cone contains phenocrysts of clinopyroxene, olivine, and embayed and sieved plagioclase in a hypocrystalline groundmass. The flow is similar but contains, in addition, sparse orthopyroxene and embayed quartz. The base of the cone overlies the lava flow and is interpreted to be younger because it is not deformed by the extrusion (Hodges, 1962).

Stop 4. Citadel Ruin (30 minutes)

Citadel Ruin is a Sinagua Indian structure next to the paved road that crosses Wupatki National Monument. No collecting is allowed within the National Monument. The ruin was built on a Woodhouse-age basalt lava that flowed down a tributary of the Little Colorado River. The tributary's channel was eroded in the Moenkopi Formation, and subsequent erosional stripping of the Kaibab Formation resulted in topographic inversion; the lava now forms a low ridge that extends over 7 km northeast of the ruin. The purpose of this stop is to examine the relationships of the lava flow with erosion surfaces and faulting.

Rocks that are exposed in Citadel Sink (Figure 4) include the Kaibab Formation, the Moenkopi Formation, which thickens to the left (east) away from the paleovalley, and the Citadel basalt lava flow, which thickens to the right (west) toward the paleovalley. The southeast-bounding fault of the Citadel graben displaces the Kaibab Formation by a down-to-the-northwest throw of about 16.5 m (54 ft), of which about 5.5 m (21 ft) postdates the lava flow. The sink originated by karst solution in the Kaibab Formation along the fault.
Stop 4. Citadel Sink (40 minutes)

Citadel Sink is the prominent scoria cone northeast of the picnic area on the paved road that crosses Wupatki National Monument. The scoria cone is an associated row of three small coalesced cones to the south erupted along the fracture system of the Black Point section of the East Kaibab monocline. The purpose of this stop is to examine the volcanic products of the Citadel Mountain eruption, see their relationships with Colorado Plateau structures and erosion surfaces, and review the volcanic stratigraphy of the eastern San Francisco volcanic field.

Stop 5. Doney Mountain (40 minutes)

Doney Mountain is the prominent cone northeast of the picnic area on the paved road that crosses Wupatki National Monument. The Doney scoria cone and an associated row of three small coalesced cones to the south erupted along the fracture system of the Black Point section of the East Kaibab monocline. The purpose of this stop is to examine the volcanic products of the Doney Mountain eruption, see their relationships with Colorado Plateau structures and erosion surfaces, and review the volcanic stratigraphy of the eastern San Francisco volcanic field.

Stop 6. The Sproul (40 minutes, optional)

The stop is at the low cone situated on the west side of prominent Merriam Crater. Park on the west side of The Sproul and follow the foot trail to the crater rim. The Sproul is a large spatter cone of Merriam age. The purpose of this stop is to observe the structure of the volcano and the lithology of the lava.

Stop 7. Vent 235 Tuff Ring (20 minutes)

The purpose of this stop is to examine a wide-rimmed vent of Tappan age that is typical of maars in the volcanic field. The tuff contains fragments of basalt from the underlying flow of Woodhouse age, Moenkopi and Kaibab Formations, Coconino Sandstone, and oxidized basalts in a sandy palagonitic matrix. Depositional structures include cross-beds, low-angle dune forms, and inversely graded planar beds. Gravity data indicate a steep-walled crater in the subsurface floored at about 150 m depth; a magnetic high in the center may reflect an intrusive body beneath the crater floor (H.D. Ackerman, J. Hassemer, and J.D. Hendricks, unpub.). The focus of explosion is interpreted to have occurred in the Coconino Sandstone. A subsequent lava flow from the cone just to the north appears to have spilled into the maar.
Stop 8. Merriam Crater (20 minutes)

The stop is on top of a cinder-mantled lava flow that extruded from the northeast base of Merriam Crater, the largest cinder cone in this part of the volcanic field and the type cone of the Merriam age group. Southeast of Merriam Crater is a smaller cinder cone, and nestled between the two cones is the pushed-up, pluglike, layered mass of a third vent. Two flows that were extruded from the latter vent flowed northeast on a low surface beside a Woodhouse-age mesa. The flow from the north side of the vent moved directly northeast toward the Little Colorado River, but the flow from the southwest side first skirted the south side of the southeastern cone before heading down the regional slope; both flows moved in lava channels and formed prominent levees, which can be seen south of this stop. To the north are Woodhouse-age basalt lavas capping east-facing mesas over which lavas of Tappan age cascaded. Between this locality and Roden Crater to the north is the basalt lava that flowed about 13 km (8 mi) to dam the Little Colorado River at Grand Falls (next stop).

Stop 9. Grand Falls (30 minutes)

Grand Falls was formed 0.15±0.03 Ma ago when a basalt flow or series of flow lobes from the Merriam Crater vent group poured into the canyon of the Little Colorado River, overfilling it at the pourover and flowing 24 km down-canyon. The lake behind the new lava dam filled and eventually overflowed around the distal end of the flow (Figure 6), cascading approximately 43 m down the canyon wall. The river canyon is cut into the Kaibab Formation (sandy dolomite), which is overlain by the basal part of the Moenkopi Formation just northeast of Grand Falls. The uppermost cross-bedded dunes of the Coconino Sandstone are exposed beneath the Kaibab Formation below the falls. Alluvial deposits caused by the damming of the canyon extend at least 45 km up the broad river valley.

Second Day: Stop 10. Elend Mountain (40 minutes)

The stop is in a railroad spur between the Ralston Purina plant and the connecting ramp between U.S. 89 and I-40. The southern part of Mount Elden is a composite lava dome composed of bulbous, dacite lobes that flowed radially from at least two extrusion points onto the flat-lying Kaibab Formation 0.57±0.03 to 0.49±0.06 Ma ago. Prior to construction of the exogenous lava dome, Pelean-style eruptions about 4.8 km (3 mi) north of this locality generated pyroclastic flows that deposited a block and ash fan south of the vent. The purpose of this stop is to examine the structure of the lava dome and to study the block and ash deposit.

Six flow lobes, several of which overlap, are visible on the southeast dome of Elden Mountain. The lowest flow displays subhorizontal concentric benches that appear to be related to rAMPING shear fractures, whereas the highest flow is broken by longitudinal tension fractures (Kluth, 1974). The preserved thickness of the block and ash deposit at this locality is about 5 m where it thickens in a paleovalley in the Kaibab Formation. The deposit consists of two parts: 1) a lower layer (0 to 25 cm thick) composed of poorly sorted, structureless to crudely stratified ash and fine lapilli, and 2) an upper layer composed of very poorly sorted ash, lapilli, and blocks up to 1 m in diameter. The basal part of the upper layer (10-25 cm) ranges in character from fines-retained and matrix-supported to fines-depleted and clast-supported. The upper part of the upper layer, apparently structureless, contains matrix-supported essential blocks that range from dense dacite vitrophyre to poorly vesiculated pumice. Concordant paleomagnetic poles of the blocks indicate deposition above the Curie temperature (K. L. Tanaka, 1981, oral comm.).

Stop 11. Black Bill Park (20 minutes)

The stop is on the paved Timberline Estates road a few hundred feet west of U.S. 89. Black Bill Park is an intercone basin bounded by San Francisco Mountain, Elden Mountain, and Tappan-age scoria cones. The purpose of this stop is to examine the eastern part of the San Francisco Mountain volcanic system.

From south to north the following features can be seen on the west side of U.S. 89 (Figures 7 and 8): 1) southeastern dome of Elden Mountain; 2) broad recess in Elden Mountain underlain by east-dipping (30°-65°) Paleozoic strata in a continuous section down from the Kaibab Formation (Permian) to the Temple Butte Formation (Devonian) in contact with intrusive dacite at the base of the cliffs; 3) Little Elden Mountain composed of dacite flow lobes; 4) uplifted block of Paleozoic strata that dips northwest at 17°; strata are overlain by basalt lava flows and a block and ash deposit from a dome on Fremont Peak; 5) Schultz Peak (Brunhes geochron), a composite dacite dome partly buried on its north end by lavas from San Francisco Mountain; 6) Fremont, Doyle, and Reese Peaks on San Francisco Mountain, all capped by outward-dipping andesite lavas about 0.43 Ma old; 7) Sugarloaf Mountain, rhyolite dome extruded 0.22±0.02 Ma ago; and 8) block lava flow of dacite extruded 0.40±0.03 Ma ago from a vent on the upper east side of Doyle Peak.
Stop 12. Bonito Lava Flow (80 minutes)

The Bonito lava flow extruded from the northwest base of Sunset Crater (Hodges, 1962), a scoria cone built during an eruption that began in 1064-1065 A.D. (Smiley, 1958) and continued episodically for about 120 years (D. Champion, written comm., 1985). The basalt lava was extruded in at least three stages and ponded in an intercone basin. North of the Bonito flow is the O’Leary Peak center, which is composed of several silicic lava domes and flows (0.25 to 0.17 Ma) and an andesitic lava flow. The purposes of this stop are to 1) examine the flow units and structures of the Bonito lava flow; 2) discuss the history of the Sunset Crater eruption; and 3) review the volcanic history of San Francisco Mountain.

Beginning at the parking lot on the Bonito flow, follow the route on Figure 9 counterclockwise to observe 1) stage 1 of the Bonito flow, covered with a thick mantle of tephra (Colton, 1967); 2) pahoehoe-type structures on stage 2B, lava that is at a similar level as stage 1 but has a thin and patchy mantle of tephra; 3) a squeeze-up that breaks through the crust of stage 2B; 4) stage 3 lava at a low topographic level; the flow is covered with aa clinkers and lacks a tephra blanket; 5) several hornitos above a lava tube; 6) large spheroidal bombs from the last summit eruption of Sunset cone on top of a small unit of stage 3 lava; 7) the stage 2A unit that was extruded onto the surface of the 1st stage; 8) a spatter rampart at the extrusion point of the stage 2A unit; 9) a pit crater that collapsed when stage 3 lava extruded; and 10) several large mounds of spatter, agglutinate, and rootless flows, some injected by shallow dikes, that were rafted by stage 1 lava when it breached an early cone of Sunset Crater. The high part of the stage 2A unit provides a vantage point from which San Francisco Mountain can be seen (Figure 10). The Inner Basin originated between 0.43 and 0.22 Ma ago as a result of collapse that displaced the top of
San Francisco Mountain outward in debris avalanches and debris flows. Truncated lava and pyroclastic units and buried silicic domes form the walls of the caldera, and the central conduit system of the composite volcano is exposed on the northeast-trending Core Ridge; Sugarloaf Mountain erupted through the largest debris fan.

Figure 9. Geologic map of the Bonito lava flow. Dashed line shows traverse at stop 12. Strike and dip symbols indicate attitude of bedding in agglutinate mounds. FR, forest road; P, parking lot.

Stop 13. Debris Fan (30 minutes)

The stop is at the information sign for Sunset Crater National Monument on Forest Road 545, the paved access road to the Monument. One of nine debris fans deposited around San Francisco Mountain as a result of its collapse forms the surface at this locality. The estimated volume of all the fans is 7.7 km$^3$, which compares favorably with the 8 km$^3$ calculated for the Inner Basin and restored cone.

The debris-fan slopes gently away from San Francisco Mountain, which gives rise to intermittent streams that have dissected the top of the fan. The deposit is coarse, polymictic, very poorly sorted, and poorly consolidated; it is part of the Sinagua Formation of Updike and Peeve (1970). Clasts of a wide variety of San Francisco Mountain lithologies can be seen in a partly excavated Sinagua Indian pit house and gullies south of the information sign.

Return to east Flagstaff on U.S. 89 and take I-40 west to exit 195B; follow the signs to Lake Mary Road, FH-3.

Stop 14. Mormon Lake (25 minutes)

The stop is at a scenic overlook next to FH-3 on the northeast side of Mormon Lake. The lake is bounded on its east side by a normal fault and on its other sides by the Mormon Mountain silicic center and basaltic cones and lavas. The purpose of this stop is to examine the volcanic stratigraphy and structure of the northern part of the Mormon volcanic field.

Basalt sheet lavas (Late Miocene?) form the low-relief surface east of the lake and are exposed in the cliffs along the fault (Figure 11), which has a throw of more than 200 ft. South of the lake a small shield volcano overlies the sheet lavas and both volcanic units are offset by the lake-bounding fault, as well as by a smaller southeast-trending fault that intersects the former to create a wedge-shaped graben in the meadow. Posttectonic dacite lava that was emplaced along the smaller fault forms the dome near the summit of the shield volcano. The shield volcano, dome, and fault scarps are overlain by small scoria cones and basalt lava flows. The Mormon Mountain silicic center on the west side of the lake is composed of several bulbous, block lava flows of dacite (3.1±0.6 Ma), which were extruded radially on top of andesite lavas, and a rhyodacite dome that forms the rounded peak on the south side of the mountain (Gust, 1978).

Figure 11. Photograph of view south along the east side of Mormon Lake at stop 14. Sheet lavas and shield volcano are basalt; both are cut by the faults. Dacite dome was emplaced along the fault that bounds the west side of the graben.

Holm and Ulrich
Stop 15. Mormon Mountain (25 minutes)

The stop is on the west side of Mormon Lake at a wide pullout between the lake and cliffs more than 100 ft high. The cliffs are at the toe of a thick (>55 m, 180 ft) block lava flow of pyroxene dacite that was extruded from the south side of the summit of Mormon Mountain, about two miles west of here. The purpose of this stop is to examine the lava-flow structures and petrography of the dacite.

The bottom part of the flow is a highly fractured zone consisting of randomly oriented open fractures bounding blocks of small or no displacement and autoclastic flow breccia of angular blocks. The middle part of the flow is a dense zone broken by moderately spaced (2-5 cm) ramping shear fractures that dip 60° to 70° west and northwest. Above the cliffs, the top of the flow is mantled by rounded (weathered) blocks of dacite, many of which appear to have little or no displacement. The dacite contains scattered phenocrysts of clinopyroxene and orthopyroxene, and sparse hornblende, in an aphanitic groundmass rich in plagioclase.

The route to stop 16 is south on FH-3 and AZ 87, and west on FH-9 toward Camp Verde; FH-9, the Zane Grey Highway, closely follows the General Crook Trail that connected Fort Verde and Fort Apache in the 1870's as an army supply route.

Stop 16. Mogollon Rm (25 minutes)

The stop is on a wide pullout on the right (north) side of the road 4.8 mi west of the Yavapai County line; at this point, the road begins to descend toward the Mogollon Rim, the southern Colorado Plateau marginal escarpment. Although the present Mogollon Rim is about 4.8 km (3 mi) west of here, sections of middle to late Miocene basalt lavas greater than 550 m (1,800 ft) thick in West Clear Creek to the north (Ulrich and Bielski, 1983) and in Fossil Creek to the south (Twenter, 1962; Weir and Beard, 1984) indicate that a buried ancestral escarpment, developed by pre-middle Miocene erosion into the Supai Group (Pierce and others, 1979), is several miles to the east, where the lavas thin abruptly on the Kaibab Formation. The youngest basalt lavas in this part of the volcanic field flowed westward down a structural ramp from the Colorado Plateau into the Transition Zone (Elston and others, 1974). The purpose of this stop is to review the general geology, volcanism, and tectonism of the southern Colorado Plateau-Transition Zone boundary.

In a general sequence toward the west are 1) Thirteenmile Rock volcanics of Elston and others (1974); 2) Mogollon Rim, which trends north here, but swings around to the west along the north side of the Verde Valley where red and tan Permian strata form cliffs; 3) Verde Valley, an erosional and tectonic basin; 4) light-colored fluvial and lacustrine sediments of the late Miocene-Pliocene Verde Formation that were deposited in the valley when it was blocked at the south end by faulting and volcanism by the Hackberry Mountain center; 5) scarp of the Verde fault, which accommodated down-to-the-northeast throw of as much as 1,800 m (6,000 ft) since 8 Ma ago; 6) Black Hills, a horst capped by subhorizontal, mid-Miocene basalt lavas of the Hickey Formation (14-10 Ma; Elston and others, 1974) that were extruded onto an erosion surface cutting across gently northeast-dipping Paleozoic strata.

Stop 17.

Thirteenmile Rock Volcanics (60 minutes)

Stop 17 is 2 miles west of stop 16; park on the shoulder of the road in a gap between the guard rails. The upper part of the Thirteenmile Rock volcanics of Elston and others (1974) are well exposed in road cuts for the next 3 miles. The section consists of valleyward-dipping basalt lava flows, scoria beds, felsic air-fall tuffs and lapilli tuffs, felsic ignimbrite, a rhyolite lava flow or dome, and volcaniclastic sediments and debris. The basalt lavas were extruded on or near the plateau margin and flowed west toward the Verde Valley, locally filling channels. The felsic pyroclastic deposits presumably originated at the Hackberry Mountain silicic center about 11 km (7 mi) south-southwest of here. Small-displacement normal faults are generally downthrown on the west. The purpose of this stop is to review the stratigraphic relationships of the basaltic and silicic units in the southern Mormon volcanic field and to examine the lithologies and structures of the volcanic deposits.

Walk down the road for about a half mile to reboard the bus. Features of interest include 1) basalt lava flows and air-fall scoria deposits; 2) planar-bedded air-fall tuffs and lapilli tuffs, generally of dacite composition; 3) feeder dikes (1 m thick) of columnar-jointed basalt lava flow and associated agglomerate; 4) air-fall pumice bed overlying a flow-banded rhyolite vitrophyre; 5) basalt lava flow containing plagioclase megacrysts up to 2 cm; 6) small-scale horsts and grabens; and 7) normal faults.

The road down to the Verde Valley passes cuts that expose debris-filled channels, non-welded ignimbrite, felsic tuffs, basalt lava flows, and volcaniclastic sediments. The Hackberry Mountain silicic center can be seen on the left (south) shortly after passing the historical marker at Thirteenmile Rock; several dome-shaped mountains compose the center.

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A Field Guide to the Jemez Mountains
Volcanic Field, New Mexico

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INTRODUCTION

This field trip will examine the evolution of the calc-alkaline Jemez Mountains volcanic field during the past 13 million years. Specific attention will be paid to the types and spatial distribution of the wide variety of volcanic rocks in the field and the pyroclastic deposits of the culminating rhyolitic eruptions of the Bandelier Tuff (1.45 and 1.12 Ma), which caused the formation of the Valles caldera complex. Pre- and postcaldera eruptions, both lavas and pyroclastic rocks, will be examined, as well as surface evidence for the active hydrothermal system under the caldera. The Jemez field is situated on the western edge of the Rio Grande rift in north-central New Mexico (Figures 1 and 2). Products of contemporaneous effusive and explosive basaltic volcanism in the adjacent rift can be seen interstratified with Jemez Mountains lavas on the eastern margin of the field. Rift-filling sedimentary sequences containing many primary pyroclastic units and a wide range of volcaniclastic sediments further allow us to interpret the tectonic and eruptive history of the field.

