• 9 •

Supai Group and Hermit Formation

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INTRODUCTION

The brilliant red cliffs and slopes present throughout the Grand Canyon comprise strata of continental, shoreline, and shallow-marine origin. Assigned to the Supai Group and Hermit Formation of Pennsylvanian and Early Permian age, these sedimentary rocks consist of a broad variety of lithologies—including sandstone, mudstone, limestone, conglomerate, and gypsum. Geologists have divided the rocks by their red-to-tan color and their weathering characteristics into five formations recognizable throughout the region. The five formations, in ascending order, are the Watahomigi Formation and Manakacha Formation (both Lower Pennsylvanian), Wescogame Formation (Upper Pennsylvanian), Esplanade Sandstone (Lower Permian), and Hermit Formation (Lower Permian). Geologists traditionally assign the first four formations to the Supai Group.

NOMENCLATURE AND DISTRIBUTION

Although the rocks now assigned to the Supai Group and Hermit Formation have a long and complicated nomenclature history (see McKee 1982a, page 3), three papers are responsible for the present terminology now in use in Grand Canyon. Darton (1910) proposed the name of Supai Formation for all the predominantly red strata between the Redwall Limestone and the Coconino Sandstone. Noble (1922) divided out the Hermit Shale from the top of Darton's Supai. McKee (1975) raised the Supai to group status and proposed four formations within the group: the Watahomigi, Manakacha, and Wescogame formations and the Esplanade Sandstone. The type section for each formation is near Supai in the Grand Canyon. Most publications within the last 35 years have used the term "Hermit Formation," in preference to shale, and this trend is followed here.

The Supai–Hermit terminology is sound throughout Marble Canyon and eastern and central Grand Canyon; however, several problems arise in western Grand Canyon. McNair (1951) assigned the lower part of the Supai interval to the Callville Limestone and recognized a Permian carbonate unit not present to the east, to which he assigned the term "Pakoon Limestone." McKee (1982a) did not recognize the Callville Limestone within the confines of Grand Canyon, and although he recognized the Pakoon Limestone, he did not include it within the Supai Group. A further complication exists in the western portions of the Grand Canyon where McNair (1951) assigned sandstone partly equivalent to the Esplanade and partly equivalent to the Hermit to the Queantoweap Sandstone.

The solution to this nomenclature problem is controversial and beyond the scope of this chapter, but the approach followed herein, pending further work, is to follow McKee's terminology in the Grand Canyon and Grand Wash Cliffs and to assign equivalent strata to the Callville and Pakoon limestones, Queantoweap Sandstone, and Hermit Formation (where applicable) in areas farther to the west, northwest, and north (Fig. 9.1).

To the southeast, along the southern margin of the Colorado Plateau, we can correlate the Supai–Hermit interval as far east as Sedona (Blakey 1979; Blakey, 1990; Blakey and Knepp 1988). Correlation to the southeast of Sedona is complicated by the addition of strata of Pennsylvanian age that is not directly time equivalent to the Supai. Termed the Naco Formation, these rocks were laid down in a basin separate from that in which the Supai was deposited. The fact that the Esplanade Sandstone is not present southeast of Sedona complicates the separation of the Supai and Hermit. Adding to the dilemma is the presence of a younger stratigraphic unit not present in the Grand Canyon, the Schnebly Hill formation (Blakey 1980, 1990).

To the east on the Defiance Plateau, Permian redbeds have been assigned to the Supai Formation (Read and Wanek 1961). Recent, Blakey (1990) demonstrated that only a portion of these redbeds are equivalent to part of the Supai in the Grand Canyon (see Fig. 9.2).

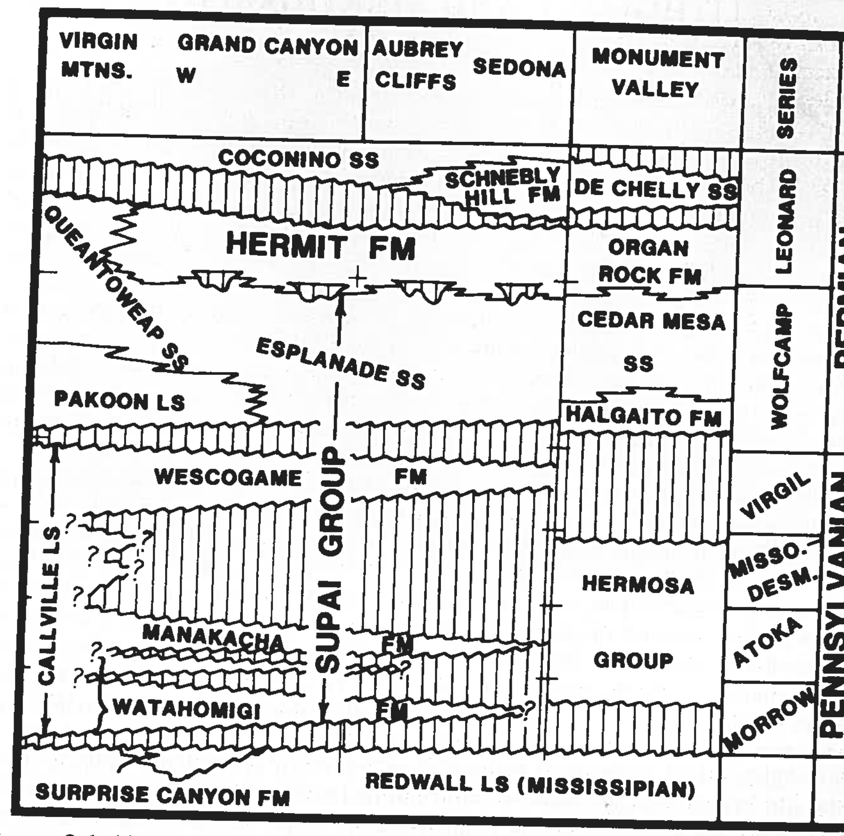

Figure 9.1. Time-rock chart of Pennsylvanian and Lower Permian rocks of northern Arizona. Vertical ruled lines show time represented by unconformities.