The evolution of a complex intracontinental volcanic field such as the Jemez Mountains often leads to rhyolitic ignimbrite volcanism, the largest-scale style of silicic volcanism found on earth. Examination of the Jemez Mountains presents an opportunity to trace the petrologic, volcanologic, and structural events involved in the development of this well-exposed volcanic field.

Some of this guide is adapted from the previous field-trip guides of Bailey and Smith (1978) and Goff and Bolivar (1983).

JEMEZ MOUNTAINS STRATIGRAPHY

The general stratigraphy of the Jemez volcanic field comprises these three groups: Keres, Polvadera, and Tewa (Figure 3). The Keres Group is volumetrically dominated by the Faliza Canyon Formation (10 to 7 Ma; dominantly andesite with subordinate basalt and dacite), but includes two high-silica rhyolite formations (Canovas Canyon and Bearhead) that constitute a continuum from <13 to 6 Ma. Intimately related to and commonly interbedded with the volcanic rocks of the Keres Group are the volcaniclastic basin-fill deposits of the Coohtii Formation. The Polvadera Group includes the Lobato Basalt, Tschicoma Formation (dominantly dacite), and El Rechuelos Rhyolite. As defined, the group spans >13 to 2 Ma,
finally ceased at about 0.13 Ma, but vigorous hydrothermal activity persists to the present.

TECTONIC AND VOLCANIC HISTORY OF JEMEZ MOUNTAINS

The Jemez volcanic field lies at the intersection of the Jemez lineament with the western boundary faults of the Rio Grande rift. The rift is a major crustal feature of Miocene to Recent age characterized by thin (35 km) crust and by late Cenozoic basaltic volcanism (see McKenzie, 1979). Several en-echelon sedimentary basins are included in the rift, and volcanic rocks of the Jemez Mountains overlie and interfinger with sediments of the western Española basin. The Jemez lineament was originally recognized as a northeast-trending alignment of young volcanic fields stretching from eastern Arizona to southeastern Colorado (Mayo, 1958; Laughlin and others, 1976), but recent work (Baldridge and others, 1980, 1983; Aldrich and Laughlin, 1984) has shown that regionally, it represents a complex zone of concentrated tectonic activity. Within the Jemez volcanic field the lineament is an actual structural entity inherited from a fault zone originating in the Precambrian basement. It is expressed as the Jemez fault zone in San Diego Canyon, as the structure within the resurgent dome of Valles caldera (Smith and Bailey, 1968; Goff and Gardner, 1980), possibly as the depression northeast of Valles caldera (Gardner and Goff, 1984), and as faults occurring northeast of there (Aldrich, 1986). The depression northeast of Valles caldera, previously known as the Toledo caldera (Smith and others, 1970), is called herein the Toledo embayment (Figure 2).
Figure 3. Chronostratigraphic chart of major units in the Jemez Mountains. Information from Bailey and others (1969), Smith and others (1970), Gardner and Goff (1984), McPherson and others (1984), and Self and others (1986).

Smith and others (1978) suggest that the unique compositional variety and much larger volume of silicic to intermediate eruptives in the Jemez field, relative to other volcanic fields along the Jemez lineament, may be due to its position at the intersection of the lineament with the Rio Grande rift. Tectonic activity at the intersection of these two structures has not only controlled the spatial focus of volcanism, but has influenced the petrogenesis of the volcanic field as well (Gardner, 1983; Gardner and Goff, 1984; Gardner and others, 1986).

Volcanic activity in the Jemez region began with mantle-derived alkaline basalts at about 16.5 Ma. The distribution and vent localities for these basalts are not known because they are poorly exposed and interbedded with basin-fill sediments of the Santa Fe Group. In the period 13-10 Ma the Jemez Mountains volcanic edifice began to build. Early Keres Group volcanism was predominantly basaltic and rhyolitic with minor intermediate products. Evolved basaltic lavas (Mg/(Mg+Fe2) = 0.40 to 0.65) of this period are interbedded with coarse-grained debris-flow deposits and basin-fill gravels (Cochiti Formation), and high-silica rhyolite lavas and pyroclastic deposits were erupted from vents aligned along N-S-trending faults. These relations, together with the geometry of Cochiti deposits, suggest that intense tectonic activity occurred during this period and that faults provided conduits for the magmas. Although exposures are sparse, it appears that the basalts were erupted in the vicinity of the Jemez lineament-Rio Grande rift intersection and flowed eastward into the rift.

Basalt and high-silica rhyolite continued to be erupted during the period 10-7 Ma but were volumetrically overwhelmed by effusion of some 1000 km³ of Paliza Canyon andesite and subordinate rhyodacite. This activity was centered on the lineament-rift intersection even more strikingly than the basaltic activity prior to 10 Ma, and half the volume of the entire Jemez volcanic field was erupted in this brief 3-m.y. period.

Before discussions of the more recent history of the Jemez field it is necessary to examine some events that took place beginning about 7 Ma. A compilation of available K-Ar dates for predominant rock types from the Jemez Mountains region shows three noteworthy features (Gardner and Goff, 1984): (1) a probable 3-m.y. gap in basaltic volcanism occurred from 7 to 4 Ma; (2) at 7 to 6 Ma, coincident with the beginning of the lull in basaltic volcanism, there was a change in the composition of intermediate volcanism from dominantly andesitic to dacitic; and (3) the revival of basaltic volcanism at about 4 Ma is nearly contemporaneous with the onset of the earliest eruption of "Bandelier-type" rhyolitic magma. Furthermore, at about 7 Ma there was apparently a sharp reduction in the "rate" (expressed as volume per unit of time) of volcanism. This lull in Jemez basaltic activity is also coincident with an apparent regional gap in basaltic volcanism (Baldridge and Perry, 1983), suggesting a link to a more regional event. Gardner and Goff (1984) suggest that all these temporally coincident events indicate a lull in tectonic activity. Furthermore, Gardner and others (1986) point out that the geometry of the Puye Formation, built largely in this period, is also indicative of tectonic stability.

Gardner (1982, 1983, 1985) suggests that transition from predominantly andesitic to dacitic volcanism at about 6-7 Ma reflects a major change of
magma processes in response to the tectonic lull. Instead of being rapidly erupted, as in earlier, more tectonically active stages of the field's development, pockets of basalt and basaltic andesite began to coalesce with pods of high-silica rhyolite, giving rise to growing intermediate magma chambers, from which small portions were erupted as the hybrid dacitic lavas of the Tschicoma Formation. The focus of volcanism in this period (7-4 Ma) was again centered on the lineament-rift intersection. It is noteworthy that while there was a lull of basaltic volcanism, available evidence suggests that there was ongoing basaltic magmatism at depth in the form of replenishing pulses into growing dacitic magma chambers (Eichelberger, 1980; Loeffler, 1983, 1984).

Since 4 Ma, with renewed tectonic activity in the rift, further basaltic volcanism was distinctly peripheral to the main body of the volcanic field and rift-related faulting shifted eastward of its former position to the Pajarito fault zone (Gardner and Goff, 1984). These events, together with essentially contemporaneous small-volume eruptions of Bandelier-type

Figure 4. General geologic map of Jemez Mountains showing field-trip stops. Random dash = Tewa Group; coarse, regular stipple = Polvadera Group; irregular stipple = Keres Group; horizontal rule = young, flanking basalt fields; CTE = Cañada de Cochiti fault zone; PPF = Pajarito fault zone; JFZ = Jemez fault zone; SFZ = Santa Ana Mesa fault zone; SPD = St. Peter's Dome; VC = Valles caldera; R = resurgent dome of VC; T = Toledo embayment.
rhyolite magma (Kite and others, 1982) and ongoing eruptions of dacitic lavas in the Tschicoma volcanic center (Doell and others, 1968; Smith and others, 1970; Smith, 1979; Gardner and others, 1986), indicate the existence of large, shallow magma chambers, centered beneath the lineament-rift intersection by about 4 Ma. Compositional data indicate that these large magma chambers, whose bulk compositions are best approximated by the hybrid dacites, were possibly parental to the rhyolitic-dacitic Bandelier magma chamber (as suggested by Smith, 1979), from which the rhyolitic deposits discussed below were erupted.

**DAY ONE: TECTONIC AND MAGMATIC DEVELOPMENT OF THE JEMEZ VOLCANIC FIELD**

The route of the first day of the field trip takes us across the southern Jemez Mountains where we will see representatives of most precaldera magma types recognized by Gardner (1985) and Gardner and others (1986; Figure 4). Although most of the rocks we will look at today are of the Keres Group, the group contains rocks petrologically analogous to rocks of the Polvadera Group. We will also view the Pajarito fault zone, which is a major active boundary fault for the Rio Grande rift, and deposits of the Cochiti Formation, which have implications for tectonic activity early in the volcanic field's history.

**Day One Road Log**

<table>
<thead>
<tr>
<th>Mileage</th>
<th>Begin Day One Field Trip</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.0</td>
<td>Begin mileage for this day's field trip on south side of Santa Fe at overpass at junction of I-25 and US 285. Head south on I-25 into the southern Española basin of the Rio Grande rift. Between 1 and 3 o’clock in the distance are the Jemez Mountains. In the middle distance are the low hills of the Cerros del Río. At 10 o’clock are the Ortiz Mountains that have a white scar marking the Ortiz gold mine.</td>
</tr>
<tr>
<td>10.6</td>
<td>Turnoff #271 at La Cienega; continue south on I-25; on left Cerrillos Hills, Eocene volcanic hills with turquoise mines. On right road cuts of Santa Fe Group and basalts of Cerros del Río.</td>
</tr>
<tr>
<td>14.4</td>
<td>Santo Domingo basin of the Rio Grande rift comes into view with Sandia Mountains on skyline.</td>
</tr>
<tr>
<td>15.8</td>
<td>Head of La Bajada grade, a fault scarp separating the Santo Domingo and Española basins.</td>
</tr>
<tr>
<td>16.1</td>
<td>Basalt dike on left.</td>
</tr>
<tr>
<td>16.6</td>
<td>Tilted Mesozoic and Eocene rocks on right.</td>
</tr>
<tr>
<td>17.1</td>
<td>Exit on turnoff #264 toward Cochiti Pueblo.</td>
</tr>
<tr>
<td>17.3</td>
<td>Turn right on NM 16.</td>
</tr>
<tr>
<td>18.2</td>
<td>White Zia sandstone at base of La Bajada fault on right.</td>
</tr>
<tr>
<td>23.3</td>
<td>Junction with road to Cochiti Lake; continue on NM 16.</td>
</tr>
<tr>
<td>25.3</td>
<td>Junction with NM 22, turn right toward Cochiti Pueblo.</td>
</tr>
<tr>
<td>27.9</td>
<td>Turn left toward Cochiti Pueblo on NM 22.</td>
</tr>
<tr>
<td>29.6</td>
<td>Turn right on FS 266 (dirt road) toward Bear Springs.</td>
</tr>
<tr>
<td>30.8</td>
<td>Rift-fill gravels on hills at right.</td>
</tr>
<tr>
<td>32.1</td>
<td>Bandelier Tuff overlies rift sediments on right.</td>
</tr>
<tr>
<td>34.5</td>
<td>Turn right on dirt road toward white cliffs.</td>
</tr>
<tr>
<td>34.6</td>
<td>Park anywhere on dirt road.</td>
</tr>
</tbody>
</table>

**STOP 1: PERALTA TUFF MEMBER OF THE BEARHEAD RHYOLITE**

This is the type area for the Peralta Tuff Member of the Bearhead Rhyolite (Bailey and others, 1969). Here the bedded deposits of high-silica rhyolite pyroclastic material are mainly reworked or water-lain tuffs with subordinant fallout and flow units (Figure 5). The tuff was apparently vented from within a dome-and-flow complex of high-silica Bearhead Rhyolite about 10 km
northwest of here. Sanidine from pumice lumps, collected about 200 m north of here, yielded a K-Ar date of 6.85±0.15 Ma (Gardner and Goff, unpub. data, 1985). The Peralta Tuff forms a prominent stratigraphic marker in the southern Jemez Mountains, and as such, approximates the somewhat arbitrary boundary at about 7 Ma between the Canovas Canyon and Bearhead Rhyolites (Gardner and others, 1986). These two rhyolite formations comprise a continuum of high-silica rhyolite volcanism from >13 to about 6 Ma, with the Bearhead representing an apparent volumetric pulse that postdated most of the voluminous andesitic activity. Working from the rift basin to the south, Kelley and others (1976) included the tuffs at this locality as a facies of the Santa Fe Group rift-fill sequence. These are also some of the most photographed "tent rocks" in New Mexico.

43.1  Park on right side of road just beyond summit of hill.

STOP 2:  VISTA OF SOUTHEASTERN JEMEZ

From this vantage point we can look WNW and see Tertiary to Quaternary domes and ignimbrites of the Jemez volcanic field on the west margin of the Rio Grande rift (Figure 6). The rounded hills are primarily andesitic and rhyolitic domes of the Keres Group (>13 to 6 Ma). Bearhead Peak is the type locality of the Bearhead Rhyolite. Cerro Boletas is a sequence of bedded rhyolite tuffs of the Bearhead Rhyolite. Cerro Picacho is yet another Bearhead dome and St. Peter's Dome is a complicated pile of andesites and inter-stratified volcaniclastics. Low places in the Keres Group volcanics were subsequently filled by the mesa-forming Bandelier Tuffs (1.45 to 1.12 Ma) from Valles and Toledo calderas. The visible scarp is the Pajarito Fault which has been periodically active in this area for the last ~16 Ma. In the foreground are various Quaternary units (poorly mapped) that are mostly Quaternary in age and have been shed from the Jemez Mountains toward the Rio Grande. Continue straight ahead (FS 268).

44.3  Golf course on right; begin dirt road.

45.1  Junction FS 289 on right; continue straight ahead.

46.5  Junction FS 89; bear left toward Bland Canyon.

47.8  Mouth of Bland Canyon; mesas on both sides capped with Bandelier Tuff.

48.1  Park along road near culvert and walk 150 m south to small hill along Bland Creek.

STOP 3:  PAJARITO FAULT IN QUATERNARY TERRACE GRAVEL

The outcrop adjacent to Bland Creek shows a high-angle trace of the Pajarito Fault that dips SE toward the Rio Grande rift. The fault juxtaposes 6.85 Ma Peralta Tuff (west) against Quaternary terrace gravels shed from the Jemez Mountains. The fault is easily followed both SW and NE from here and produces spectacular benches in mesas of Bandelier Tuff.

51.3  Cattle guard; as you cross over ridge note that Bandelier Tuff is capped with Quaternary gravels.

52.2  Ford on Cochiti Creek.

Figure 6. View of southeastern Jemez Mountains looking WNW; BP = Bearhead Peak, BC = Bland Canyon, CC = Cochiti Canyon, CB = Cerro Boletas, CP = Cerro Picacho, and SD = St. Peter's Dome.

100  Self and others
Contact between Upper and Lower Bandelier Tuffs can be seen in road cut on left and across canyon to right.

Ascend Pajarito fault scarp.

Pull off alongside of road so traffic can pass.

STOP 4: PAJARITO FAULT OVERLOOK: UNCONFORMITY

From here we can look SE into the Santo Domingo basin of the Rio Grande rift toward the Sandia Mountains on the far side of the rift. The Pajarito fault creates a large bench in the Bandelier Tuff. The canyon below us exposes Paliza Canyon andesite flows (9 Ma) and interbedded sedimentary rocks of the Cochiti Formation. On the other side of the road is a fine exposure of surge deposits at the base of the Upper Bandelier Tuff that is draped over the Cochiti Formation (Figure 7). To the east along the road, these gravels of the Cochiti Formation are overlain by hydrothermally altered, two-pyroxene andesite. Continue straight ahead on FS 289.

Ridge at right is capped with two-pyroxene andesites of Paliza Canyon Formation (Keres Group) but flanked by younger Upper Bandelier Tuff.

Landslide overlook into Cochiti Canyon on left.

Pull off road so other cars can pass.

STOP 5: INTRUSIVE BEARHEAD RHYOLITE

The hill on our right is Cerro Boletas, which is composed primarily of well-bedded Peralta Tuff (6.85 Ma). Along the road we can see a vertically sheeted intrusion of Bearhead Rhyolite that is devitrified to the SE but is glassy to the NW along the intrusive contact with tuff. This intrusion has a K-Ar date of 6.2 Ma and is very typical of Bearhead and Canovas Canyon Rhyolite intrusives (13 to 6 Ma) exposed in the labyrinth of canyons in the southeastern Jemez Mountains. Continue straight ahead on FS 289.

Ridge at right is capped with andesites of Paliza Canyon Formation (Keres Group) but flanked by younger Upper Bandelier Tuff.

Landslide overlook into Cochiti Canyon on left.

Pull off road so other cars can pass.

STOP 6: COCHITI CANYON OVERLOOK; LUNCH

We are looking SE into Cochiti Canyon, which exposes orange cliffs of Bandelier Tuff overlying a thick sequence of gravels of the Cochiti Formation. Near the bottom of the canyon some hydrothermally altered andesite flows are interbedded with the gravels. The Cochiti Formation dips primarily to the south and east towards the Rio Grande rift and apparently fills in the topography between Keres Group andesitic stratovolcanoes. Ages of the various andesites in this area range from 9.5 to 8.5 Ma.

Continue straight ahead on FS 289.

Junction FS 142 to St. Peter's Dome; bear left on FS 289.

Road now winds through cluster of andesites and dacites of Keres Group.

Pull off road at summit of hill so other cars can pass.

STOP 7: ANDESITE VENT AND DOME

This hill is the vent for a large andesite dome and flow that are part of a complex of similar rocks exposed in this area. Scoriaceous rocks are exposed near the summit of the hill. The andesite is typical of many two-pyroxene andesites of the Paliza Canyon Formation.

Continue straight ahead on FS 289.

Road winds over north shoulder of dacite dome.

Figure 7. Photograph of unconformity showing Upper Bandelier Tuff over Cochiti Formation.
Rabbit Mountain, a post-Toledo caldera rhyolite dome, looms into view.

Junction FS 248; bear left on FS 289.

Junction FS 36 at Graduation Flats; turn left on FS 35.

Bedded rhyolite tuffs associated with Rabbit Mountain along road.

Hydrothermally altered Keres Group andesite.

Junction FS 268; turn right on FS 268.

Redondo Peak on skyline.

Junction FS 284 on left; continue ahead on FS 268.

Del Norte Pass; descend into Valles caldera along southeast caldera wall in hydrothermally altered andesite.

Junction with NM 4; turn left on pavement. Redondo Peak and South Mountain rhyolite dead ahead. Los Griegos and Los Conches Peak to left (south). Drive into southeastern most of Valles caldera.

Vertically sheeted core of South Mountain rhyolite flow on right.

Pull off into parking area of campground or alongside of NM 4.

STOP 8: SOUTH MOUNTAIN RHYOLITE AND HYDROTHERMALLY ALTERED BASALT

At this location we can observe the geology of the southern most of the Valles caldera. South Mountain rhyolite (north side of highway) is a post-caldera, crystal-rich, high-silica rhyolite (0.49 Ma) that flowed over and against caldera wall rocks. At this point, caldera wall rocks consist of hydrothermally altered basalt of the Paliza Canyon Formation. The age of the alteration is not known. It may be correlative with hydrothermal events in the Cochiti mining district (26 Ma) or it may be associated with hydrothermal systems formed along with creation of Valles and Toledo calderas (c.45 Ma). Alteration such as this occurs at a few places along the west, north, and southeastern wall of the caldera.

Continue west on NM 4.

El Cajete Pumice overlies South Mountain rhyolite.

Turn left on FS 10 (unmarked) where the long row of mailboxes occurs.

Begin ascent up south caldera wall in Bandelier Tuff.

View of Redondo Peak and southernmost of Valles caldera.

Junction FS 135 at rim of caldera; continue straight ahead on FS 10 toward Ponderosa.

Junction FS 269; bear left on FS 10.

Upper Bandelier Tuff on right.

Pull over along side of road so other cars can pass.

STOP 9: TWO-PYROXENE ANDESITE OF THE PALIZA CANYON FORMATION

This andesite, although a bit glassy, is fairly typical of the rocks that originally constituted half of the volume (1000 km³) of the entire volcanic field. The Paliza Canyon andesites contain augite, bronzite-hypersthene, labradorite-andesine, magnetite-ilmenite + minor olivine as phenocrysts. All phenocryst phases occur in clots that apparently represent the form in which the minerals were removed from the magmas during fractional crystallization.

Continue ahead of FS 10.

Upper Bandelier Tuff on left; Cerro del Piño dead ahead.

Descend grade through tilted and faulted sequence of Santa Fe Group sediments and Keres Group basalt.

Junction FS 270; bear right on FS 10.

Pull over alongside of road so other cars can pass.

STOP 10: CERRO DEL PINO DACITE OF THE PALIZA CANYON FORMATION

We have just driven around the toe of a flow vented from a dome about 2 km east-northeast of here. This is the westernmost dome of a chain of dacite domes that span about 6 km with a curious east-west trend. We have identified no structural control on the trend of this dome complex, which is, in fact, transverse to the main structural grains of the Jemez Mountains. The Cerro del Piño dacite can be easily examined here in large boulders of float in the gully at the bend in the road. The rocks are typical of Paliza Canyon dacite and are megasporically identical to dacites of the Tship coma Formation. The dacite contains plagioclase (cores of phenocrysts are identical to the andesite phenocrysts, but rims and microlites range to oligoclase), augite, bronzite-hypersthene, hornblende, ± minor biotite. The megasporic "mafic" inclusions consist of vesiculated glass, acicular hornblende, and skeletal Ca-plagioclase. These inclusions represent basaltic magma that was injected into the differentiating dacitic chamber (see, for example, Etsehberger, 1990).

Continue ahead on FS 10.

Paliza Canyon unfolds on left.

Pull off road at bend so other cars can pass.

STOP 11: OVERLOOK OF PALIZA CANYON AND BORREGO MESA

From our perch atop the Tshirege Member of the Bandelier Tuff, we can see heaps of Paliza Canyon andesite cut by faults of the Canada de Cochiti fault zone (east to southeast), Borrego Mesa, basalt capped with andesite (9 Ma; Luedke and Smith, 1978; south-southeast), the transition from the Albuquerque basin of the Rio Grande rift to the Colorado Plateau (south to southwest), a distant Mount Taylor, the next volcanic field of the Jemez lineament to the southwest, and the Nacimiento Mountains, a Laramide structure (southwestern to western skyline).

Continue ahead on FS 10.

Poorly exposed Triassic red shales of Chinle Formation.

Turn left on FS 271 toward Paliza Canyon.

Gravels of Cochiti Formation on left.

Pull off side of road beyond cattle guard.
STOP 12: DACITIC PLUG AND COCHITI FORMATION DEPOSITS

This near-volcanic plug is one of several in the southern Jemez Mountains that show geochemical similarities to dacites of the Tschicoma Formation. Petrologically, these dacites appear to have been generated by mixing of andesite and high-silica rhyolitic magmas. This plug probably intruded surrounding Cochiti Formation gravels, which are nicely exposed about 100 m farther up the canyon. This is close to the western limit of the Cochiti Formation. Here the maximum thickness of the immature basin-fill sequence is about 30 m, but the formation thickens to more than 300 m into the rift to the east. Time permitting, a 2- to 3-km hike up the canyon brings one to some excellent exposures of Canovas Canyon high-silica rhyolite tuff with overlying Paliza Canyon basalt (13±1.0 Ma; Gardner and Goff, 1984) on the north wall of the canyon.

Turn around and retrace route toward south. Junction FS 10; turn left to south. Junction FS 12; continue ahead on FS 10. Group campground on left; cliffs of Bandelier Tuff over Chinle Formation on right.

91.3 Pavement begins.
91.8 Chinle Formation on left in village of Ponderosa.
92.3 Borrego Mesa capped with andesite on left.
92.9 Two ledges of basalt with rhyolite tuff in between exposed on Borrego Mesa.
93.7 Ponderosa Bar on left; begin NM 290.
94.7 Santa Fe Group exposed at bottom of Borrego Mesa.
98.0 END OF DAY ONE: Junction NM 4; turn left and drive to NM 44 at San Ysidro; turn left and drive to I-25 at Bernalillo; turn left on I-25 and return to Santa Fe.

DAY TWO: BANDELIER TUFFS, VALLES CALDERA, AND HYDROTHERMAL SYSTEM

Today's fieldwork concentrates on the rhyolitic eruptives of the Bandelier Tuffs, two major ignimbrite sheets, each with volumes of several hundred cubic kilometers (Figure 8). Rhyolitic ignimbrites precursory to the Bandelier eruptions will be seen, as well as the products of the youngest (0.13 Ma) postcaldera ring fracture eruption from the Valles caldera, here called the El Cajete system (ECS) deposits. The hydrothermal system circulates at temperatures of 220° to 300°C at depths of 600 to 2000 m in fractured intracaldera tuffs and precaldera andesites. No surface hot springs representative of the deep hydrothermal system discharge in the caldera. Instead, surface

Figure 8. Map of the distribution of the Bandelier Tuff in the Jemez Mountains [adapted from Smith and others (1970)].
manifestations are characterized by acid-sulfate hot springs, mud pots, and fumaroles. A hydrothermal outflow plume discharges from the caldera in the subsurface down the Jemez fault zone and forms derivative hot springs in San Diego Canyon. Dilute hot springs representing relatively shallow circulation issue from several locations in the western moat zone of the caldera (Goff and others, 1981; Goff and Grigsby, 1982; Goff and others, 1985).

Day Two Road Log

Cumulative mileage  

Start  

Today’s road log begins at Plaza de Santa Fe.

0.0  
Northeast corner Plaza de Santa Fe/Palace St., Santa Fe. Proceed east on Palace.

0.4  
Left turn at Paseo de Peralta (by Posada de Santa Fe Hotel). Proceed along Paseo de Peralta to junction with Guadalupe St.

2.3  
Turn right onto Guadalupe St.

2.9  
Merge with US 285 (also US 64 and 84) going north. Route runs through Española basin of the Rio Grande rift.

11.6  
On left Camel Rock, unusual erosional pedestal rock in Santa Fe Group (Skull Ridge member of Tesuque Formation) composed of sandstones and silts.

15.7  
Turn left (west) off US 84/285 onto NM State Road 4 (NM 4), signedposted to Los Alamos, at Pojoaque. Road traverses Santa Fe Group, Rio Grande rift-fill sediments, and largely unconsolidated sands and silts. Pojoaque Member of the Tesuque Formation forms prominent west-dipping sediments off to the right. Recent uplift and degradation of these sediments explain the absence of young Jemez Mountains tephra in the Española basin. View ahead of Jemez Mountains. Road descends into Velarde graben (Figure 2). Cross Rio Grande at Otowi bridge.

25.6  
Pass road up Guaje Canyon on right; exposures of Puye Formation overlying Chamita Formation.

28.2  
Pull into parking area on left.

STOP 13: END OF OTOWI MESA: BANDELIER IGНИMRITES

The two Bandelier Tuffs, lower (Otowi) and upper (Tahirege), herein designated LBT and UBT, are seen here with an erosional unconformity between them. They rest on lacustrine silts and soils, which in turn overlie a basal unit dated at 2.4 Ma. The general Bandelier ignimbrite stratigraphy and details of the two ignimbrites will be discussed in detail as we ascend the mesa. The two plinian deposits will be discussed in more detail later in the guidebook.

Continue straight ahead (west) on NM 4. Y-junction (split of NM 4); continue straight ahead toward Los Alamos.

28.9  
Clinton P. Anderson Memorial overlook. View of Pueblo Canyon, North Mesa, Kwage Mesa, and Otowi Ruins. Alternate photo stop.

30.2  
Junction NM 4 and Canyon Road, Los Alamos. Right turn onto Canyon Road.

33.4  
View of medial Pueblo Canyon on right and Upper Bandelier Tuff, more intensely welded than at Stop 13. Approaching Diamond Drive, see ahead andesite/dacite lava domes and flows of southern part of the Tschicoma center.

35.3  
Turn right (north) at junction with Diamond Drive.

36.4  
Turn right into First Baptist Church parking lot. Opposite are upper welded flow units of Upper Bandelier Tuff with well-developed interbedded surge deposits.

STOP 14: SURGE DEPOSITS IN THE UPPER BANDELIER TUFFS

The proximal, welded, upper part of the UBT here contains beds of pyroclastic surge material with prominent cross-bedding (Figure 9). Surge unit thicknesses are approximately 0.5 m. Amplitude and wave lengths of dune forms are approximately 0.4 m and 2.0 m, respectively. Bedset angles vary up to 35°. These units occur in the upper cooling unit of the UBT, which has a gray color because of welding. The crystal-rich surge beds are intimately related to welded flow units. Sparse lithics of country rock (Madera Limestone of Carboniferous age) occur. These surges may be either ground surge deposits left by the passage of flow units or surge layer deposits at the edge of the flows channeled down local drainages. Flow direction was roughly eastward perpendicular to the outcrop face.

37.6  
Turn left (south) on Diamond Drive.

39.0  
Turn right onto unsignposted road opposite hospital just before bridge crossing Los Alamos Canyon. Follow road into, along, and out of canyon, which is cut in Upper Bandelier Tuff. Good example of box canyon cut in Upper Bandelier Tuff.

42.2  
Turn right (west) onto NM 4 (West Jemez Road). On right is scarp (about 70 m) of Pajarito fault, which offsets Upper Bandelier Tuff by more than 100 m. Latest

Figure 9. STOP 14: Pyroclastic surge bed between two welded flow units of Upper (Tahirege) Bandelier Tuff showing steep dune-form, laminar bedding, and pinch-and-swell structure. Flow direction from left to right obliquely out of page. Scale is 15 cm long.
movements are younger than 20 ka (J. N. Gardner, unpub. data). West Jemez Road parallels base of fault scarp.

On right and left, pass through low road cuts of welded Upper Bandelier Tuff with interstratified surges. Equivalent to Stop 14 in stratigraphic position.

42.9 T-junction. Turn right onto other branch of NM 4. Road immediately climbs Pajarito fault scarp. Exposures of densely welded Upper Bandelier Tuff on left.

43.8 Pull off road, EXERCISE GREAT CARE BECAUSE OF DANGER FROM TRAFFIC. Roadcuts of the most conveniently reached, densely welded Upper Bandelier Tuff ignimbrite. Not densely welded on a "world scale" but density of 2.2-2.3 g/cm³ (c.f. 1.1 g/cm³ of nonwelded ignimbrite) and flattening ratios of 6:1 in flanche (squashed pumices).

STOP 15: WELDED UPPER BANDELIER IGNIMBRITE

The Upper Bandelier tuff exposed on the Pajarito fault scarp is about the most welded found in the outflow sheet. These are presumably late-erupted flow units, perhaps hotter upon emplacement than lower units because of formation from collapse of lower eruption columns. The grey pumice-poor ignimbrite is devitrified and slightly vapor-phase altered. Lithic clasts are sparse. Thin surge beds can be found, indicating the presence of several flow units. The surge beds are not as intensely welded as the body of the ignimbrite flow unit. Note also vista from top of fault scarp over Pajarito plateau (Bandelier Tuff surface) and Río Grande rift to Sangre de Cristo Mountains. Continue west on NM 4.

45.4 Tschicoma dacitic lava exposed above level of Upper Bandelier Tuff surface. Alternative stop to examine grey porphyritic lava. View of upper Frijoles Canyon on left. Canyon is incised deeply into Lower Bandelier Tuff.

46.4 Cross head of Frijoles Canyon, poor exposures of welded, grey Upper Bandelier Tuff. Pass junction of road to Cochiti Lake (FS 289).

49.6 Road begins to descend into Valles caldera. On left, dacitic lavas of Tschicoma center in caldera wall.

50.9 Pull off side of road.

STOP 16: VALLES CALDERA; VALLE GRANDE

The caldera is about 22 km from east to west and contains 3000 m of fill down to the basement (Nielson and Hulen, 1984). Most of the depression you see was formed at 1.45 Ma by the Otowi (Lower Bandelier Tuff) eruption. Domes outline the ring fracture systems. Redondo Peak (3431 m; 11,254 ft) on the resurgent block of the caldera, towers 800 m above the caldera. Continue southwest on NM 4.

51.6 On the left skyline lies Rabbit Mountain, a remnant of an obsidian dome of Toledo age (post-Lower Bandelier Tuff, pre-Upper Bandelier Tuff).

54.6 Pass South Mountain rhyolite lava flow (0.49 Ma), one of the Valles caldera (post-Upper Bandelier Tuff) lava domes. This dome and flow were examined on Day 1. Road crosses east fork of Jemez River and rises onto South Mountain lava flow.

Two fall units of thick El Cajete plinian pumice; EGS: 0.13 Ma) overlie South Mountain rhyolite. This is the youngest plinian deposit from the Valles caldera. Road passes through several exposures of El Cajete deposit in the next 3 km. Turn left at row of mailboxes onto FS 10 (unmarked). Follow route of Day 1 to FS 135, across caldera moat, and up southern wall of caldera. El Cajete plinian deposit in road cuts on left.

 Junction with FS 135. Turn right to Cat Mesa.

Proceed on FS 135 through forest preserve into dry (generally) gully and out again. On steep rise, just before hairpin left hand curve, turn right (north) into forestry track and park. Follow track north for about 100 m until you reach the promontory on the edge of the caldera.

STOP 17: CAT MESA: PROXIMAL LOWER BANDELIER IGNIMBRITE (LUNCH)

View over southwest part of caldera. Refer to Figure 10.

Lithic breccias: proximal ignimbrite lithic breccias occur in Lower Bandelier Tuff in cliffs below the mesa. WARNING: STEEP CLIMB DOWN AND UP TO OBSERVATION POINT. REMEMBER YOU ARE AT 8,400 FT. Self and others (1986) interpret this breccia horizon as indicative of ring fracture vents within 2-3 km north of this site during the Otowi eruption. Turn around and retrace route on FS 135 and 10 to paved NM 4, turn left, and proceed west.

Pull into parking area off road to right at bottom of descent to east fork of Jemez River.

STOP 18: TYPE SECTION OF EL CAJETE PUMICE DEPOSITS; BANCO BONITO LAVA

The section exposed here (Figure 11) shows South Mountain rhyolite (0.49 Ma) overlain by El Cajete pumice fall deposits (0.13 Ma), surge beds, and pyroclastic flow deposits, in turn overlain by the Banco Bonito obsidian flow. A disconformity occurs between the El Cajete and two thin units associated with the lava flow. The vents for this youngest eruptive episode of the Valles caldera are located about 2 km to the north.

The El Cajete plinian deposit is widespread in the southeast Jemez Mountains and has a bulk volume of about 5 km³. It was formed at the same time as small local pyroclastic flows and surges, seen interstratified with the plinian units. More voluminous pyroclastic flows were generated, which flowed down San Diego Canyon, forming the Battleship Rock ignimbrite.

Proceed west on NM 4.

On right is exposure of proximal, non-incipiently welded proximal Battleship...
Figure 10A. View of Cat Mesa looking SE, showing caldera wall exposure of Lower (Otowi) Bandelier Tuff (LBT). Note: Lens-shaped proximal lithic breccia horizon (lower arrow) some 5 m thick; offset of LBT by small fault (f on downthrown side). Breccias rapidly die out into lithic-rich ignimbrite to right of fault; onset of strong vapor-phase alteration at prominent notch (upper arrow); onset of welding in LBT (above upper arrow). Total exposure thickness shown is 50 m.

Figure 10B. Detail of lithic lag breccia bed in LBT at Cat Mesa. Large block (arrowed) is about 2 m in diameter. Note lithic breccias in flow unit below main breccia bed.

Figure 11. Interbedded plinian pumice fall beds (P) and nonwelded ignimbrite flow units (i) with associated surge deposits (S) in the El Cajete member of the youngest (0.13 Ma) eruption products from the Valles calderas. Above is flow foot breccia on Banco Bonito obsidian lava, produced during same eruptive event. Note disconformity between El Cajete pumice deposits and pyroclastics associated with lava extrusion (dashed line, right). State Road 4 at crossing of East Fork, Jemez River.
Rock ignimbrite with lenses and pipes of lithics, representing a type of proximal breccia facies. These are overlain by rubble base of Banco Bonito obsidian flow. Road rises onto Banco Bonito flow surface. Flow-banded obsidian exposed on left of road 50 m to west.

On Banco Bonito lava surface. Indistinct undulations are pressure ridges on lava-flow surface.

Road descends off Banco Bonito lava flow. Poor exposures of nonwelded Battleship Rock ignimbrite on the right.