Figure 9.2. Index map of area of study.

LITHOLOGY AND STRATIGRAPHY

Watahomigi Formation

The Watahomigi Formation, the oldest formation in the Supai Group, consists chiefly of red mudstone and siltstone and gray limestone and dolomite (Fig. 9.3). The unit forms a broad, slightly westward-thickening sheet that ranges in thickness from 100 feet (30 m) in eastern Grand Canyon and at Sycamore Canyon on the east to 300 feet (90 m) in western Grand Canyon and along the Grand Wash Cliffs (Figs. 9.4 and 9.5). Carbonate content, which chiefly is very fine-grained (aphanitic) limestone to the east and granular (grainstone and packstone) limestone to the west, increases dramatically to the northwest. Grains include abraded fossil fragments, accretal grains, and pellets. Mudstone consists primarily of nondescript, slope-forming, poorly exposed units with thin, intercalated, bioturbated (disturbed by organisms), bright-orange, limey sandstone. Occasional bedding-plane exposures of the redbeds reveal various tracks, trails, and burrows. Most sections contain a basal chert–pebble conglomerate in which the clasts were derived from the underlying Redwall Limestone.

Throughout the Grand Canyon and the adjacent southern Colorado Plateau, the Watahomigi Formation can be divided informally into a lower redbed slope, middle carbonate ledge, and upper redbed slope (Blakey 1980; McKee 1982a). Based on included fossils and the presence of local conglomerate beneath the upper slope, McKee (1982a) assigned the lower slope and the middle ledge a Morrowan age and the upper slope an Atokan age (Fig. 9.2).

The lower contact of the Watahomigi Formation is everywhere sharp and unconformable with the underlying Redwall Limestone. Where the Surprise Canyon formation is present (see Chapter 8, this volume), geologists suspect an unconformity, but the contact is poorly exposed. The upper contact probably is conformable and was assigned by McKee to a zone of gray, jasper-bearing limestone and bright-orange sandstone and intercalated red mudstone. The contact apparently occurs at the change from primarily slope below to steep slope or cliff above and, therefore, may not be a consistent stratigraphic level across the region.

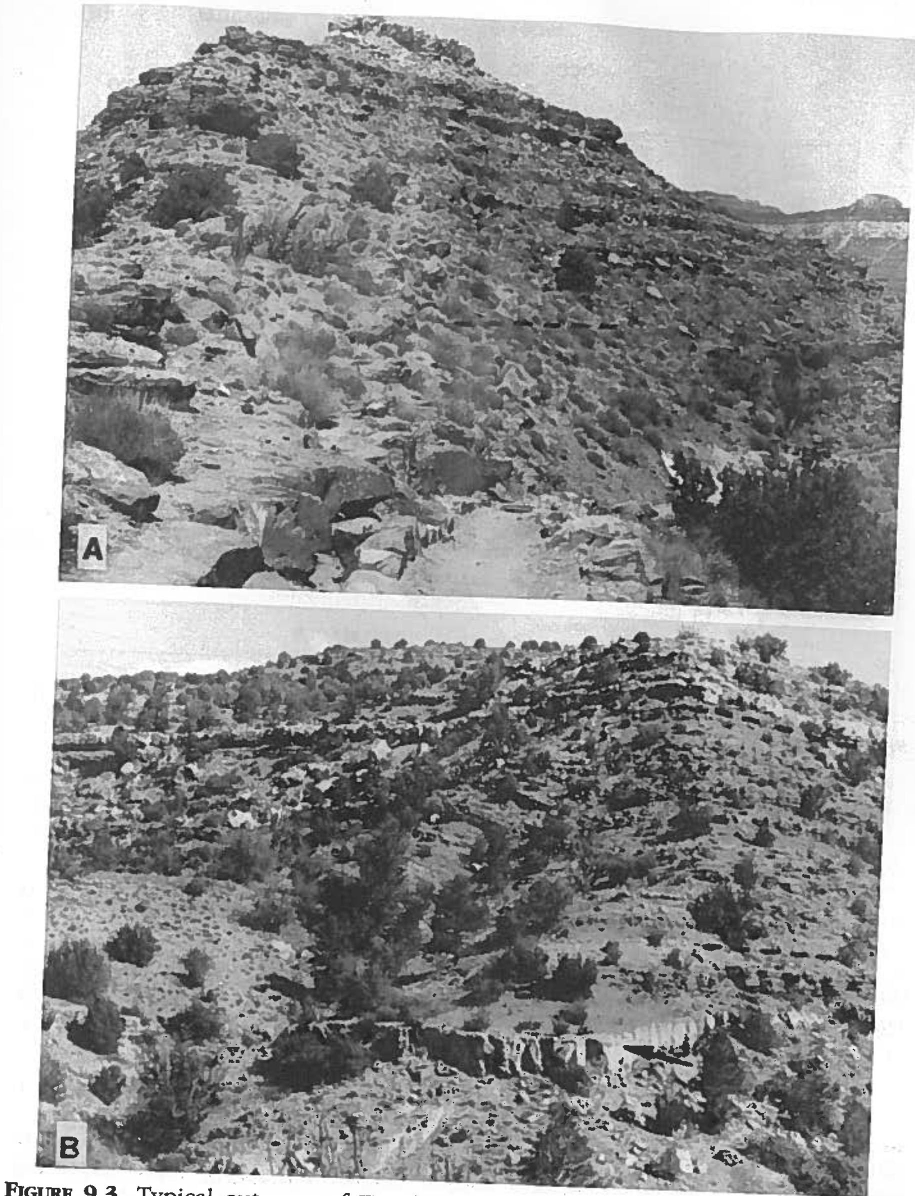

Figure 9.3. Typical outcrops of Watahomigi Formation. (a) Along Hermit Trail in Grand Canyon. (b) In western Mogollon Rim near Hell Canyon; arrow points to limestone marker bed in region. In both photos, dashed line marks top of formation.