Right turn to Sulphur Springs on FS 105. On right are Recent reworked caldera sediments containing high proportions of hydrothermally altered material. Road runs NNE from caldera moat towards Redondo Border, a block-faulted area on resurgent block to west of Redondo Peak.

Note: Sulphur Springs is private property. This stop is marked as alternative because of lack of free access. Road-log mileage continues without addition of 5 mi round trip to the springs.

STOP 19: SULPHUR SPRINGS

Sulphur Springs was a small resort where people bathed in waters from the springs and mud pots. Most buildings have fallen into ruin. The hot springs occur at the intersection of the northeast-trending Sulphur Creek fault and several cross-faults (Goff and Gardner, 1980). A variety of thermal features are visible here: fumaroles, hot springs, and mud pots. Temperatures approach boiling and pH may be less than 1. Geochemistry is discussed by Goff and others (1985). Continental Scientific Drilling Project core hole VC-2A was drilled here to a depth of 528 m in 1986 and achieved a temperature of 215°C. The core hole penetrated a thick sequence of intracaldera tuffs and volcaniclastic rocks. Retrace route to NW 4 and turn right.

On either side of the road is rhyolite lava flow of Redondo Creek Member (Smith and others, 1970), an undated post-Valles moat rhyolite flow. Poor exposures of caldera lake sediment also occur. This is the approximate position of the Valles caldera ring fracture.

Junction of NW 126 and NW 4. La Cueva settlement nestles in the caldera moat at the point where San Diego Canyon enters the caldera. The canyon head is now largely blocked by Battleship Rock ignimbrite overlain by Banco Bonito obsidian lava flow on the left, with a narrow pass cut by San Antonio Creek, which drains the western part of the caldera. Right turn immediately after NW 4 crosses San Antonio Creek into private, unmarked road.

STOP 20: VALLES CALDERA WALL; BANDELIER AND PRE-BANDELIER TUFFS

Note: This stop is again on private property. Persons following the road log on an individual basis must make their own arrangements for access. The road log continues without addition of the 3-mi round trip to the outcrop and back to NM 4. At this outcrop the Bandelier ignimbrites display erosional forms known as hoodoos, tent rocks, or teepee rocks. Found here is as complete an exposure of the ignimbrites from the Valles caldera as exists anywhere on the caldera rim. The top unit is the Upper Bandelier ignimbrite; at its base is a rare exposure of the proximal Upper Bandelier plinian (Figure 12). The main cliff-former is the Lower Bandelier Tuff. Near the top is a lithic-rich horizon marking the same level as the lithic breccias at Cat Mesa (Stop 17). Below the Lower Bandelier are two other high-silica rhyolite, nonwelded ignimbrites, informally called A and B, of late Pliocene age (2-3 Ma). ‘B’ has spectacular, coarse, bedded, proximal lithic breccia, indicating that these exposures were near the source for these ignimbrites.

Retrace route to NW 4 and turn right. Pass La Cueva campground on left. Above campground is grey welded Battleship Rock ignimbrite with cave after which La Cueva was named. Begin to enter narrow part of San Diego Canyon where it slices through the Valles caldera wall. Proceeding down canyon on left is glassy, flow-banded Banco
Madera Limestone. Massive limestone beds sit on a small cliff of Pennsylvanian Battleship Rock, note that the tuff above one of the splays of the Jemez fault.

Return to NM 4 and turn left. Ahead is a solfatara to pumiceous tuff breccia about 65 m above the Battleship Rock is a consequence of Madera Limestone and Abo Formation. The vent near El Cajete crater about 0.13 Ma. Initially, these ash-flow deposits (the Battleship Rock Member of the Valles Rhyolite) extended a considerable distance down San Diego Canyon and filled it to a depth of about 100 m, but subsequent erosion has removed all but the outcrops in the Battleship Rock area and one or two other small remnants down-canyon. Battleship Rock itself is the filling of a narrow vertical-walled gorge cut into Madera Limestone and Abo Formation. The curved columnar jointing in the lower part of Battleship Rock is a consequence of cooling against the gorge walls. Subsequent erosion has removed the adjacent, less resistant sedimentary rocks and has left the more resistant welded tuff standing as a promontory—a remarkable example of inverted topography. The tuff at Battleship Rock is about 80 m thick and contains up to six flow units that constitute a cooling unit. The tuff is entirely vitric from bottom to top. The basal 15 m are composed of poorly consolidated pumiceous ignimbrite, which becomes increasingly compacted upward and grades into partly welded tuff that has a minimum porosity of 15% at approximately 36 m above the base. The tuff becomes gradually less welded and passes again into unconsolidated pumiceous tuff breccia about 65 m above the base.

If you look at the contact zone between tuff and underlying rocks on the east side of Battleship Rock, note that the tuff site on a small cliff of Pennsylvanian Madera Limestone. Massive limestone beds are separated by thin shale partings and are extremely fossiliferous.

Return to NM 4 and turn left. Ahead is Madera Limestone at road level, overlain by Permian Abo Formation.

As you descend into San Diego Canyon, note the smell of H2S. This is due to a small solfatara to the left of the road, located above one of the splays of the Jemez fault. Between here and the next stop, get a fine view above and to the right of the sequence in the San Diego Canyon walls. The road has now left the Valles caldera and is following the fault-controlled canyon.

STOP 21: BATTLESHIP ROCK

Battleship Rock, a spectacular outcrop of columnar-jointed, rhyolitic welded tuff, was formed by a series of postcaldera small-volume ash flows that issued from a vent near El Cajete crater about 0.13 Ma. Initially, these ash-flow deposits (the Battleship Rock Member of the Valles Rhyolite) extended a considerable distance down San Diego Canyon and filled it to a depth of about 100 m, but subsequent erosion has removed all but the outcrops in the Battleship Rock area and one or two other small remnants down-canyon. Battleship Rock itself is the filling of a narrow vertical-walled gorge cut into Madera Limestone and Abo Formation. The curved columnar jointing in the lower part of Battleship Rock is a consequence of cooling against the gorge walls. Subsequent erosion has removed the adjacent, less resistant sedimentary rocks and has left the more resistant welded tuff standing as a promontory—a spectacular example of inverted topography. The tuff at Battleship Rock is about 80 m thick and contains up to six flow units that constitute a cooling unit. The tuff is entirely vitric from bottom to top. The basal 15 m are composed of poorly consolidated pumiceous ignimbrite, which becomes increasingly compacted upward and grades into partly welded tuff that has a minimum porosity of 15% at approximately 36 m above the base. The tuff becomes gradually less welded and passes again into unconsolidated pumiceous tuff breccia about 65 m above the base.

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STOP 22: SODA DAM AND JEMEZ FAULT ZONE

The travertine dam across the gorge in Precambrian granite was built by carbonated thermal waters that discharge from a strand of the Jemez fault zone. There are roughly 10 springs and seeps in this area, including one across the river to the left of the high travertine knob. About 15 years ago, water discharged along the top of the dam. Today Soda Dam is slowly falling apart.

The waters discharge at ~48°C, Cl = 1500 mg/l. The ratios of very soluble elements, Na, Li, Cl, and Br, are practically identical to those from the deep fluid within Valles caldera (Trainer, 1973; Goff and others, 1981). This and other evidence indicate that the waters here are derivatives of those from deep within Valles. Leakage and mixing occur southwest down various strands of the Jemez fault zone.

Older travertine deposits crop out high above us west of Soda Dam and include river gravels with cobbles of Bandelier Tuff. Uranium-thorium disequilibrium dates show that the Soda Dam hot-spring system has been active for the last 1 million years, which is nearly as old as caldera formation (Goff and Shevenell, 1987).

The Jemez fault zone is very complex in this area. The main trace trends north-east across the highway. Generally, the displacement is down to the east about 200-250 m because the Jemez fault zone is the westernmost of the Rio Grande rift faults. At Soda Dam, a local horst of sheared Precambrian is uplifted and contorts Paleozoic rocks all around it. If you gaze carefully at the upper east wall of San Diego Canyon, you can see a white band of Abiquiu Formation (25 Ma) overlying orange Permian shales. The Abiquiu is overlain by a sequence of volcanic units of the Valles Canyon Formation: from bottom to top, basalt flows, andesite flows, and flow breccias. The Bandelier Tuff is thin; only the Tahisreg Member covers the older volcanics.

The trip now returns towards Santa Fe, retracing the day's route.

Junction of NM 4 and Loop NM 4 at bottom of Pajarito fault scarp. Left to Los Alamos, straight ahead to Bandelier National Monument and White Rock. Go straight ahead. Road runs down surface of Upper Bandelier Tuff on the Pajarito plateau.

Subsidiary fault to Pajarito system downdrops Upper Bandelier Tuff about 30 m at large S bend in road. After bend, on left is upper part of Ancho Canyon; on right are upper reaches of Frijoles Canyon. On right skyline is St. Peter's Dome. Junction with road to Bandelier National Monument.
123.3 Road descends into Ancho Canyon.
Exposures of Upper Bandelier Tuff on top of Lower Bandelier Tuff.
129.2 Four-way junction of Pajarito Road and NM 4. Continue on NM 4 past White Rock.
134.0 END OF DAY TWO: Y-junction. NM 4 recombines; turn right towards Pojoaque and Santa Fe. Return to Santa Fe.

DAY THREE: PRE-BANDELIER RIFT SEDIMENTS AND VOLCANICS

This day examines the Pliocene Puye Formation, volcaniclastic sediments shed off the Jemez Mountains into the Española basin of the Rio Grande rift (Bailey and others, 1969; Turbeville, 1986). We will also see basaltic volcanics from the Cerros del Rio volcanic field, which mainly lies to the east of the Rio Grande, but which has a western extension buried under the Bandelier Tuffs. The Puye Formation contains valuable evidence of explosive volcanism for both the Tschicoma volcanic center and the Cerros del Rio field. Numerous pumice and scoria fall deposits, phreatomagmatic ash beds, and small pyroclastic flow deposits interbedded with the Puye sediments attest to this explosive activity. During the course of the day we will also see spectacular outcrops along the Pajarito fault zone will also be examined.

Day Three Road Log

Cumulative mileage

BEGINDAYTHREEFIELDTRIP

0 Follow route of Day 2 from Plaza de Santa Fe to US 285. Retrace route to Pojoaque; turn left on NM 4 to Los Alamos. Cross Rio Grande at Otowi bridge. Pass junction

Day Three Field Trip

Day Three Road Log

Cumulative mileage

BEGINDAYTHREEFIELDTRIP

25.6 Grey Puye Formation gravel overlies Chamita Formation (Santa Fe Group). This distal part of Puye volcanogenic alluvial fan is dominated by mudflows, fluvioglacial, and lacustrine deposits (Figure 13). Enter dirt road, PS 57, and proceed. Do not stop without permission of San Ildefonso Indian Tribe on this section of road.

28.1 Cattle guard. Begin Santa Fe National Forest. Stopping is permissible without permit.

29.4 Turn right off road onto gravel track which in 50 m makes a circle. Park. Walk to stream bed and follow (north) unnamed arroyo (about 1 hour on foot required for this stop). On west side of canyon is outcrop of Guaje plinian tuff.

STOP 23: MIDFAN PUYE FORMATION AND TSCHICOMA CENTER FAN AND FLOW DEPOSITS

This stop explores an unnamed dry arroyo that gives an excellent strike section through the mid-fan Puye Formation and consists of primary dacitic pumice fall deposits derived from explosive eruptions in the Tschicoma volcanic center, small nonwelded dacitic ignimbrites, e.g., the Puye ignimbrite, 2.5 Ma (Turbeville, 1986), and debris flows representing reworked ignimbrite and reworked talus from associated lava domes (Figure 13). Stream-channel and pumice-rich mudflow deposits are also common in this mid-fan section.

Return to FS 57/422; turn right. Junction with Rendija Canyon (FS 57 and 442). Continue ahead on FS 442.

Figure 13. Schematic sections through typical depositional cycles in Puye Formation from proximal (west) to distal (east) locations (after Waresback, 1986). Typical medial sequences will be seen at Stop 23.
31.0 Junction of FS 442 (Guaje Canyon) and FS 416. Take hairpin curve to right uphill on FS 416. Beware of oncoming traffic. Road climbs through Puye Formation, including coarse debris-flow deposits.

31.6 Top of mesa; road bears sharply left. Good view down Guaje Canyon. Continue on FS 416 for about 100 m. FS 416 bears left; continue ahead through pumice quarries. Enter pumice pits.

31.8 Stop at bar gate to Copar pumice mine. Park.

STOP 24: GUAJE PUMICE FALL DEPOSIT ON DISPERSAL AXIS OF DISTRIBUTION

Copar pumice mine. The quarry workings expose more than 8 m of lower Bandelier plinian pumice deposit (Guaje pumice bed of Bailey and others, 1969). A massive-graded fall unit (A) underlies bedded fall units B-E, which in turn are overlain by the nonwelded lower (Otowi) Bandelier ignimbrite with pumice dunes at the base. Casts of felled trees are seen at the plinian-ignimbrite contact. This exposure is on the dispersal axis of fall unit A, which has an easterly distribution. In the southern Jemez Mountains unit A was not deposited, but units B-E can be found. Turn vehicles around and retrace route to confluences of Rendija Canyon and Guaje Canyon, FS 57 and FS 442.

34.3 Turn right up Rendija Canyon, FS 57.

34.8 Stop under prominent cliffs with huge rock in Puye Formation conglomerates.

STOP 25: PUYE FORMATION: MAFIC PYROCLASTIC DEPOSITS, DEBRIS FLOWS, AND PUMICE-FALL DEPOSITS

In Rendija Canyon, the middle to upper parts of the Puye Formation are dominated by coarse debris-flow deposits, plinian deposits, thin ignimbrites, and mudflow and hyperconcentrated flood-flow deposits. Some of these are dominated by andesitic scoria and ash, suggesting that andesitic volcanism in the Tschicoma center persisted until about 3 Ma. The steep walls of conglomerates, both debris-flow deposits and coarse stream gravels, erode into pillars. Proceed up Rendija Canyon on FS 57.

37.6 Pass Sportsmans Club. You are now on the surface of the Puye Fan, with Bandelier Tuff forming cliffs above.

37.8 Pull onto track (unmarked) on right side of road. Park vehicles and walk north for about 40 m, following small path down slope into stream. Cross stream bed and head for prominent cliffs to north (straight ahead).

STOP 26: UPPER BANDELIER TUFF SURGE BEDS, TOLEDO PYROCLASTIC DEPOSITS, GUAJE MOUNTAIN DACITE, AND YOUNG FAULTING ALONG PAJARITO ZONE

A short walk northward leads to a mesa of upper Bandelier ignimbrite overlying Toledo (1.4 to 1.2 Ma) pyroclastic deposits (Heiken and others, 1986). The Toledo deposits are plinian pumice falls representing explosive eruptions that accompanied lava-dome growth between the two Bandelier eruptions. The related lava domes for these fall deposits fill the Toledo embayment. At the base of the UBT the plinian deposit is about 1 m thick and is overlain by several meters of fine-grained, spectacularly cross-bedded surge deposits. These in turn are overlain by the nonwelded base of the UBT. Proceed up Rendija Canyon.

Surfaced road begins. Road immediately dips into small valley eroded along fault. Steep climb up through Upper Bandelier Tuff. Junction with Barranca Road, Los Alamos. Turn right and follow main road around curve to left.

Four-way junction. Turn right onto Diamond Drive; pass Los Alamos Golf Course. Pass Stop 14 of Day 2 by Baptist Church. Retrace route of Day 2 to junction of West Jemez Road and NM 4. Turn left onto NM 4, signedpost Bandelier National Monument and White Rock.

Junction with road into Bandelier National Monument. As a pleasant side trip we recommend a visit to the monument to see the Indian ruins hewn from Bandelier Tuff. Return to NM 4; turn right. Continue to White Rock and follow NM 4 to T-junction. Take right to return towards Santa Fe. Turn off NM 4 into wide parking area on right-hand side.

STOP 27: BASALTIC VOLCANISM OF CERROS DEL RIO FIELD

The basalt underlying the Bandelier Tuff at this locality erupted from a vent in the gorge immediately south of the road (age 2.4 Ma). The flow grades downward into pillow-palagonite breccia, which displays forest bedding—an indication that the flow spread eastward from the vent into a lake that probably formed by damming of the Rio Grande. About 200 m down the highway and to the east, water-laid basaltic ash and lacustrine clays underlie the palagonite breccia; the ash and clays provide further evidence of eruption into a former lake. Farther down the road, a tongue of basalt at the toe of the flow has injected basalt into sand and gravel beds, causing intense deformation of the sand and gravel. On the north side of this tongue, the basalt was in steep contact with ripple-marked sediments that have since been stripped from the contact, exposing a cast of the ripple marks in the basalt surface.

END OF DAY THREE: Return to Sante Fe using route described at end of Day Two.


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Stratigraphy, Correlation, and Tectonic Setting of Late Cretaceous Rocks in the Kaiparowits and Black Mesa Basins

INTRODUCTION

The purpose of this field trip is to examine the Late Cretaceous (Cenomanian through Campanian) marine and nonmarine strata deposited near the western margin of the Western Interior Seaway in the area of the southwestern Colorado Plateau (Figures 1 and 2). The Dakota, Tropic Shale, Straight Cliffs, Wahwes, and Kaiparowits Formations in the Kaiparowits Plateau in southern Utah will be examined and contrasted with the Dakota, Mancos Shale, Toreva and Wepo Formations, and Rough Rock and Yale Point Sandstones at Black Mesa in northeastern Arizona. These sequences provide information concerning similarities and local variations in tectonic and eustatic influences and the tools for detailed comparisons with the better known equivalents to the east and northeast. Cretaceous strata record a complex interplay among tectonic uplift, foreland basin subsidence, and several major tectonoeustatic transgressive-regressive cycles.

GEOLOGIC SETTING AND HISTORY

The Cretaceous strata of the southwestern Colorado Plateau were deposited in a basin bounded by the Sevier orogenic belt of western Utah and the Mogollon Highlands of southern and central Arizona (Molenar, 1983). These elevated areas, in association with reactivation of basement structures (Peterson, 1969a), provided the tectonic controls on deposition in this region and formed a V-shaped embayment at peak transgression (“Grand Canyon Bight” of Stokes and Heylman, 1963).

Folding and eastward-directed decollement-style thrusting along the north-northeast-trending Sevier orogenic belt, which lies in southeastern Nevada and western Utah (Figure 3), occurred during the late Early Cretaceous through the Paleocene (Armstrong, 1968). A tectonically induced foreland basin formed immediately adjacent to and east of this active source terrain into which thousands of meters of nonmarine and marine strata were deposited.

Harbaugh and others (1957), Drewes (1981), and Bilodeau (1986) have proposed that uplift occurred in the Mogollon Highlands (centered in central or southwestern Arizona; Figure 3) and erosion of northeastward-dipping, pre-Cretaceous strata occurred prior to the deposition of the Late Cretaceous (Cenomanian) Dakota Formation. Gravels deposited at the base of the Dakota were primarily derived from Paleozoic strata. By the late Cenomanian and through the late Turonian, arkosic sediments were shed from the south, whereas the Sevier orogenic belt continued to supply lithic sediments eroded from Paleozoic rocks to the west.

The other dominant influence on deposition in the
area was eustatic sea-level fluctuations, particularly the large-scale cycles corresponding to Kaufmnan's (1977) Greenhorn, Niobrara, Craggett, and Bearpaw cyclothems. The Greenhorn cyclothem, represented by the Dakota Formation, Tropic Shale, and Tibbet Canyon Member of the Straight Cliffs Formation in the Kai­parowits Plateau and the Dakota Formation, Mancos Shale, and Toreva Formation at Black Mesa (Figure 4), spanned the Cenomanian through middle Turonian. During the transgressive phase, the middle, nonmarine member of the Dakota Formation aggraded on a subsiding allu­vial plain (Figure 4). The shoreline transgressed beyond the western margin of the Kolob Terrace and probably just southwest of the present-day Mogollon Rim. During the late Cenomanian, prior to peak transgression, rapid subsidence within the Sevier foreland basin (whose axis lies in the present-day Kolob Terrace) preserved at least seven progradational shoreface sequences in the upper Dakota Formation of Utah (Figure 5). Similarly, several late Cenomanian stacked shoreface sequences form the eroded top of the Cretaceous section along the western Mogollon Rim. Peak transgression occurred in the early Turonian (Lower Hamsites podosoides Zone) when stacked shoreface sequences, included in the Straight Cliffs Formation, continued to be deposited on the western margin of the seaway (Figure 4). The lower part of the Tropic and Mancos Shales are correlative to these shoreface units. At peak Greenhorn transgression the shoreline paralleled the structural highlands in Utah and Arizona (Molenaar, 1983; Cobban and Hook, 1984), forming a V-shaped embayment of the seaway, which probably amplified tidal waves and affected local marine circulation during deposition of the upper part of the Dakota and lower Straight Cliffs Formations. During the lower Collignoniceras woollgari Zone, foreland basin subsidence slowed, sea level fell, and the shoreline rapidly retreated northeastward in a steplike fashion until peak regression occurred east of the Kaiparowits and Black Mesa basins during the middle Turonian Prionocyclus hyatti Zone. Regres­sive shoreface sands are represented by the Tibbet Canyon Member of the Straight Cliffs Formation in the Kaiparowits Basin and the lower sandstone member of the Toreva Formation in Black Mesa Basin. These shore­face sands intertongue with and are overlain by the nonmarine Smoky Hollow Member of the Straight Cliffs Formation in the Kaiparowits Basin and the middle carbonaceous member of the Toreva Formation in the Black Mesa Basin. These members have swamp deposits at their bases that grade upward into fluvial meander­belt deposits and are capped by the braided stream conglomeratic sandstone of the Calico bed in the Kaiparowits region and the arkosic, upper sandstone member of the Toreva Formation in Black Mesa Basin. The unconformity at the base of the upper Toreva sandstone and at the top of the Calico bed (Peterson, 1969a, b) represents the ravinement between the Greenhorn and Niobrara cyclothems and can be correlated throughout the western interior (Arizona to Montana).

The Niobrara cyclothem is represented in the Kaiparowits Plateau by the John Henry Member of the Straight Cliffs Formation and in the Black Mesa region by the upper sandstone member of the Toreva Formation through the Yale Point Sandstone. During maximum transgression of the Niobrara cyclothem, marine deposits were stacked on the east side of the Kaiparowits Plateau and the northeast side of Black Mesa. The earliest (Coniacian) marine deposits of the Niobrara cyclothem at Black Mesa are represented by the Wind Rock Tongue of the Mancos Shale and in the Kaiparowits Plateau by the lower marine mudstone tongue of the John Henry Member. Rocks younger than

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Figure 2. Correlation of Upper Cretaceous sequences present on the Kaiparowits Plateau and at Black Mesa. Field-trip stops are indicated to the right of the thickness columns.

Figure 3. Cretaceous tectonic pattern in the Colorado Plateau, including large-scale folds and orogenic belts. Modified after Cooley and others (1969).
the Santonian Yale Point Sandstone were removed by Cenozoic erosion in the Black Mesa area. The near equilibrium of eustacy and subsidence led to the development of thick coal deposits landward of the stacked shoreline deposits.

The Drip Tank Member of the Straight Cliffs Formation in the Kaiparowits region represents the regressive phase between the Niobrara and Claggett cycloths. The overlying Wahweap Formation consists dominantly of meander-belt and floodplain deposits formed during the early Campanian Clagget cyclothem. During this phase the shoreline probably was somewhere east of the Henry Basin as indicated by the time-equivalent nonmarine Masuk Member of the Mancos Shale.
INTERTONGUING DAKOTA FORMATION AND TROPIC SHALE

Figure 5. West–east cross section from the western Kolob Terrace across the southern Kaiparowits Plateau. Offshore marine deposits of the Tropic Shale intertongue westward with progradational strand-plain sequences of the Dakota Formation.

in that area. The late Campanian portion of the Bearpaw cyclothem is represented by the Kaiparowits Formation. This thick (850 m) sequence of floodplain deposits was folded, truncated, and then unconformably overlain by conglomerates of the latest Campanian Canaan Peak Formation. This unconformity occurred near peak transgression, and the tectonically controlled sedimentation in the Kaiparowits Basin shows no obvious effect of the seaway, which is well to the east.

FIELD-TRIP GUIDE

The field trip begins along the northern part of the Kaiparowits Plateau, progresses southward along the west flank of the plateau, and then moves to the north end of Black Mesa (Figure 1).

Stop 1
Cannonville Area - 1 mile west of Cannonville on Highway 54; bluffs on north and south sides of the highway.

The Dakota Formation can be divided into three distinct members based on lithology. The lower member is composed predominantly of clast-supported, granule and pebble conglomerate, deposited by gravelly braided streams within previously eroded, southeast-trending paleovalleys. The middle member is composed of interstratified sandstone, mudstone, claystone, and coal, arranged in upward-fining, meander-belt and floodplain environments. The upper member is composed of interstratified claystone, mudstone, siltstone, and sandstone arranged in upward-coarsening, claystone-to-sandstone sequences, capped by rooted carbonaceous mudstone or by intensely bioturbated, fossiliferous, sandstone beds. The upper member was deposited by progradational shorelines during the overall transgressive phase of the Greenhorn cyclothem.

In the Cannonville area, the Jurassic-Cretaceous contact is conspicuous as drab, gray-buff lithologies of the Dakota overlie white and red lithologies of the Entrada Sandstone. Within the middle member at least seven, upward-fining, meander-belt and floodplain sequences can be traced throughout the area. The basal contact is erosional and each sequence or cyclothem typically contains numerous, laterally isolated, lenticular to discontinuous channel-sandstone bodies, or lithosomes. Where sandstone lithosomes are present, they consist of, in sequential order, intraformational conglomerate, cross-stratified and ripple cross-laminated sandstone, mudstone, carbonaceous claystone, and coal. Epsilon crossbedding, representing lateral migration of point bars, is common. The fact that these cyclothems can be correlated regionally suggests an allocyclic control (climate, sea-level fluctuation, or basin subsidence) rather than autocyclic meander-belt avulsion or temporally equivalent delta-locus switching. Many vertebrate fossil localities have been discovered in the middle member in this area and have produced the remains of fish, turtles, crocodiles, and dinosaurs.

Stop 2
Pardner Canyon - 4 miles east of Henrieville; canyon northeast of Highway 54.

Pardner Canyon provides excellent exposures of the Tropic Shale and overlying Straight Cliffs Formation (Figure 6). The Tropic-Straight Cliffs contact is placed at the lowest readily traceable sandstone bed overlying the marine shales. The lowest member of the Straight Cliffs Formation, the Tibbet Canyon, represents the regressing shoreline of the Greenhorn cyclo-
Figure 6. Tropic Shale and Straight Cliffs Formation on the west side of Pardner Canyon (Stop 2). Kt-Tropic Shale; Straight Cliffs Formation: Ksc-Tibbet Canyon Member; Kscs-Smoky Hollow Member; Kscsc-Calico bed, Smoky Hollow Member; Kscj-John Henry Member.

them. The lower part of the overlying Smoky Hollow Member consists of coals, carbonaceous mudstones, greenish-gray mudstones, and sandstones that accumulated in swamps and marshes. The middle part of the member (the barren zone of Peterson, 1969b) consists of bentonitic mudstones and sandstone deposited on alluvial plains by meandering streams. Terrestrial vertebrates have been recovered from this unit, including the first Turonian mammals known from this hemisphere. The unit is overlain by the bleached, white, pebbly sandstones of the Calico bed deposited by braided streams. An unconformity spanning the late Turonian through early Coniacian is present between the Calico bed and the base of the overlying John Henry Member (Peterson, 1969a, b). Ryer (this paper, Stop 7) suggests that an unconformity may also be present at the base of the Calico. The Calico bed thickens to more than 100 m to the northeast (north of Escalante) and is thinner to the south, reflecting lack of deposition or post-Calico erosion to the south and greater subsidence to the north. The John Henry Member in this area consists of coals, mudstone, carbonaceous shales, and sandstones deposited in lagoonal to coastal plain environments, and fossils indicate a distinct brackish influence. This member has produced abundant vertebrate fossils including the first record of Coniacian and Santonian mammals in this hemisphere. The John Henry Member is predominantly marine to the east (Stop 7). The dramatic sandstone cliff in Pardner Canyon is formed by the Drip Tank Member of the Straight Cliffs Formation, which consists of medium-grained sandstone with quartz and chert pebble conglomerates deposited in a predominantly braided-stream system (Peterson, 1969a, b). This sandstone is thickest in the Tropic area (160 m). Some exposures of the overlying Wahweap Formation can be seen farther up the canyon.

Stop 3

East of Pardner Canyon - 5 miles east of Henrieville; cliff southeast of Highway 54.

Here, the braided-stream deposits of the Drip Tank Member are more than 100 m thick and form a sheer cliff along Henrieville Creek (Figure 7). The base of the overlying Wahweap Formation is marked by the first laterally continuous mudstone unit. The Wahweap Formation is about 450 m thick in this area and consists predominantly of sandstones, mudstones, and carbonaceous shales deposited on floodplains by meandering rivers. Unlike the underlying Straight Cliffs Formation, there appear to be no marine or brackish deposits. The Wahweap (below its capping sandstone) correlates to the Maisuk Member of the Mancos Shale to the east in the Henry Basin. Both units have produced abundant vertebrate fossils, including mammals, that support an early Campanian age for these formations based on comparisons with the Milk River faunas of Canada (Eaton, unpublished data).

Stop 4

Henrieville Creek - 8 miles east of Henrieville, area of gravel pit north of Highway 54 where Henrieville Creek crosses the highway; exposures on east facing cliff.

The capping sandstone of the Wahweap Formation is more than 100 m thick here and contains chert pebble conglomerates. Most of this unit was deposited by braided streams. The base of the Kaiparowits Formation is placed at the first laterally persistent mudstone (or laterally equivalent sandstone). Generally, this contact is marked by a distinct change in slope as the much less resistant Kaiparowits forms a bench on top of the thick, resistant, capping sandstone of the Wahweap, but at this location the contact occurs on the cliff face (Figure 8). The lower part of the Kaiparowits is dominated by fluvial sandstones and...
dark greenish-gray, silty, alluvial-plain mudstones that are less fossiliferous than the rest of the formation.

Stop 5

The Blues Overview - 10 miles east of Henrieville; pull off north of Highway 54.

The Kaiparowits Formation is more than 850 m thick in this area and is composed of about equal proportions of mudstone and sandstone (Figure 9). The sandstones are very friable and sedimentary features are often difficult to detect. The distinctive gray arkosic sandstones, volcanic rock fragments, and lack of chert clasts throughout the sequence suggest source rocks different from those of the Straight Cliffs and Wahweap Formations. The Kaiparowits was deposited by meandering streams on a relatively well-drained floodplain, as indicated by the rarity of carbonaceous shales as compared to the underlying Cretaceous formations. The Kaiparowits is richly fossiliferous in this area and has produced abundant remains of turtles, fish, crocodiles, dinosaurs, mammals, freshwater molluscs, and fragmentary egg shells.

Stop 6

Top of South Rim - bench above The Blues; 13 miles east of Henrieville on Highway 54.

The uppermost Kaiparowits Formation is preserved in the area around Canaan Peak, where it is overlain unconformably by the latest Campanian through Paleocene (?) (Bowers, 1972, based on palynomorphs) conglomerates of the Canaan Peak Formation. The Kaiparowits Formation was strongly folded into an elongate north-south trending syncline and deeply eroded prior to deposition of the Canaan Peak Formation. If the late Campanian date is correct for the lower part of the Canaan Peak Formation, the folding and erosional truncation of the Kaiparowits Formation must have been a rapid event.

The route between stops 6 and 7 crosses the east flank of the Table Cliff's syncline and the Dutton monocline and passes through steeply dipping beds of the Wahweap Formation and the more gently dipping beds of the Straight Cliffs Formation.

Stop 7

West of Escalante - 5 miles west of Escalante on Highway 54.

The lower part of the Straight Cliffs Formation near Escalante includes strata equivalent to the Ferron Sandstone Member of the Mancos Shale in central Utah and the Frontier Formation of northeastern Utah. These rocks constitute a widespread clastic wedge that defines the boundary between the Greenhorn and Niobrara cycles. The outcrops west of Escalante (Figure 10) provide critical evidence bearing on the interpretation of this clastic wedge.

Conformably overlying the Tropic Shale, the Tibbet Canyon Member of the Straight Cliffs Formation records eastward regression of the Greenhorn Sea during middle Turonian time. The great thickness of the Tibbet Canyon Member, and westward intertonguing of the Tibbet Canyon middle and upper shoreface deposits with carbonaceous shale and coal in the basal part of the Smoky Hollow Member, demonstrate that relative sea level continued to rise in this area during regression.

Mudstone, carbonaceous shale, and coal of the lower Smoky Hollow Member were deposited in deltaplain and alluvial-plain environments. The cross-bedded, pebbly Calico sandstone forms the upper part

Figure 9. Typical exposures of the Kaiparowits Formation in The Blues (Stop 5).

Figure 10. Straight Cliffs Formation at Stop 7. Ksct-Tibbet Canyon Member; Kscs-Smoky Hollow Member; Kscsc-Calico bed, Smoky Hollow Member; Kscj-John Henry Member.
of the member and was deposited by braided streams. The contact at the base of the Calico bed is sharp at this locality.

The contrasting styles of sedimentation in the lower and upper parts of the Smoky Hollow Member can be interpreted in terms of base level which, for marine and brackish-water facies, equates with relative sea level. The lower part of the member was deposited under conditions of rising relative sea level and represents a continuation of the rising base level that prevailed during deposition of the Tibbet Canyon Member. The Calico bed accumulated under conditions of gradually rising base level. The abrupt contact between the two units may represent an unconformity produced by fluvial erosion during a lowering of base level.

Deposition of the Calico bed was terminated by transgression of the sea in Coniacian time. The fact that the Calico is separated from the overlying middle shoreface strata of the John Henry Member by only a thin transgressive lag indicates that the transgression was rapid. Further, that the transgression was the result of rapid rising of relative sea level.

Relative sea level is a combination of two independent variables, eustatic sea level and subsidence. Stratigraphic relationships elsewhere in Utah, particularly in the Ferron Sandstone in central Utah and the Frontier Formation in northeastern and north-central Utah, suggest that the relative sea-level/base-level curve reconstructed for this outcrop is not a simple eustatic sea-level curve, but records a major event of tectonic uplift in areas to the west. The tectonic event occurred during late Turonian time, when eustatic sea level was rising, resulting in submergence of distal Ferron-Frontier deposits. The tectonic event resulted in rapid rising of relative sea level.

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vertebrates known anywhere in the world.

Stop 12

Wahweap Creek Overlook - 1/2 mile east of Big Water, where Smoky Mountain road crosses Wahweap Creek.

The thin yellow- to buff-colored, ridge-forming, upper sandstone member of the Dakota Formation (Figure 12) represents the seaward feather edge of the two upward-coarsening shoreface sequences examined at Stop 10. The ridge is composed of an intensely bioturbated, fossiliferous sandstone, interpreted as a transgressive lag. The lower part of the overlying Tropic Shale contains numerous fossiliferous, limestone concretion horizons. Detailed lithostratigraphic, biostratigraphic, and chronostratigraphic (bentonites) correlations have revealed that each concretion horizon can be traced into transgressive lags overlying progradational strand-plain deposits (Figure 5). Stacked shoreline sequences in the Kolob Terrace suggest greater relative subsidence to the west.

Figure 12. Dakota Formation (Kd), Tropic Shale (Kt), and Tibbet Canyon Member of the Straight Cliffs Formation (Ksct) at Stop 12.

Stop 13

Tropic Shale Section East of Big Water - 4 miles east of Big Water on road to Smoky Mountain.

The marine strata of the Greenhorn cyclothem are represented by the Tropic Shale in southern Utah (Figure 13). Coals of the middle member of the Dakota Formation are overlain by sandy shale containing a brackish water fauna capped by about 2 meters of interbedded, rippled sandstone-shale, which locally forms the top of the Dakota. The basal few meters of Tropic Shale consist of noncalcareous shale. The feather edge of the shoreface sequence seen at Stop 12 correlates with the rather abrupt transition to calcareous shale here, marking the onset of open marine deposition of the Greenhorn cyclothem.

At the base of the calcareous shale is a sequence of three major and several minor bentonite beds representing the first of a series of chronostratigraphic marker beds, tied to molluscan biozones, that can be correlated with great precision across most of the Western Interior Seaway (Pratt and others, 1985; Figure 14). Overlying this first sequence of bentonites are two limestone concretion horizons, containing a very diverse, well-preserved Sciponoceras gracile Zone fauna, from which Koch (1978) was able to identify 170 taxa.