The sharp increase in limestone west of a line paralleling the Hurricane Cliffs can be used to divide the Watahomigi Formation into an eastern redbed facies and a western carbonate facies (see McKee 1982a, Fig. 6). The westward increase in thickness, carbonate content, and marine fossils is typical of many Paleozoic rock units in the Grand Canyon region.

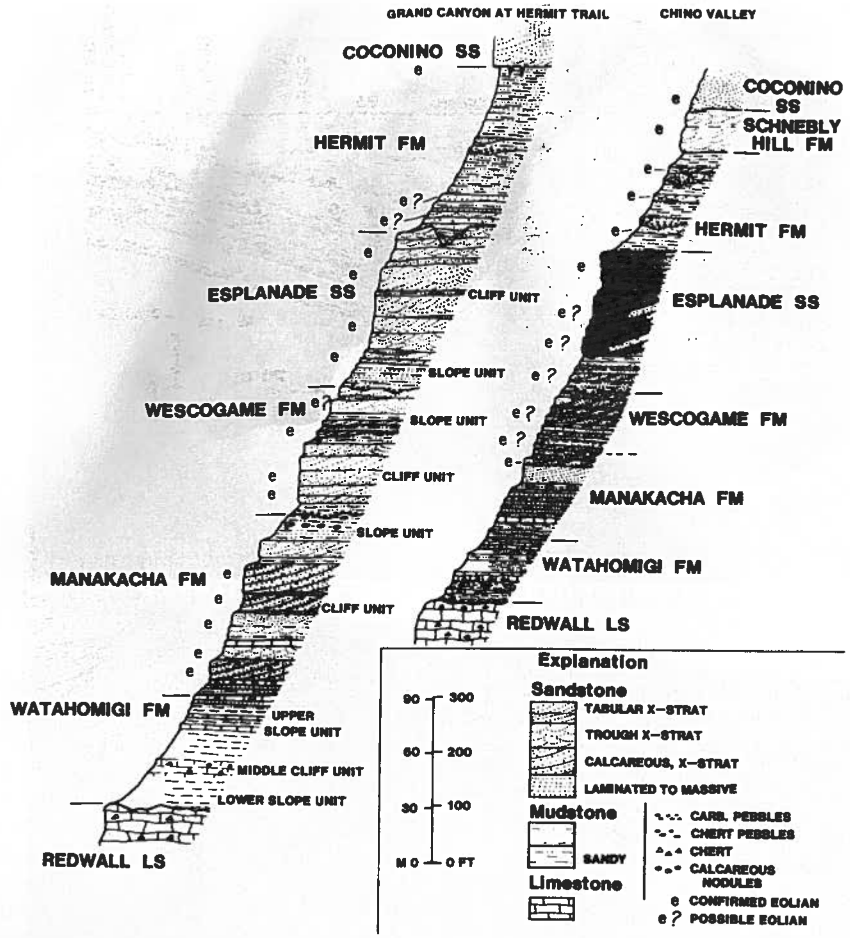

FIGURE 9.4. Columns of Supai Group and Hermit Formation and related strata showing distribution of known and suspected eolian strata.

Manakacha Formation

The Manakacha Formation marks an important change in the trend of Paleozoic depositional patterns in the Grand Canyon region. Following initial Cambrian sand deposition, the area was dominated by carbonates and minor mudstones from Middle Cambrian to Early Pennsylvanian time. The influx of quartz sand during the deposition of the Manakacha reflects a significant change across the western interior of the United States.

The Manakacha formation consists chiefly of quartz sandstone and intercalated red mudstone (Fig. 9.6). Unlike most other Paleozoic rock units, the formation is thicker in central Grand Canyon than it is in western Grand Canyon (McKee 1982a). The Manakacha forms a broad, sheetlike deposit across northwestern and central Arizona that averages about 300 feet (90 m) thick in Grand Canyon and 150 feet (45 m) thick across the Verde and Chino valleys.

Sandstone is the dominant lithology in the Manakacha Formation (Figs. 9.4 and 9.5). The composition ranges from very fine- to medium-grained quartz