Three and one-half meters above the Sciponoceras concretions lies one of the most widely recognized bentonites in the Western Interior, referred to as PBC-11 by Elder and Kirkland (1985; see that paper for equivalent terminologies). This bentonite ranges from 20 to 50 cm thick across the Colorado Plateau and always is associated with a 1-cm-thick bentonite a few centimeters below. It is generally referred to as the "Neocard" bentonite in this area for the zonal index fossil Neocardioceras juddii. PBC-11 serves as a primary datum for Greenhorn cyclothem in this region (Figures 4, 5, and 14). The Cenomanian/Turonian boundary occurs a few meters above this datum.

Peak transgression is recorded between marker beds PBC-17 and PBC-30 in the lower part of the lower Turonian Mammites nodosoides Zone (Figure 14), and is marked by the presence of marlstones grading eastward into limestones.

About halfway up the Tropic Shale, at the base of the middle Turonian Collignoniceras woolgari Zone, there is a bentonite swarm consisting of several meters of alternating shale and bentonite. By this time the shoreline to the west had begun to prograde eastward (Figure 4).

The lower half of the Tropic Shale in the Kaiparowits Plateau can be precisely correlated, on the basis of its numerous marker beds, with the lower third of the Mancos Shale at Black Mesa. Both of these areas had a remarkably similar history during the late Cenomanian and Turonian.

The contact with the Tibbet Canyon Member of the Straight Cliffs Formation at the top of the Tropic Shale is typical of a wave-dominated coast line. The conglomeratic white Calico bed caps the overlying Smoky Hollow Member of the Straight Cliffs Formation and marks the end of the Greenhorn cyclothem in this area.

The route between stops 13 and 14 follows Highway 98 from Page to Black Mesa and crosses Jurassic strata exposed on the Kaibito Saddle, which separates the Kaiparowits and Black Mesa basins. At White Mesa the Dakota Formation overlies the Cow Springs Member of the Entrada Sandstone.
Stop 14

Coal Chute - On Highway 160, just northeast of the entrance to Navajo National Monument.

The Late Cretaceous strata at this locality are remarkably similar to those of the Greenhorn cyclothem in the Kaiparowits Basin. We will examine strata from the base of the Dakota up through the lower one-third of the Mancos Shale. All three members of the Dakota Formation are well-developed here. A bioturbated, transgressive lag is present at the top of the Dakota Formation (Figure 15). Brackish water deposits underlying this lag are only preserved in structural lows, and intertonguing of the Dakota and Mancos has not been observed at Black Mesa. This may indicate that the Black Mesa Basin was not subsiding as rapidly as the Kaiparowits, and subtle tectonic movements during transgression were controlling the preservation of the marginal marine sediments (Kirkland, 1983; Fürsich and Kirkland, 1986).

Though it is thinner here, the lower half of the Mancos Shale can be correlated with the Tropic Shale in the Kaiparowits area almost bed for bed based on biostratigraphy and marker beds (Figures 14 and 16). Two-thirds of the way up in the Mancos section the shales become noncalcareous and, for approximately the next 10 m, the slope is covered with thin rippled sandstone slabs containing the first occurrence of Collignoniceras woollgari regulari. This sequence of thin sandstones and shales can be correlated all around Black Mesa and is referred to as the Hopi sandy member by Kirkland (1983; Figures 4 and 17; equivalent to the middle sandy member of Repenning and Page, 1956). This member is not recognized in the Kaiparowits region. The sandstones in this sequence thicken in the southwestern part of Black Mesa where they form a major break in the Mancos Shale slope (Kirkland, 1983). This member also records a peak in the offshore deposition of terrestrial plant debris. The source of the Hopi sandy member is southwest of Black Mesa and the unit represents prodelta storm deposits.

The bentonite swam at the base of the Collignoniceras woollgari woollgari Subzone, together with the Hopi sandy member, serves to divide the Mancos Shale at Black Mesa into four informal members: (1) a lower fossiliferous calcareous shale member (to the top of the bentonite swarm); (2) a middle well-laminated

Figure 14. Event correlation of the lower Mancos and Tropic Shales with the standard reference section of the Greenhorn cyclothem from near Pueblo, Colorado. Marker beds numbered after Elder and Kirkland, 1985. Zone numbers indicate: 1-Metiococeras mosbyense; 2-Sciponoceras gracile; 3-Neocardioceras juddii; 4-Watinoceras coloradoense; 5-Mammites nodosoides; 6a-Collignoniceras woollgari woollgari; 6b-C. w. regulari; and 7-Priocyclocus hyatti.
calcereous shale member; (3) the Hopi sandy member; and (4) an upper noncalcereous claystone member that is almost devoid of calcereous fossils and thins and becomes siltier to the southwest (Figure 17). A prominent tongue (the Toreva tongue of Repenning and Page, 1956; Blue Point tongue of the Toreva Formation, this paper) of the Toreva Formation at Blue Point, in the area of southwesternmost Black Mesa, contains a very diverse nearshore marine fauna which includes C. woolligeri regulari, and fossils collected from the Mancos-Toreva transition include Inoceramus lasamucki, I. flaccidus, and C. woolligeri regulari (Kauffman, pers. comm., 1987?) indicating a position high in the C. woolligeri regulari Subzone. Fossils indicative of the younger Prionocyclus hyatti Zone (I. howelli, I. flaccidus, Mytiloides codellana, Lopha bellaplicata bellaplicata, Prionocyclus hyatti; Cobban and Kauffman, pers. comm., 1987) have been found in the transition with the overlying Toreva Formation only along the northeast side of Black Mesa. The lower sandstone member of the Toreva Formation represents a wave-dominated regressive coastline. At this stop the shoreline was oriented nearly perpendicular to the trend of the outcrop, and accretion surfaces can be seen dipping to the north-northeast.

The late Cenomanian through middle Turonian Mancos Shale at Black Mesa is very similar to the Tropic Shale in the Kaiparowits region (Figure 4), whereas the type Mancos Shale in the San Juan Basin spans the late Cenomanian through middle Campanian, recording several major marine cycles (Pike, 1947; Molenaar, 1983).

Navajo Highway 41 Roadcut - at crest of cliff, northwest side of Black Mesa, east of Navajo National Monument.

At this stop we will examine the Toreva Formation. The lower sandstone member of the Toreva, exposed on the lower part of the cliff above the Mancos Shale, was deposited as the Greenhorn sea regressed northeast across Black Mesa during the middle Turonian Collignoniceras woolligeri regulari and Prionocyclus hyatti Zones (Figure 4). Fossils from this area were identified in Repenning and Page (1956) as Inoceramus dimidius of the Prionocyclus macombi Zone. Large volumes of feldspathic sediment discharged from delta distributary channels were redistributed by high-energy waves and longshore currents to form extensive, elongate, northwesterly trending sand bodies oriented parallel to the coastline. The middle carbonaceous member of the Toreva is more than 30 m thick at its type area to the south, but is thin and poorly developed in the area of northern Black Mesa, and may locally be cut by the base of the coarse, arkosic, upper sandstone member of the Toreva Formation (Figure 17). Sandstones of this member form the rim at the top of the road cut. The sandstone beds at the base of the upper sandstone member have sheetlike geometries, abundant tabular planar cross-stratification, and no consistent vertical variations in grain size. These characteristics and paleocurrent measurements of the sandstone beds indicate deposition in northeasterly flowing braided and bed-load meandering rivers. A vertical change to more lenticular channel-fill deposits a decrease in grain size, and an increase in overbank deposits, in the upper part of the member indicate a change to meandering rivers with a lower depositional gradient.

The Coal Chute section is near the landward pinchout of the Wind Rock Tongue of the Mancos Shale and Rough Rock Sandstone. As a result, the lower and upper carbonaceous members of the Wepo Formation join to form the undifferentiated Wepo Formation that caps Black Mesa in this area.

The route north from Stop 15 follows the Organ Rock monoclone northward along exposures of Triassic and Jurassic strata. The prominent hogback southeast of the highway is formed by the lower member of the Dakota Formation. Near Stop 14 the underlying thick, white, crossbedded Cow Springs Member of the Entrada Sandstone (Peterson, in press) is visible in the lower part of the hogback. Here the Morrison is thin, although it thickens rapidly to the northeast. A few
miles southwest of Stop 14, the Dakota directly overlies the Cow Springs Member. The amount of strata missing under the pre-Dakota unconformity increases to the south, and, in the area of the Mogollon Rim, Upper Cretaceous strata rest on Pennsylvanian rocks. Farther south at Deer Creek, southeast of Phoenix, Upper Cretaceous marginal marine and terrestrial sediments are intimately associated with volcanics and rest on Pennsylvanian strata, reflecting the margin of the Late Cretaceous marine embayment.

Stop 16
Chilchinbito - just west of Chilchinbito on Highway 59.

To the south at Crevise Point, the entire Cretaceous section at Black Mesa is exposed. The resistant, cliff-forming, braided-stream sandstones of the lower member of the Dakota Formation unconformably overlie a thick section of the Late Jurassic Morrison Formation. The lower member of the Dakota Formation here has produced fossils of dinosaurs, crocodilians, turtles, and fish. The middle and upper members, together with the Mancos Shale, form the broad slope leading up to the cliff-forming sandstones of the Toreva Formation (Figure 18). The coarse, arkosic upper sandstone member of the Toreva Formation rests on an erosional surface cut into the middle carbonaceous member of the Toreva Formation. This surface, formed as base level lowered during eustatic sea-level drop, a tectonic event, or both, represents the unconformity between the Greenhorn and Niobrara cyclothsms. The subsequent base-level rise is reflected in the changing depositional environments from braided to mix-load meandering fluvial systems through the upper sandstone member and from lower alluvial plain to coastal plain through the overlying lower carbonaceous member of the Wepo Formation. After transgression of the sea into northern Black Mesa, sediments deposited in offshore and marginal marine environments are preserved as the Wind Rock Tongue of the Mancos Shale and the Rough Rock Sandstone, which form the broad slope leading up to the cliff-forming sandstones of the Toreva Formation (Figure 18). These units were considered early Santonian (Molenaar, 1983), but new collections of inoceramids from this interval (Kirkland, unpublished data) date this sequence as latest Turonian or Coniacian based on the occurrence of Mytiloides feigei in the lower part of the Wind Rock Tongue. A second minor slope is formed by the paludal and floodplain deposits of the upper carbonaceous member of the Wepo Formation. The large upper cliff is formed by the late Santonian marginal marine Yale Point Sandstone. Thus the stacked Niobrara beach sands along the northern side of Black Mesa represent a depositional setting correlated to the stacked beach sands on the east side of the Kaiparowits Plateau. This stacking led to the accumulation of the thick coal deposits on the delta plain behind them (Garr, 1987). These coals are under development at the Kayenta and Black Mesa mines near the north side of Black Mesa.

Repenning and Page (1956) originally included the entire sequence above the Mancos Shale through the Rough Rock Sandstone in the Toreva Formation. Subsequently Franczyk (in press) restricted the Toreva in northern Black Mesa to the strata that correlate with the Toreva in its type area in southern Black Mesa (Figure 17).

Because the type Mesaverde Formation of southern Colorado is Campanian in age and represents a post Niobraran regression, and the Mesaverde Group as used elsewhere is younger than the sequence at Black Mesa, the Mesaverde Group on Black Mesa should be renamed or its use discontinued.

Stop 17
Rough Rock - exposures west of Rough Rock.

The road from Rough Rock to the top of Black Mesa provides a rare, easy access through the "Mesaverde Group." The Toreva Formation is partly covered along the road cut, but excellent exposures occur to the east at Yale Point. The lower sandstone member of the Toreva Formation in the northeastern part of Black Mesa generally consists of two to three stacked delta-front sandstone units separated by shale and siltstone beds. An extensive distributary channel system existed in the Rough Rock area and resulted in distributary channel deposits erodingly overlying thin delta-front sandstones at the base of the member. The upper sandstone member of the Toreva Formation is locally well-developed and at Yale Point rests unconformably on the lower sandstone member (Figure 19). Channel deposits in the upper part of the upper sandstone member are locally uranium mineralized. The northeast-trending Rough Rock Sandstone consists of thick shore-
face deposits locally capped by extensive tidal-channel and tidal-inlet deposits.

The predominantly fluvial upper carbonaceous member of the Wepo Formation conformably overlies the nearshore marine Rough Rock Sandstone and is unconformably overlain by the nearshore Yale Point Sandstone (Figure 18). The upper carbonaceous member consists of a framework of sheet and ribbon scour-based sandstones encased in an interbedded sequence of siltstone, mudstone, claystone, and coal (Figure 20).

Six sedimentary facies have been recently recognized within the upper carbonaceous member in the northeastern part of Black Mesa (Carr, 1987): 1) scour-based, trough-crossbedded sandstones interpreted as fluvial and distributary channel deposits; 2) heterolithologic sequences of sandstone, siltstone, and mudstone associated with trough-crossbedded sandstones interpreted as levee deposits; 3) sharp-based, ripple-laminated sandstones and siltstones interpreted as crevasse splay deposits; 4) rooted mudrocks interpreted as well-drained marsh and backswamp deposits; 5) coals and carbonaceous shales with rooted substrates interpreted as poorly drained marsh and backswamp deposits; and 6) burrowed sandstones and mudrocks interpreted as interfluvial lake and interdistributary bay deposits.

Facies distributions and relationships suggest that the upper carbonaceous member can be subdivided for the purpose of analysis into lower, middle, and upper stratigraphic intervals (Carr, 1987).

Within the lower interval, trough-crossbedded sandstones are encased in interbedded sequences of finer-grained facies dominated by rooted mudrocks. Coals tend to be laterally discontinuous and thin. This interval is interpreted as a deltaic plain. Distributary channels were flanked by poorly developed levees, which graded into interdistributary areas occupied by poorly drained marshes and backswamps and brackish-water bays.

The middle interval consists of thick, laterally extensive, trough-crossbedded sandstones that grade eastward into thinner, laterally restricted sandstones encased in finer-grained facies dominated by rooted mudrocks and coals. Coals are thicker and more continuous than those in the lower interval. This interval is interpreted as an alluvial plain characterized by well-drained backswamps. Fluvial channels were flanked by well-developed levees, which graded into flood basins dominated by well- to poorly drained backswamps.

Within the lower part of the upper interval, trough-crossbedded sandstones are relatively rare; in the upper part they are virtually absent. The lowermost part of the interval is dominated by thick, laterally extensive coals which decrease upward in abundance and thickness. Burrowed sandstones and mudrocks increase upward and dominate the upper part of the interval. The lower part of this interval is interpreted as an alluvial plain characterized by diminishing fluvial activity and extensive interfluvial areas. Narrow, laterally stable fluvial channels were flanked by levees that graded into flood basins dominated initially by well- to poorly drained backswamps and later by interfluvial lakes. The uppermost part of the interval is interpreted as a marginal marine coastal plain with brackish-water bays grading landward into poorly drained, brackish-water marshes.

The upper carbonaceous member of the Wepo Formation represents the landward segment of a regressive-transgressive depositional cycle, which occurred along the southwestern margin of the western interior foreland basin during late Coniacian into the Santonian. The vertical succession of depositional environments within the lower and middle intervals records a regressive fluvial-deltaic sequence; the vertical succession within the upper interval records a transgressive marginal marine to fluvial sequence.

With regression of the shoreline during late Coniacian, a deltaic plain was established, which prograded over nearshore marine deposits of the Rough Rock Sandstone. Marshes and backswamps were restricted to relatively narrow belts between poorly developed levees and interdistributary bays. Marsh/backswamp peats were prevented from attaining significant thicknesses by frequent overbank flooding and influence of brackish water.

With continued regression, the deltaic plain developed into an alluvial plain. Backswamps were defined by the configuration of the interfluvial areas and were more laterally continuous than those of the deltaic plain. Backswamps were protected from overbank flooding by well-drained levees and persisted for prolonged periods of time, producing thick peat deposits. Near the end of this period of regression, fluvial activity diminished, which resulted in the progressive expansion of the interfluvial areas and in production of thicker and more laterally continuous peat deposits.

With the onset of transgression, extensive interfluvial lakes produced by rising base level gradually drowned the backswamps. With continued transgression and decrease in fluvial activity, the alluvial plain developed into a marginal marine coastal plain. The influence of brackish water on coastal-plain marshes resulted in the preservation of thin peat deposits. The marginal marine coastal plain ultimately yielded to a period of erosional transgression prior to deposition of the nearshore marine deposits of the Yale Point Sandstone.

Figure 20. Upper part of the Cretaceous section at the northeastern escarpment of Black Mesa, 4 km north-northwest of Stop 17. Ky—Yale Point Sandstone; Kw—upper carbonaceous member of the Wepo Formation; Kr—Rough Rock Sandstone. Section is approximately 300 m thick.
REFERENCES CITED


INTRODUCTION

This field-trip guide describes the common sedimentary structures that occur in eolian sands. The outcrops that are described occur in the Navajo and Entrada Sandstones between the areas of Page, Arizona and St. George, Utah (Figure 1), but the sedimentary structures of these two sandstones are typical of most eolian deposits. The main part of the guide discusses the geologic setting and the origin of the various structures, and the road log discusses which structures are best displayed at selected outcrops.

Geologic Setting

The late Paleozoic and early Mesozoic were the peak times of eolian deposition in the western interior of the United States. On this field trip we will be looking at the Upper Triassic-Jurassic Navajo Sandstone and the Middle Jurassic Entrada Sandstone, the principal lower Mesozoic eolian deposits of the southwestern Colorado Plateau.

Kocurek and Dott (1983) have summarized the Jurassic paleogeography and paleoclimates of the western interior. During this time the southern Colorado Plateau lay at a paleolatitude of 10°-20° north, and paleomonth was a little west of present north. Eustatic sea levels were relatively low through the late Paleozoic and early Mesozoic, and marine waters were much less extensive than during the early to middle Paleozoic and the Cretaceous. The climate was generally warm and dry. Winds, as interpreted from crossbedding, were from northerly directions ranging from the present northwest to present northeast (Poole, 1962; Kocurek and Dott, 1983).

The entire upper Paleozoic to lower Mesozoic section thickens and contains a greater proportion of marine beds toward the west. The axis of the depositional basin ran through the present Basin and Range province, an area from which Triassic and Jurassic deposits were later largely removed (Peterson, 1972; Kocurek and Dott, 1983). West of the basin axis were tectonic highlands of which little is known. The sections thins and contains a greater proportion of nonmarine beds eastward toward the Transcontinental Arch and toward local uplands (the remnant Ancestral Rockies in Colorado and the Mogollon Highlands in southern Arizona). Sources of fluvial sediment were largely to the south and east of the depositional areas (Poole, 1961). Sources for the huge volumes of upper Paleozoic and Mesozoic eolian sand are poorly known, but a northern cratonic source on the Canadian Shield was probably important (Kocurek and Dott, 1983).

Stratigraphy

The Navajo Sandstone is the uppermost formation of the Glen Canyon Group, which also contains, in descending order, the Kayenta Formation and the Moenave Formation or its lateral equivalent, the Wingate Sandstone (Figure 2). The entire group may be Jurassic in age, although a Triassic age cannot be ruled out for the lower part (Peterson and Pipirinos, 1979). At the base of the group is a regionally traceable unconformity, designated the J-0 unconformity by Pipirinos and O’Sullivan (1978), that locally separates the Moenave Formation or Wingate Sandstone from the underlying Chinle Formation, of Triassic age.

The Moenave and Kayenta Formations are both composed largely of fluvial redbeds. The Moenave Formation contains the ledge-forming Springdale Sandstone Member at its top. Reddish-brown mudstone of floodplain origin is the dominant rock type in the Kayenta Formation, although fluvial sandstone and siltstone are also present. Thin intervals of eolian sandstone occur in the upper part of the Kayenta, and the contact with the overlying Navajo Sandstone is gradational (Middleton and Blakey, 1983). The fluvial sandstones of the Kayenta are less well sorted and mineralogically less mature than the eolian sandstones of the upper Kayenta and Navajo, and the fluvial crossbeds dip westward in contrast to the southeastward crossbeds in the eolian sandstones in the upper Navajo.

The Carmel Formation consists of sandstone, siltstone, mudstone, and evaporites. The Carmel Formation is overlain by the Entrada Sandstone, a widespread unit of dominantly tidal flat and sabkha on the east and south (Blakey and others, 1983). Farther to the southeast, in the Page area, the lower part of the Carmel intertongues with the eolian Page Sandstone (Peterson and Pipirinos, 1979). The crossbedding in the Page Sandstone dips to the south-southwest (Blakey and others, 1983).

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few miles to the east (Thompson and Stokes, 1970). To the west, between Page and Kanab, the Entrada pinches out and the Dakota rests on the Carmel Formation (Figure 2).

SEDIMENTARY STRUCTURES

Sedimentary structures in eolian sands fall into several categories: small-scale primary structures (fine structures) that record the depositional processes that operated at the sediment surface, larger-scale cross-stratification that records dune morphology and behavior, and miscellaneous structures such as trace fossils, deformational structures, or evaporite-related structures that record other postdepositional or syndepositional physical or chemical processes. The following discussion considers the origin and significance of these structures.

Fine Structures

Fine structures, as defined by Hunter (1985), are the smallest sedimentary structures, particularly the thinnest strata, that exist in a rock. No upper limit can be placed on the scale of these structures, but most strata that qualify as fine structures are less than a few centimeters thick. In general, strata that qualify as fine structures are without finer lamination internally. The classification of fine structures used here is genetic and, like all genetic classifications, cannot always be applied with confidence.

The three principal types of fine structure that are formed in dry sand are climbing-wind-ripple structures, rainfall lamination, and sandflow cross-stratification (Hunter, 1977; Fryberger and Schenk, 1981; Kocurek and Dott, 1981). Several types of fine structure form in moist to submerged interdune deposits, but only a few types have been studied sufficiently to be distinguished with confidence (Hunter, 1981).

Climbing-wind-ripple structures

Deposits formed by the climbing of wind ripples are common in most outcrops of the Navajo and Entrada Sandstones of the southern Colorado Plateau. This commonness is fortunate for the geologist concerned with the interpretation of depositional environments because the distinctive internal structure of such deposits provides the best single piece of evidence for an eolian origin of these formations. Excellent examples of climbing-wind-ripple structures can be seen in the Navajo Sandstone at the bottom of Water Holes Canyon (Stop 1) and in the Entrada Sandstone northwest of Page (Stops 2 and 3). To see the structures to best advantage, look for places where the outcrop surface bevels the bedding at a low angle.

Climbing-wind-ripple structures can form on any wind-rippled surface that is receiving net deposition. In the Navajo and Entrada Sandstones, climbing-wind-ripple structures occur most commonly in bottomset or toeset (basal-apron) deposits that interfinger upward with foreset (including slipface) deposits. In horizontal exposures of trough crossbedding, climbing-wind-ripple structures are best developed at the lateral margins of the troughs. Climbing-wind-ripple structures can make up complete sets of crossbeds; provided the slope angles were everywhere less than the angle of repose. In a few thin sets of crossbeds that represent nearly critically climbing small superimposed dunes (especially in the Entrada Sandstone at Stops 2 and 3), climbing-wind-ripple structures make up the preserved topset deposits.

Any climbing-ripple structure can be thought of as consisting of two components: translatent strata and rippleform strata, the latter including ripple-foreset cross-laminae (Hunter, 1977). In climbing-wind-ripple structures, however, the rippleform strata are generally not visible. A translatent stratum, or what Fryberger and Schenk (1981) call a ripple-produced stratum, is the deposit left by a single migrating ripple during its lifetime. Where the ripples climbed at an angle so low that the stoss sides of the ripples were eroded (in which case the angle of climb is called subcritical or stoss-erosional), the translatent strata have sharp, erosional contacts. The translatent stratification formed by stoss-erosionally climbing wind ripples long went unrecognized as having anything to do with ripples, for it rarely has any waviness suggestive of a rippled bed or any ripple-foreset cross-lamination suggestive of ripple migration (Figure 3). However, this kind of stratification can often be recognized (or at least suspected) at a glance by its "pin-stripe" appearance, a small-scale cyclicity that is suggestive of varves (Figure 3). On closer inspection, the translatent strata can often be seen to be inversely graded, another good indicator of wind-ripple origin.

Translatent strata differ from other strata in that the imaginary time lines within them are not parallel to the bounding surfaces but rather intersect them; almost any set of crossbeds is a translatent stratum. Furthermore, when the translatent strata are of climbing type, the time lines cross from one layer into adjacent ones, meaning that
all the strata in a set were forming at the same time. Of course, the law of superposition still holds along any vertical column.

The rarity of ripple-foreset cross-laminae within the translatent strata formed by climbing wind ripples derives from the nature of wind ripples. In wind ripples, in contrast to current ripples in water, the lee slopes dip so gently that avalanching does not occur, and consequently there are no miniature sandflow cross-strata defining the ripple foresets. Moreover, the thin lamination produced by wind gusts is naturally faint and tends to be blurred by the impacts of saltating grains. Although the rarity of ripple-foreset cross-laminae makes the translatent nature of the strata formed by climbing wind ripples difficult to recognize, this rarity helps in distinguishing climbing-ripple structures formed by wind ripples from those formed by current ripples.

Grainfall Lamination

Grainfall lamination is formed when previously saltating grains pass into the relatively calm air in a zone of flow separation leeward of a dune crest and settle out onto the lee slope or basal apron in front of the dune. As they fall, the grains lose the forward momentum that was imparted to them by the wind and strike the surface so gently that the thin lamination produced by grain segregation during wind gusts is not destroyed. Grainfall lamination is distinguished from translatent stratification formed by climbing wind ripples by the

extreme evenness of grainfall lamination and by its typical faintness and lack of small-scale cyclicity (pin-stripping). Distinguishing grainfall lamination from plane-bed lamination, which is formed under conditions of intense saltation in strong winds, is probably impossible in hand specimens. Winds strong enough to produce plane beds are not common in modern dunes, however, especially on the moderately steep lee slopes where grainfall deposition is most common.

Grainfall lamination is uncommon in the Navajo and Entrada Sandstones of the field-trip area. Probably this is because of the generally large size of the dunes that formed these sandstones. On large dunes grainfall deposition may never reach the base of the slipface, and the grainfall deposits of the upper slipface tend to be destroyed by avalanching. Even if they escape destruction by avalanching, grainfall deposits of the upper slipface are usually destroyed in the course of dune migration. Preserved grainfall deposits in the field-trip area are generally found in thin sets of cross-strata formed on small dunes; foreset cross-strata of grainfall type are associated with topset deposits formed by climbing wind ripples in some thin sets of cross-strata in the Entrada Sandstone at Stops 2 and 3. A 1-m-thick set of cross-strata of grainfall type can be seen at Stop 5 (Figure 4). Grainfall or, less probably, plane-bed deposits also occur in some large-scale sets of concordant cyclic cross-strata in the Navajo Sandstone, as at Stop 7 (Hunter and Rubin, 1983).

Sandflow cross-strata

The large-scale, steeply dipping cross-strata that are generally thought of as classically eolian were formed by sandflows (non-coherent avalanches) on the slipfaces of tall dunes. In dip cross sections, sandflow cross-strata are most easily identified by their lack of curvature and by their distinctive toes or basal pinches (Figure 5), and in strike cross sections or horizontal exposures they are most easily identified by their thickness (typically a few centimeters) and lenticularity.

Two measurable characteristics of sandflow cross-strata are useful in quantitatively interpreting eolian deposits (Hunter, 1981). Because the original dip angles of sandflow cross-strata have a narrow range (the average dip angle being 32-33°), the present dip angle is a good indicator of the amount of compaction. A reduction in dip angle to 27°, a common amount of reduction in the upper Paleozoic and lower Mesozoic sandstones of the Colorado Plateau, indicates 19 percent compaction. And because the thickness of sandflow layers is positively correlated with slipface height (although the relationship has

Figure 3. Climbing-wind-ripple structure (showing typical “pin-striped” appearance) in the Entrada Sandstone northwest of Page, Arizona (near Stop 3), quarter-dollar for scale. A few ripple-foreset cross-laminae are visible within some of the cross-erodingly climbing translunat strata; they indicate that the ripples migrated toward the right. The 5-cm-thick layer with indistinct cross-laminae (near the bottom of the photograph) is a translunat stratum that probably was produced by a small dune. The faint lamination within this stratum is of grainfall type except for some climbing-wind-ripple translatent strata just above the base of the dune-formed translunat stratum.

Figure 2. Columnar sections of Glen Canyon and San Rafael Groups in southwestern Utah (Zion-Kanab area) and northwestern Arizona (Page area).
much scatter), the sandflow cross-strata can give some indication of dune size; this relationship has recently been quantified by Kocurek (unpublished data).

Sandflow cross-strata are common in both the Navajo and Entrada Sandstones in the field-trip area. However, they are not common in all sets of cross-strata, and their rarity in some sets does not indicate a subaqueous origin for those sets.

**Wet-interdune fine structures**

Interdune areas can be dry, moist, or ponded. Where the sand is dry, the same fine structures that form on gently sloping dune surfaces can form. Where the sand is moist, the deposition of wind-blown sand involves the adhesion of grains to the surface, and the resulting fine structures are called adhesion structures (Kocurek and Felder, 1982). Where interdune ponds or streams occur, the various types of fine structure formed by oscillatory flow, unidirectional flow, and settling out in quiet water can form. Besides having distinctive structures, some interdune deposits differ texturally from associated dune deposits. The interdune deposits, although mostly sand, can have both coarser material (coarse sand and very fine pebbles) and finer material (silt and mud) than dune deposits.

The most common wet-interdune fine structure (or complex of fine structures) in the upper Paleozoic and Mesozoic eolian sandstones of the western interior is a type of stratification characterized by a small-scale irregular lenticularity or crinkly appearance (Figure 6). This lenticularity must reflect some combination of deposition on irregularly wavy surfaces and small-scale deformation soon after deposition. Many physical, chemical, and biological processes may have been involved (Hunter, 1981), but recent work in Arabian dune fields (Fryberger and others, 1983) suggests that the precipitation and dissolution of evaporite minerals within the sediment were the processes most important in giving the stratification its distinctive appearance. The best examples of this type of fine structure in the field-trip area are in the transition zone between the Navajo Sandstone and the underlying Kayenta Formation, for example, at Snow Canyon (Stop 10). Besides occurring in interdune deposits, the structure is common throughout the arid floodplain or sabkha deposits of the Kayenta Formation.

Easily identified adhesion structures are surprisingly rare in the upper Paleozoic and Mesozoic interdune deposits of the western interior, probably because of processes involving evaporite minerals, which can either prevent adhesion structures from assuming their typical form or alter the structures after they form. The only undoubted adhesion structures that we know of in the field-trip area are some thin layers characterized by quasiplanar adhesion lamination and climbing adhesion ripples in dune deposits (not interdune deposits) of the Entrada Sandstone at Stop 3. The formation of adhesion structures at levels above the interdune troughs requires rainfall, not just the intersection of the water table by the troughs.

**Crossbedding**

**Simple and compound crossbedding**

Where fine structures are deposited on an inclined surface such as the flank of an eolian dune, the resulting deposit is an inclined bed called a crossbed. As a dune migrates, successive crossbeds are deposited on its advancing surface, thereby producing a set of crossbeds. Unless net deposition occurs while the dune migrates, the set of crossbeds will be thoroughly reworked by the next dune having an equally deep trough that migrates across the depositional site. In contrast, in an area undergoing net deposition, dunes will move upward (climb) relative to the generalized depositional surface (Allen, 1963; Brookfield, 1977; Rubin and Hunter, 1982). The result of this process of dune climbing (migration accompanied by net deposition) is a coset of crossbeds, shown in simplest form in row 1 of Figure 7.
Two-dimensional bedforms are straight and parallel in plan form; the flanks of the bedforms have the same strike at all locations. Two-dimensional bedforms produce two-dimensional cross-bedding: cross-bedding in which all foresets and bounding surfaces have the same strike. In plans showing the direction and inclination of dips of cross-beds and bounding surfaces, dips of all planes plot along a single straight line through the center of the plane.

Invariable
Cross-bedding deposited by invariable three-dimensional bedforms has bounding surfaces that are trough-shaped; bounding-surface dips in a single trough (or in identical troughs) plot as a nearly straight line.

Variable
Bounding surfaces have complex shapes produced by such processes as zig-zagging of scour pits; dips of bounding surfaces plot as scatter diagrams.

Perfectly transverse
Plots of cross-bed and bounding-surface dips have bilateral symmetry; dip directions are distributed unimodally.

Perfectly longitudinal
Plots of cross-bed and bounding-surface dips have bilateral symmetry; dip directions may be distributed bimodally (as shown) or may be unimodal as a result of migration of the nose of the main bedform. Perfect longitudinality is evidenced by vertical accretion of bedforms; cross-beds dip in opposing directions on opposite flanks.

Transverse, oblique, and longitudinal cross-bedding are not distinguishable unless bedforms are at least slightly three-dimensional (see below).
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Symbols:
- Main Bedforms
- Superimposed Bedforms
- Planform Enclaves
- Scour Pits
- Re-deposition Transport (Direction Only)
One of the most important goals of eolian sedimentology is to determine wind regimes and sand transport directions from crossbedded geometry. The first step of this interpretive process is to reconstruct dune geometry and behavior from crossbedded geometry. In general, dunes with simple geometry and simple behavior deposit simple crossbedding, whereas dunes with complicated geometry or complicated behavioral characteristics produce more complicated crossbedding (Figure 7). Morphologic complications are caused by such features as sinuous, discontinuous, or irregular dune crests; scour pits in dune troughs; superimposed bedforms, peaks, spurred dunes, or other topographic features; and morphologic differences between individual dunes in a dune field. Behavioral complications are produced by such processes as reversals in dune asymmetry or migration direction (in response to random fluctuations in dune geometry, thereby demonstrating that the dunes had a relatively simple two-dimensional morphology. Additional examples of crossbedding produced by fluctuating flow are given by Hunter and Rubin (1983). The cyclicity in the Navajo Sandstone is interpreted to be annual (Stokes, 1964; Hunter and Rubin, 1983) primarily because unusually strong winds would have been necessary to have caused the observed dune migration in shorter time cycles.

Morphologic complications caused by topographic features

Like crossbedding produced by fluctuating flow, the crossbedding produced by dunes with superimposed dunes or other morphologic complications can be either random (produced by random topographic features) or cyclic (produced by trains of similar superimposed dunes, such as sinuous, or superimposed dunes or other morphologic features). Migration of the superimposed features is often accompanied by erosion on the stoss side followed by deposition on the lee side. The cyclic passage of such topographic features deposits cyclic sets of crossbeds that are grouped in a coset deposited by the main dune.

Two kinds of cyclic, complicated-morphology, compound crossbedding are conspicuous throughout the Navajo and Entrada Sandstones along the route of this field trip: (1) structures formed by small dunes migrating across the lee slopes of larger dunes (Figure 10), and (2) structures formed by dunes with scour pits that migrated along the lee slopes of larger dunes (Figure 12). These kinds of structures listed above are geometrically related because intersecting troughs of the main dunes and the superimposed dunes form topographic depressions that behave geometrically like scour pits, and migration of the superimposed dunes can cause these depressions to migrate along the trough of the main dune. Regardless of the morphologic details, alongcrest migration of superimposed features causes a cyclic rotation in dip direction of crossbeds within the set (Figure 12).

Recognition of structures deposited by superimposed dunes is important, not only because such structures are useful for reconstructing in detail the original dune morphology, but also because the migration direction of the superimposed features relative to the main dune is an important indicator of whether the main dunes were transverse, oblique, or longitudinal. Superimposed features can be expected to move dominantly in one alongcrest direction if they are superimposed on an oblique or longitudinal dune. In contrast, on a transverse dune, superimposed features can be expected to show a more consistent preferred alongcrest migration (Rubin and Hunter, 1985). In the Entrada Sandstone in the vicinity of Page, Arizona, the compound crossbedding indicates that the superimposed bedforms had a pronounced tendency to migrate in an alongcrest direction (to the right of the migration direction of the main dunes). This characterizes the direction of the superimposed features relative to the main dune. In contrast, transverse beds are supported by observations to the east of the field-trip area (in northwestern New Mexico), where sand-body surface relief defines lineations (burred dunes) that trend northeast-southwest (Vincenlote and Chittum, 1981), approximately parallel to the local sand transport direction inferred by Poole (1962) and Tanner (1965).

Combined complications

As might be expected, dunes having complicated morphology and complicated behavior can produce extremely complicated crossbedding. Examples of one of the more cyclic structures deposited by such dunes are the "zigzags" produced by lee-side spurs and scour pits that reverse their asymmetry and direction of alongcrest migration (Figure 13). This kind of behavior can be expected of dunes that are transverse to the long-term resultant transport direction and are subject to winds that have a reversing crest-parallel component of flow. Such reversals could be expected to cause opposing flanks of lee-side spurs to experience erosion and deposition alternately.