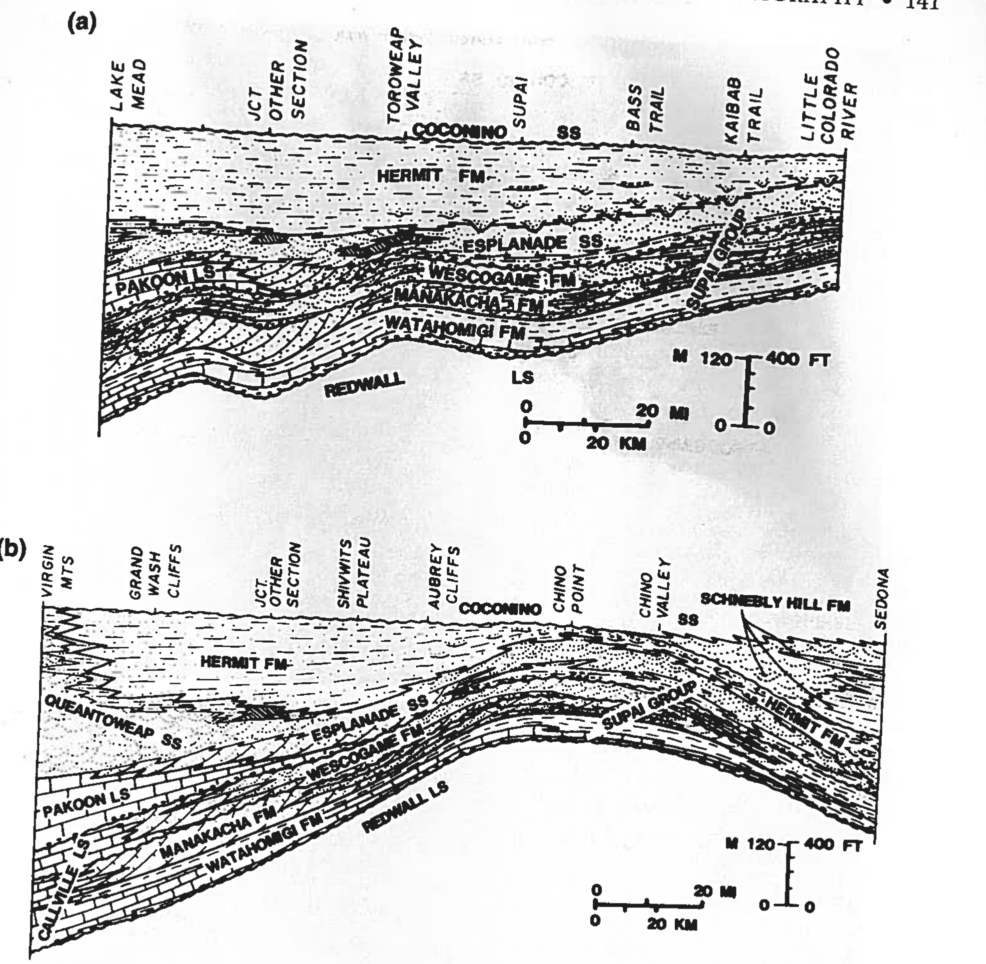

FIGURE 9.5. Restored stratigraphic cross section of Supai Group, Hermit Formation, and related strata. (a) West to east through Grand Canyon; (b) northwest to southeast from Virgin Mountains to Sedona. See Figure 9.4 for list of symbols.

grains to ooids, abraded fossils, and peloids. Most units, regardless of composition, are cemented with calcite. Jasper (red chert) is an accessory to many of the more limey units. Based on bedding types, geologists can recognize three kinds of sandstone. Cross-stratified sandstone comprises trough, planar, and compound sets that range in thickness from 1 foot (2.5 cm) to 30 feet (9 m). Careful examination of the strata within the sets reveals the ubiquitous presence of the climbing translatent strata described by Hunter (1977). Climbing translatent strata are thin laminae, generally less than several millimeters thick, that display reverse grading within each lamina. Each lamina displays the migration record of a single wind ripple (Hunter 1977) and, as such, is a powerful indicator of eolian deposition. The orientation of the strata reflects the nature and geometry of the surface on which they were deposited. Horizontal or very low-angle, climbing translatent strata form at the base of dunes,

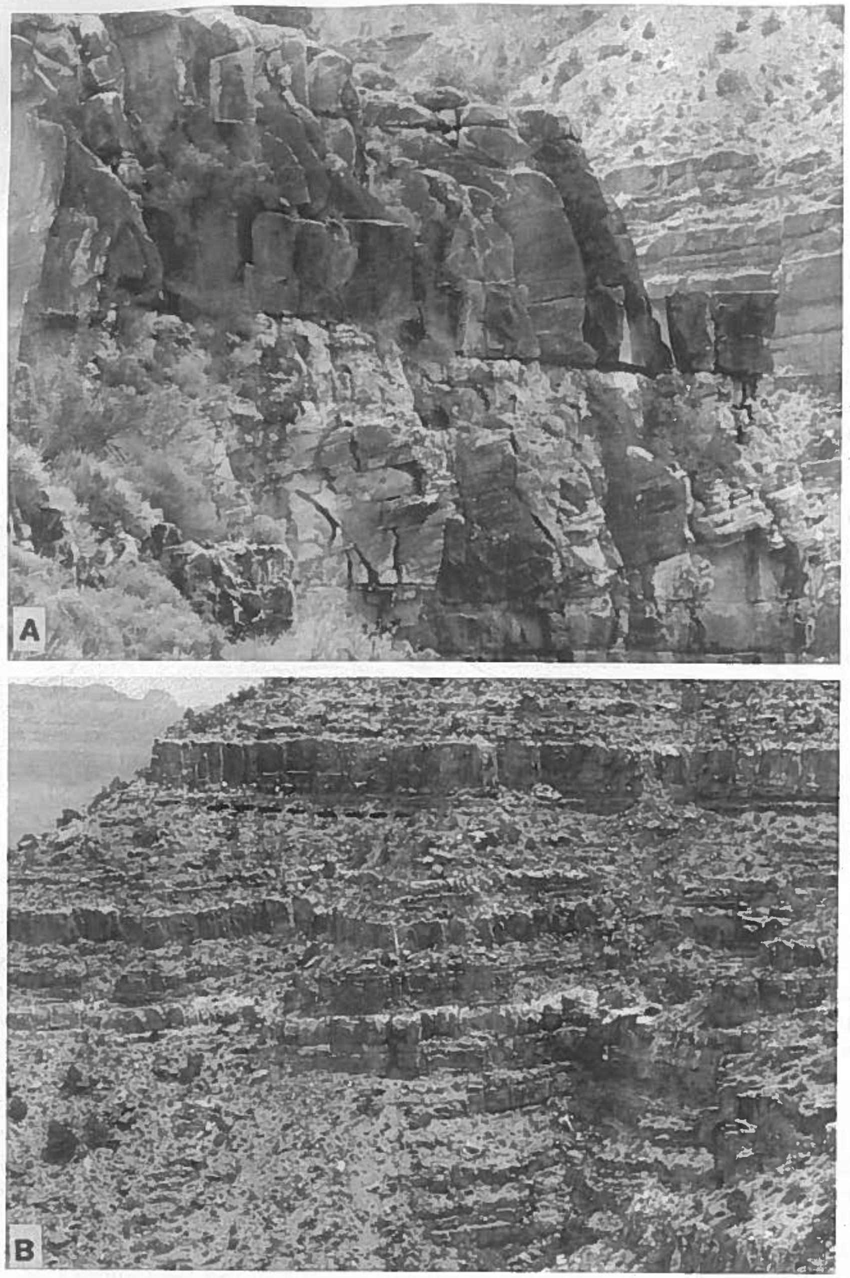

Figure 9.6. Typical outcrops of Manakacha Formation. (a) Close-up showing two large eolian sets (each about 4–5 m thick). (b) Along Hermit Trail; dashed line marks top of formation; Manakacha is about 60 m thick.

between dunes, and on eolian sand sheets. Climbing translatent strata in cross-stratified sets form from the migration of eolian dunes. A broad range of both types of deposition is present within sandstone units of the Manakacha Formation.