Miscellaneous Structures

Deformational structures

Several types of structures formed by deformation of unconsolidated dune sands are found in the Navajo and Entrada Sandstones of

Rubin and Hunter
Figure 8. Conformable cyclic foresets in a 10-m-thick set of crossbeds in the Navajo Sandstone, Zion National Park, Utah (Stop 7). The cyclicity is due to an alternation of grainfall deposition and deposition by wind ripples and is interpreted to be the result of an annual cycle of fluctuations in wind speed or direction. In this particular set of crossbeds, the grainfall deposits are more resistant to erosion than the climbing-wind-ripple structures. The mean distance of dune advance represented by individual cycles is 0.3 m.

Figure 9. Cyclic compound crossbedding (discordant cyclic foresets) in a set of crossbeds in the Lamb Point Tongue of the Navajo Sandstone, north of Kanab, Utah (Stop 5). The crossbeds are largely of sandflow type. After each period of dune advance, the slipface was eroded back. During each period of slipface erosion, a wedge-shaped body of sand (light-colored) was deposited at the base of the slipface. The mean distance of dune advance represented by individual cycles is 1.5 m. The cyclicity, which must have involved fluctuations in wind direction, is interpreted to have been annual.

The field-trip area. The structures range in scale (fold amplitude or thickness of deformed zone) from 1 cm or less to 10 m or more.

The smallest deformational structures include some of the irregularities in the stratification of wet-interdune deposits (described under "Fine Structures") and structures formed by the sliding of coherent or semicoherent sediment masses down slipfaces. The latter, whose formation has been studied experimentally by McKee and others (1971), are not common; most of the avalanches were sandflows by the time they came to a stop, even if they started as slides.

Decimeter-scale contortions are common in flat-bedded intervals in the transition zone between the Navajo Sandstone and the underlying Kayenta Formation, for example, at Snow Canyon (Figure 6; Stop 10). We interpret these structures to have formed by a combination of loading and fluid escape. Some of the flat beds are certainly waterlaid, and even the eolian deposits probably became water-saturated after slight burial. The uneven loading due to the migration of sand dunes over the flats (a "rolling-pin" mechanism) then induced compaction and fluid escape.

Zones of contorted bedding as thick as 10 m or more are common in parts of the Navajo Sandstone, for example, in the lower part of the formation and in the Lamb Point Tongue north of Kanab (Stops 5 and 5A). Another example is well exposed in the walls of Water Holes Canyon (Figure 14; Stop 1). Structures of this type have been studied by Sanderson (1974), Doe and Dott (1980), and Horowitz (1982). All these workers agree that the contortions formed while the sand was saturated with water but that the deposits are eolian, not subaqueous. An origin below the level of the dune troughs is thereby indicated.

One characteristic of the deformation poses a paradox if the deformation took place below the level of the dune troughs. The overturning of crossbeds at the bases of deformed zones indicates shearing by an overriding sediment mass that moved approximately in the direction of crossbed dip (Figure 14). How could dune migration direction have controlled the direction of shearing below the level of the dune troughs? One proposed interpretation capable of resolving this paradox is that of Horowitz (1982), which involves earthquake-induced liquefaction centered beneath an interdune trough and the consequent collapse of the dune impinging on the trough. A second possible explanation is that the overriding dunes, which may have been as much as several hundred meters in height (Rubin and Hunter, 1982), acted like huge rolling pins that deformed the sediment in the underlying beds. More study of these structures is needed before their origin is fully understood.
Figure 12. Computer-graphics model of the origin of scalloped crossbedding formed by superimposed dunes.
A. Dune morphology and vertical sections. The main dune is migrating left to right, and the superimposed dunes are migrating away from the viewer. Intersections of the troughs of the two sets of dunes produce scour pits, and the cyclic passage of these scour pits through the left vertical section produces the scallops.
B. Horizontal and vertical sections. Deposition causes the scour pits to migrate upward through the horizontal section, producing the "fingertip" structures.

Rubin and Hunter
Trace fossils

Although trace fossils are far from common in eolian sandstones, except perhaps in wet-interdune deposits, they are known (Ahlbrandt and others, 1978) and indeed are fairly common in a few beds of the Navajo Sandstone in the field-trip area. We have no doubt that the beds in which the trace fossils occur are eolian. At least three kinds of trace fossils have been seen in dune-foreset or basal-apron deposits at Water Holes Canyon (Stop 1).

Small (diameter 0.5-1.0 cm), densely clustered, diversely oriented, unlined tubes filled by structureless sand occur near the base of a set of crossbeds at the bottom of the canyon. These trace fossils are probably intrastratal trails produced by an organism that lived on the lower lee slopes of dunes. Penetrating into the same set from its upper bounding surface are roughly vertical tubes about 10 cm in diameter filled by sand, some of which is stratified. These tubes evidently stood as open burrows and probably required damp sand to prevent collapse. The set of crossbeds in which these trace fossils occur, unlike so many other sets at the locality, has no carbonate nodules. The lack of these nodules, which are interpreted to have formed by replacement of evaporites, suggests that the moisture was provided by unusually fresh water, which was evidently attractive to organisms.

Evaporite structures

Centimeter-diameter carbonate-cemented nodules are abundant at many outcrops of the Navajo Sandstone and at some outcrops of the Entrada Sandstone. The nodules have several characteristics that suggest that some of them formed by replacement of evaporite crystals or nodules. In some sets of crossbeds that contain the nodules, the nodules are concentrated near the bottom or top of the set (Figure 15). Both of these sites are inferred to be favorable for evaporite precipitation, because both tend to be near the interface between air and interstitial waters at the time of deposition. Water can rise into lower foresets and evaporate, thereby depositing evaporites that are concentrated near the base of a set of crossbeds; and water that evaporates from a dune trough may deposit evaporites in the top of the underlying set. In the Entrada Sandstone east of the field-trip area (near Gallup, New Mexico) some nodules that are concentrated at the base of a set have preserved the shape of the evaporite crystals that were replaced (Figure 15), demonstrating that at least these nodules formed by replacement of evaporites. The origin of the more typical spherical or irregular nodules that appear to be randomly distributed within sets of crossbeds may also be related to evaporites, but such an origin is not conclusively demonstrated.
ROAD LOG AND STOPS

DAY 1

Mileage

0.0 Road log begins at intersection of U.S. 89 and U.S. Alternate 89, 24 miles south of Page, Arizona. Take U.S. 89 north towards Page. Ascend Echo Cliffs, passing through section from Moenkopi Formation at base of cliff to Navajo Sandstone at top.

12.7 Begin descending hill, passing outcrops of Page Sandstone and upper part of Navajo Sandstone for next 1 mile.

17.8 STOP 1. Navajo Sandstone at Water Holes Canyon. This stop is on property of the Navajo Nation. Persons following this route should obtain permission to stop here from the Navajo Nation Minerals Department, Window Rock, Arizona. It is worth this extra effort to stop at this site, because the structures are exceptionally well displayed. Allow several hours to follow the hike and to examine the structures described. Park on north side of canyon; enter gate on east side of highway; walk east along canyon rim until reaching a location where north wall of canyon becomes less steep; climb down to canyon floor. Continue east along the canyon floor, noting the following features: climbing wind-ripple structures (including some with preserved foreset cross-laminae); sandflows (visible in vertical sections in the canyon walls and in a few horizontal sections on the floor); two kinds of trace fossils; nodules — inferred to be replaced evaporite nodules — that in some beds are concentrated above and below bounding surfaces; small subsets of crossbeds deposited by superimposed dunes; scalloped crossbedding formed by fluctuating flow; and a zone of deformed bedding that includes almost the entire section in the canyon wall. At the site of the deformed zone, climb the north wall of the canyon and continue east along the rim to inspect the conformable fluctuating-flow bedding cycles. The following features are visible along the rim on the return route to the highway: a thick bed containing burrows that weather out of the rock (just below the elevation of the rim) and horizontal sections through trough-shaped sets of crossbeds — some of the sets are cyclically filled, first from one direction, then from another.

Continue north along U.S. 89 towards Page.

22.9 South turnoff to Page (Loop 89) on right. If time permits, continue north along U.S. 89. Page Mesa on right (east) is held up by Page Sandstone.

24.2 North turnoff to Page (Loop 89) on right.

24.7 Overlook on south side of Glen Canyon (gorge of Colorado River). Turn right on gravel road.

25.2 ALTERNATE STOP 1A. Contact of Page Sandstone and Navajo Sandstone. Park along road and walk 0.3 mile northeast to small mesa. Contact is near base of mesa, marked by large polygonal cracks at top of Navajo. Return to U.S. 89.

25.7 Turn right (north) onto U.S. 89, cross bridge over Glen Canyon.

26.2 ALTERNATE STOP 1B. Visitor center at Glen Canyon Dam. Navajo Sandstone exposed in walls of canyon. Note zone of large-scale contorted bedding. Return to Page for night.

DAY 2

0.0 Retrace route from Page to visitor center at Glen Canyon Dam.

0.5 Leave Glen Canyon Dam driving on Lakeshore Drive toward Wahweap Marina. Follow road north along west side of Lake Powell. Navajo Sandstone and Page Sandstone crop out along road for first 2 miles. Entrada Sandstone forms white cliffs on north side of Lake Powell.

4.3 Wahweap Marina.

6.8 Intersection of U.S. 89 and Lakeshore Drive. Cross U.S. 89, entering parking lot of Lake Powell Motel.

6.9 STOP 2. Entrada Sandstone on hillside in back of Lake Powell Motel. Park in the motel lot, and walk up the hill toward the closest outcrops. Note the climbing wind-ripple structures. Continue walking up the hill, and note that the crossbeds generally dip toward the right (toward the west or southwest), whereas the bounding surfaces that separate the crossbeds dip into the outcrop (roughly toward the south). This set of beds was deposited by a large dune that migrated toward the south or southwest while superimposed bedforms migrated toward the west or northwest.

7.0 Turn left (west) onto U.S. 89 toward Kanab, Utah.

12.3 Turn left (south) onto dirt road. Close gate behind you. Continue to pulloff.

12.8 STOP 3. Entrada Sandstone in mesa walls. Note the excellent examples of fine structures, including climbing adhesion ripples, and note the scalloped crossbedding (on east side of dirt road) that was produced by scour pits that migrated toward the west, while the main dune migrated toward the south. Return to U.S. 89.

13.3 Turn left (west) onto U.S. 89 toward Kanab. Several outcrops of Entrada Sandstone along highway for next 7 miles.

29.7 Entrada Sandstone in cut on left (south) side of highway. North of highway for next 2 miles are outcrops of white Entrada Sandstone and underlying red beds of Carmel Formation.

35.9 For next 0.7 mile, pass downstream through steeply tilted Navajo Sandstone. Continuing on to Kanab, note Kayenta Formation and lower part of Navajo Sandstone in Vermilion Cliffs on right side of highway.

75.3 Intersection of U.S. 89 and U.S. Alternate 89 in Kanab, Utah. Turn right (north), continuing on U.S. 89 toward Mount Carmel Junction.

78.5 Thin eolian sandstone in Kayenta Formation on right (east) side of highway.

79.8 Gradational contact between red beds of Kayenta Formation and overlying Lamb Point Tongue of Navajo Sandstone.

80.6 Turn right (east) onto gravel road into Kanab Canyon.

81.4 STOP 4. Zigzagging spur in Lamb Point Tongue of Navajo Sandstone. Park along road. This structure is shown in Figure 13. Measure dip directions of crossbeds and bounding surfaces within these zigzags, and note that the beds dip roughly toward the east and south (on the northeast and southwest flanks of the southeastward-plunging spur). The spur shifted back and forth laterally while the main dune migrated toward the southeast.

Rubin and Hunter
Continue along gravel road.

81.7 Turn around at driveway to private property; return to U.S. 89.

82.8 Turn right (north) onto U.S. 89 toward Mt. Carmel Junction.

83.2 STOP 5. Compound crossbedding in the Lamb Point Tongue of the Navajo Sandstone. Structure is shown in Figure 9. Park along road. Note the cyclic character of these foresets. They are interpreted to be annual layers, and they document the net rate of dune advance (1.5 m/year). Continue northward on U.S. 89.

84.2 For next 0.2 mile, pass upsection through Tenney Canyon Tongue of Kayenta Formation.

84.6 ALTERNATE STOP 5A. Zone of large-scale contorted bedding near base of main body of Navajo Sandstone. Park along highway. Exposures on both sides of road. The top of the deformed zone is at least locally a shear surface, not an erosional surface. Continue northward on U.S. 89.

85.1 Exposure of large-scale scalloped crossbedding in Navajo Sandstone in canyon wall on left (west) side of highway.

91.0 South turnoff on left (west) to Coral Pink Sand Dunes. The dunes are 12 miles southwest of the highway and worth the drive if the wind has been strong enough to obliterate the unsightly offroad-vehicle tracks that usually mar this otherwise beautiful area.

91.5 North turnoff to Coral Pink Sand Dunes.

92.0 Cross Sevier fault. Navajo Sandstone on east side, Carmel Formation on downthrown (west) side of fault.

95.0 Mt. Carmel Junction. Turn left (west) on Utah Highway 9. Note cut through oolitic limestone in Carmel Formation on right (north) side of road 0.2 mile past intersection.

102.5 Stratigraphically highest point on road between Mt. Carmel Junction and Zion National Park. Upper Cretaceous Tropic Shale and Dakota Formations overlie Carmel Formation unconformably.

106.0 For next 0.5 mile, pass eolian sandstone outcrops of Temple Cap Sandstone. Next outcrops after that are of Navajo Sandstone.

106.7 On right (north) side of road is a transverse exposure of a trough set in which foresets are cyclic (annual layers).

107.4 Entrance to Zion National Park.

107.6 Outcrop north of road has trace fossils and a trough set with zigzagging infilling (not visible from road).

108.3 ALTERNATE STOP 6A. Parking area providing view of Checkerboard Mesa (Navajo Sandstone). Walk along road toward mesa to view very thick (more than 30 m) set of crossbeds south of road and east of mesa. Return to vehicle and continue westward on Utah 9.

108.9 STOP 6. Parking area provides view eastward of transverse section of concave-downwind foresets in a tabular set in Navajo Sandstone. The variable strike of the foresets indicates that the dune was curved in plan form, whereas the planar lower set boundary indicates that the dune trough was relatively uniform in elevation (i.e., no scour pits in the trough). Continue driving westward.

109.6 STOP 7. Parking area for exposures of conformable cyclic foresets in the Navajo Sandstone (Figure 8). Horizontal exposures reached by walking 0.1 mile back (eastward) along highway, then going south from road. Vertical exposures are visible from a point 0.2 mile west of parking area, looking at opposite wall of canyon. Return to vehicle and continue westward.

110.1 Valley-bottom exposure of compound crossbedding deposited by downslope-climbing dunes (not visible from road).

111.5 ALTERNATE STOP 7A. Parking area provides view northward of coset of cross-strata exposed on hillside (Navajo Sandstone). Walk back (north and east) 0.4 mile to see large-scale scalloped crossbedding on southeast side of same hill.

111.7 ALTERNATE STOP 7B. Parking area provides view eastward of coset of cross-strata (Navajo Sandstone) with bounding surfaces that rise southward (in dip direction of cross-strata). A few meters above road level on north side of road is an outcrop of flat-bedded wet-interdune sandstone. Walk westward 0.1 mile along road to trailhead north of road. About 0.3 mile along trail is horizontal exposure of desiccation-cracked interdune deposits.

113.4 Enter 1.1-mile-long tunnel.

114.8 First of five hairpin turns. Next 2 miles provide several good views of Navajo Sandstone exposed in canyon walls.

115.8 Stratigraphic level of thin eolian sandstone in Kayenta Formation. The sandstone forms a resistant ledge visible on opposite side of canyon.

116.7 Exposure of subaqueous climbing ripples in Kayenta Formation.

117.4 Cross bridge over Pine Creek. Springdale Sandstone Member of Moenave Formation forms resistant ledge just above road level on both sides of bridge. The sandstone is fluvial.

117.9 Intersection of Utah 9 and park road. Turn right on park road into Zion Canyon.

120.8 Zion Lodge. Stop for night.

DAY 3

Retrace route to intersection of park road and Utah Highway 9.

0.0 Turn right (west) on Utah 9.

0.0 ALTERNATE STOP 7C. Zion National Park visitor center. Exhibits; books and maps for sale.

0.6 Booth at west entrance to Zion National Park. Chinle Formation crops out at road level between here and Springdale.

2.5 Pass through Springdale.

5.5 Shinarump Member of Chinle Formation forms resistant ledge at road level. Member is composed of fluvial sandstone and conglomerate. Beyond here road traverses Moenkopi Formation.

23.5 Begin descending hill, passing outcrops of Kaibab Limestone of Permian age.

24.5 Pass through La Verkin, continuing southward on Utah 9. Road follows Hurricane fault for 2 miles from here to Hurricane.
33.5 Outcrops of Moenkopi Formation near axis of Virgin anticline.

35.5 Intersection of Utah 9 and Interstate 15. Turn left (south) on Interstate 15.

42.5 Take exit from Interstate 15 to Utah 18, passing through St. George and continuing northward toward Snow Canyon State Park.

49.8 ALTERNATE STOP 8A. Turn right on road into Winchester Hills subdivision. Continue 1.2 miles and park along road. View transverse cross section of large-scale trough crossbedding in upper part of Navajo Sandstone. The trough-shaped sets were produced by scour pits that migrated out of the plane of the outcrop. Dune geometry is inferred to have been similar to that shown in row 3 of Figure 7. Return to Utah 18.

49.8 Turn right (or continue northward) on Utah 18.

51.4 STOP 8. Intersection of Utah 18 and road through Snow Canyon State Park. Note unusually large scale of crossbedding in upper Navajo Sandstone east of Utah 18, and note Quaternary basalt flows. Turn left onto road through park.

53.1 STOP 9. Large-scale scalloped crossbedding in Navajo Sandstone (visible toward west). Structure is shown in Figure 11. Continue south along road through park.

54.2 Relatively thin sets of crossbeds in lower part of Navajo Sandstone.

54.5 STOP 10. Interbedded flat-beded sandstone and crossbedded sandstone near base of Navajo Sandstone. These beds record the transition from subaqueous deposition in the Kayenta Formation to the eolian deposition in the Navajo Sandstone. Note several structures that are inferred to have been produced by the wetness of the sediment: deformatinal structures, crinkly lamination, and nodular appearance of bedding. Park along road and walk east to east wall of canyon. Return to vehicle and continue on road through park.

58.3 Turn left at town of Ivins.

59.8 Turn left (east) on old U.S. 91 toward St. George.

64.7 Rejoin Utah 18 at outskirts of St. George. End of road log.

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REFERENCES CITED