A second type of sandstone consists of horizontally laminated to very low-angle, cross-stratified, fine-grained, calcareous, and silty units up to 20 feet (6 m) thick. Generally deeper red in color than cross-stratified units, these units also show rare ripple lamination and locally abundant tracks and trails on bedding planes. The existing evidence about the sandstone's origin suggests a subaqueous genesis, but few other clues are present.

The third type of sandstone typically is very fine-grained and grayish orange to bright reddish orange in color. It ranges from structureless (homogeneous) to extensively bioturbated. Actually, both the structureless and bioturbated sandstone are thought to be the result of bioturbation; however, the former is so extensively bioturbated that individual burrows or trackways cannot be distinguished. Because the latter is less thoroughly disturbed, we can see individual traces clearly.

Many sandstone units in the Manakacha Formation show additional alteration caused by diagenesis—chiefly the growth of calcite and dolomite crystals that alter or destroy the original fabric of the sediment. Such change is most prevalent in the upper portions of the sandstone units.

Mudstone and fine-grained limestone and dolomite are common throughout the Manakacha Formation. The mudstone is usually dark reddish brown in color, and it ranges from structureless to units that contain nodular carbonate concretions. Light gray carbonate units that bear jasper occur fairly commonly; they are seen with or without wispy lamination.

McKee (1982a) recognized a lower cliff unit and upper slope unit throughout Grand Canyon (Fig. 9.6). Sandstone dominates the former and is less abundant in the latter. These subdivisions are not as apparent in the Verde and Chino valleys; units assigned to the Manakacha Formation in these areas form a lower ledge and slope unit, a middle cliff unit, and an upper ledge unit. It should be emphasized that the lower level of relief along the southern margin of the Colorado Plateau (as opposed to the Grand Canyon) undoubtedly affects the topographic expression of all units in the Supai Group.

The Manakacha Formation displays a steady increase in carbonate content to the northwest. Some of these carbonates are cross-bedded, indicating deposition by fluid currents. In the Shivwits Plateau region, McKee (1982a, Fig. 9.8) classified the entire formation as limestone.

The upper contact with the overlying Wescogame Formation marks a regional unconformity, with most of the Middle Pennsylvanian absent. Conglomerate derived from the underlying Manakacha Formation, accompanied by scouring and channeling, marks the boundary. Poorer outcrops along the southern Colorado Plateau in the western Mogollon Rim region make this contact very difficult to locate.

Wescogame Formation

The Wescogame forms a sheetlike unit that ranges from 100 to 200 feet (30 to 60 m) thick across the Grand Canyon and the western Mogollon Rim region. Lithologic components are similar to those described above for the Manakacha Formation (Figs. 9.4, 9.5, and 9.7). Sandstone dominates the Wescogame throughout much of the central Grand Canyon and in western Chino Valley; to the east,

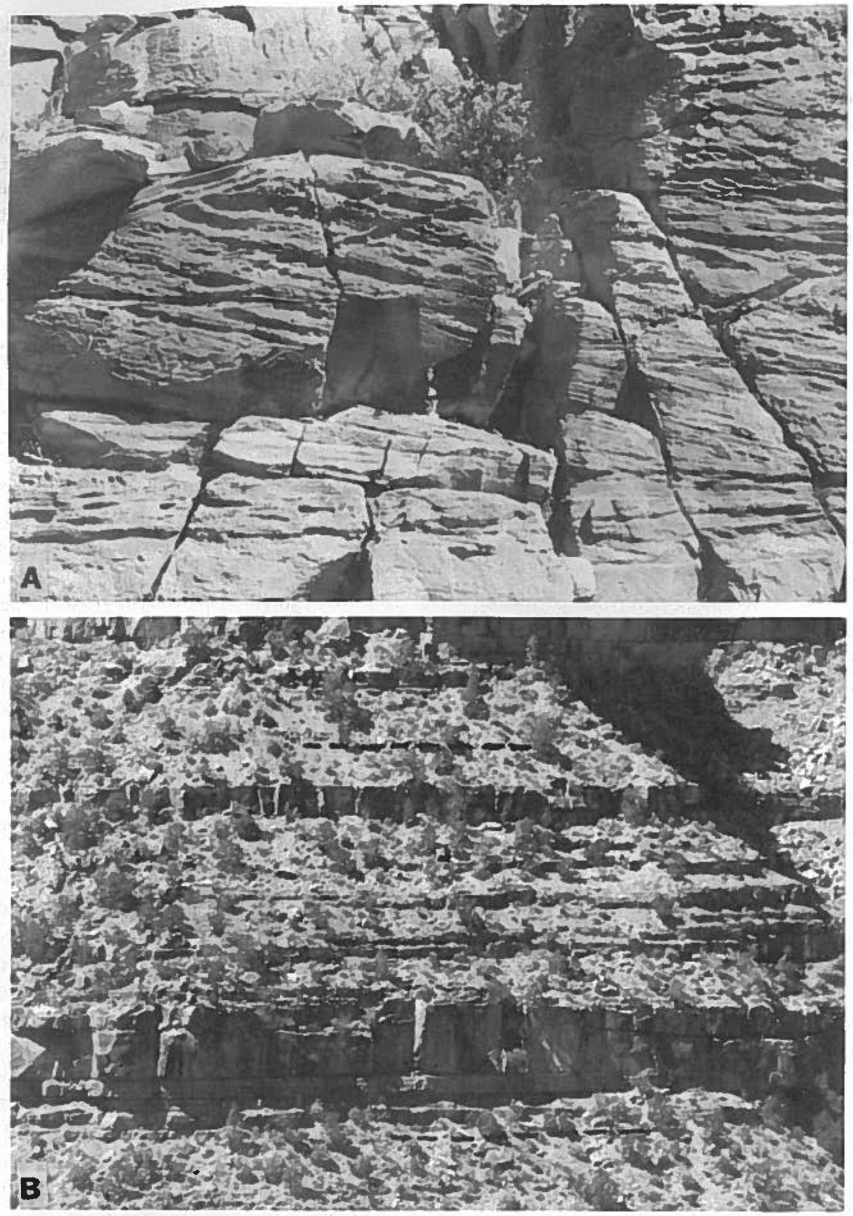

Figure 9.7. Typical outcrops of Wescogame Formation. (a) Close-up showing trough cross-bedding composed of eolian wind-ripple strata; about 2 m of section shown. (b) Along Kaibab Trail; dashed lines mark base and top of formation; Wescogame is about 40 m thick.

mudstone increases, whereas limestone increases west of Chino Valley (Blakey 1980; McKee 1982a).

McKee (1982a) informally divided the Wescogame Formation into a lower cliff unit and an upper slope unit, both traceable throughout much of the Grand Canyon. In the western part of the Mogollon Rim region, the Wescogame typically forms steep, ledgy slopes. Contrary to the implications of McKee's work, the location of the boundary between the Manakacha and Wescogame formations can be difficult to determine, both from a distance and from close range. The unconformity between the two formations varies, and even where noticeable relief is present, the boundary tends to be in a slope- or ledge-forming horizon, where debris from overlying units obscures the contact. The thickness and expression of associated conglomerate also varies and cannot be used to define the horizon in some areas. The topographic criteria provided by McKee (1982a) form the most reliable method of separating the two formations. The contact lies near the top of the slope that typically forms the upper part of the Manakacha Formation, a few feet below the overlying cliff unit of the Wescogame Formation.

The upper contact of the Wescogame Formation almost certainly marks the disconformity associated with the Pennsylvanian–Permian boundary (McKee 1982a). Although some of the problems discussed above also pertain to locating this contact, the top of the Wescogame Formation generally is easy to locate. Wherever exposures are clear, conglomerate and relief of up to 50 feet (15 m) mark the boundary. Based on regional patterns in the Grand Canyon and in the western Mogollon Rim region, geologists have defined three broad, north-northeast-trending facies belts. The eastern belt, the redbed facies, lies east of a line from near Hermit Basin to the western Verde Valley. It comprises subequal amounts of sandstone and red mudstone. The middle belt, the sandstone facies, extends west to a line running through the Shivwits Plateau region. The westernmost belt is chiefly cross-stratified limestone and limey sandstone. It forms the limestone facies.

Esplanade Sandstone (and Pakoon Limestone)

Sandwiched between the steep ledges and slopes of the underlying Wescogame Formation and the smooth slopes of the overlying Hermit Formation, the Esplanade Sandstone forms one of the most distinctive horizons in the Grand Canyon, especially in the central and western portions. The Esplanade Sandstone contains the highest percentage of sandstone of any of the units in the Supai Group (Figs. 9.4, 9.5, and 9.8). The formation consists of a steadily northwestward-thickening wedge that ranges from 200 to 250 feet (60 to 75 m) thick in the eastern Grand Canyon and western Mogollon Rim region to over 800 feet (240 m) in the western Grand Canyon and along the Grand Wash Cliffs (Blakey 1980; McKee 1982a). The thickness in the western Grand Canyon region includes the Pakoon Limestone.

Although the lithologic components are the same as those previously described in the Manakacha Formation, the dominance of cross-stratified sandstone in many sections distinguishes the Esplanade Sandstone. Most cross-stratified units are dominated by climbing translatent strata, which strongly suggests an eolian origin for the unit. The sandstone units forms beds generally 5 to 50 feet (1.5 to 15 m) thick that are separated by thin, red mudstone; by fine-grained carbonate units; or by prominent, irregular-bedding planes. These planes cause the Esplanade to weather to an irregular or ledgy cliff. Clean outcrops range in color from pale grayish orange to pale reddish orange, but staining from the overlying redbeds in the Hermit Formation causes the unit to appear dark reddish brown.

McKee (1982a) informally divided the Esplanade into a basal slope unit and main cliff unit. Typically, the upper part of the main cliff weathers back into a series of ledges, or ledges and slopes. In western portions of the Grand Canyon,

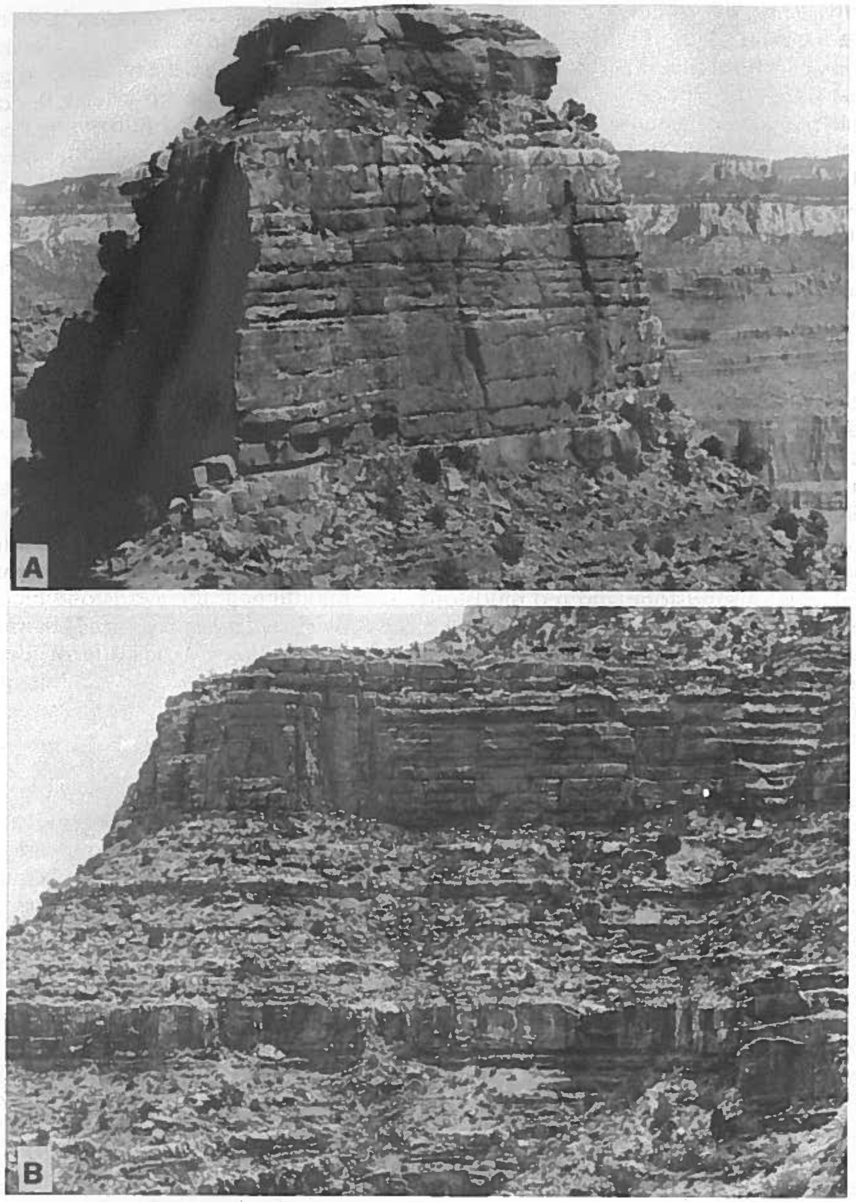

Figure 9.8. Typical outcrops of Esplanade Sandstone. (a) O'Niell Butte on Kaibab Trail. (b) On Hermit Trail. Most ledges are composed of eolian sandstone. Dashed lines mark base and top of Esplanade, which is about 90 m thick at both locations.

the basal slope unit grades westward into the dolomite and limestone of the Pakoon Limestone. The Pakoon thickens from east to west because of increased subsidence and the replacement of siliciclastic units by carbonate units in this direction. The Pakoon Limestone is approximately 300 feet (90 m) thick along the Grand Wash Cliffs. McKee chose not to include the Pakoon formally within

the Supai Group, but I would suggest that future stratigraphic work in the area consider assigning the Pakoon Limestone to the Supai Group.

The nature of the upper contact of the Esplanade Sandstone with the overlying Hermit Formation is part of a complex regional problem. McKee (1982a, pp. 169–171, 202–203) examined the nature of the contact and concluded that overall evidence suggested the presence of a regional unconformity. At many locations, there is a definite transition between the Esplanade below and the Hermit above. At other locations, the Esplanade is overlain by an erosion surface with 30 to 50 feet (9 to 15 m) of local relief. The Hermit overlies that surface. Still other locations exhibit channeling into the Esplanade, but the channels originate from within the Hermit and not from the base. This suggests that the erosion surfaces formed after the deposition of the Hermit began. In a few places (such as near Toroweap, in the Shivwits area, and around Sedona), cross-stratified sandstone typical of the Esplanade occurs as much as 100 feet (30 m) above the Hermit–Esplanade contact. This suggests that the contact is variable from place to place and that it represents a zone of transition. The erosional relief also may be related to channel cut-and-fill sequences and not to a major regional unconformity. Obviously, there is a great deal that geologists do not know about this contact zone.

The most dramatic facies change in the Esplanade Sandstone occurs at the base, where the lower slope unit grades westward into the Pakoon Limestone in the Shivwits Plateau area (McKee 1982a). The upper part of the Esplanade displays a gradual change to increased carbonate content in the west. This change, however, is less sharp than that which occurs in the Manakacha and Westcogame formations. Bedded gypsum occurs near the top of the Esplanade in the Toroweap area (Fig. 9.5). The gypsum is intercalated with cross-stratified sandstone throughout a zone as much as 200 feet (60 m) thick.

Hermit Formation

Its overall fine-grained nature and relatively poor exposures, as well as a general lack of interest in the unit among geologists, contribute to making the Hermit Formation one of the poorest known units in Grand Canyon. White (1929) provided a comprehensive discussion of the formation, but it centered on the flora. Many other workers have provided local descriptions, but no major, recent, stratigraphic or sedimentologic study is available. This chapter will provide both sedimentologic and stratigraphic description—but primarily at the reconnaissance level.

For the most part, the Hermit Formation is composed of slope-forming, reddish brown siltstone, mudstone, and very fine-grained sandstone. The formation varies from approximately 100 feet (30 m) in thickness in the eastern portion of the Grand Canyon and near Seligman to over 900 feet (270 m) in the Toroweap and Shivwits Plateau areas (McNair 1951). Northward along the Hurricane Cliffs and into adjacent Utah, the sandstone content increases. Here, the Hermit interval generally is assigned to the Queantoweap Sandstone. In the western Mogollon Rim region east of Seligman, the Hermit can be traced eastward through discontinuous outcrops to the Sedona area, where it averages 300 feet (90 m) in thickness (Blakey 1980, 1990; Blakey and Knepp 1988).

McNair (1951) pointed out the misnomer of "shale" in the Hermit and proposed the word "formation" instead. Many subsequent workers have followed this suggestion. Overall, the Hermit Formation comprises a heterogeneous, chiefly fine-grained, siliciclastic sequence (Figs. 9.4, 9.5, and 9.9). Neither the vertical

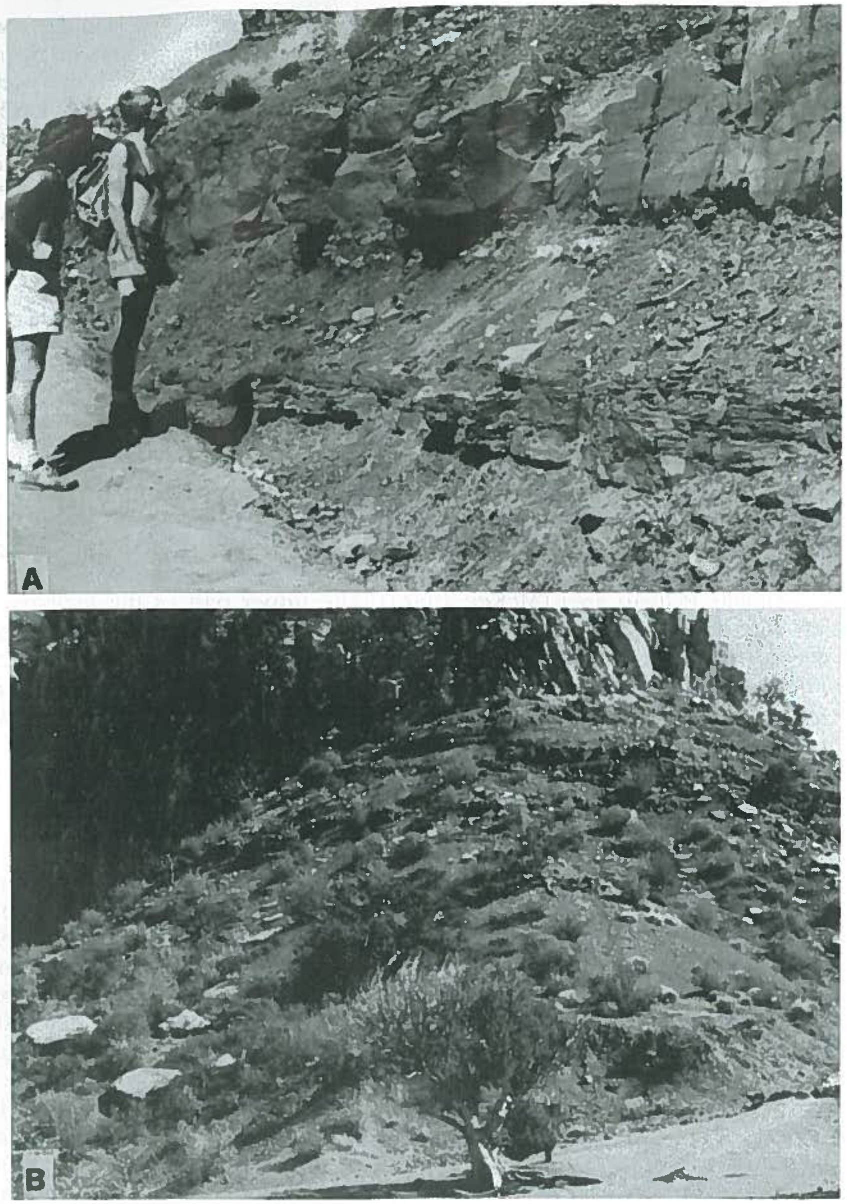

FIGURE 9.9. Typical outcrops of Hermit Formation, Kaibab Trail. (a) Close-up of ledges (fluvial channels and slopes (overbank deposits). (b) Typical slope-forming nature.

nor the areal distribution of the various components defines any sharp trends, though regional variations are apparent. Most of the lithologic components of the Hermit are present to some extent in the underlying Supai Group. Silty sandstone and sandy mudstone are the most widespread and voluminous components in the Hermit. Silty sandstone ranges from structureless to ripple laminated to trough cross-stratified. Structureless units make up ledge-forming beds that average 3 feet (1 m) in thickness and may or may not contain limy, nodular concretions. Ripple-laminated units display subaqueous, faint-to-prominent ripple cross-lamination. Trough cross-stratified sandstone consists of troughs up to several feet across (which may be organized within gently dipping master sets in which the troughs are roughly perpendicular to the included master sets). Termed epsilon cross-stratification, this structure is associated commonly with point-bar deposition in meandering streams. Rare trough to planar–tabular sets of cross-stratified sandstone, fine-grained and well-sorted, with climbing translatent strata, occur near the base of the Hermit at several localities within Grand Canyon and in the Sedona area.

Sandy mudstone forms slopes in the Hermit Formation. The mudstone commonly is featureless, though rare clean rock outcrops display fine ripple lamination and calcareous nodular concretions. A minor component overall, but locally abundant, is an intraformational conglomerate composed of sedimentary pebbles. Most pebbles are carbonate grains obviously derived from the adjacent carbonate concretions contained within intercalated sandstone and mudstone, but some fine-grained, limey sandstone and siltstone pebbles also are present. The conglomerate occurs both as beds and as an accessory to sandstone units. McKee (1982a) reported conglomerate in the Hermit at 14 of his 35 localities in Grand Canyon; it is most abundant in the Sedona area, where 10 to 15 horizons are typical within the 300-foot (90-m)-thick section.

Most of the Hermit Formation consists of interbedded, silty sandstone and sandy mudstone. At most locations, sandstone is more abundant near the base of the formation, and mudstone increases upward. The two lithologies are interbedded rhythmically, typically with 15 or more cycles present in most sections. The extent or geometry of individual sandstone bodies is uncertain because of poor exposures. Reconnaissance work suggests an increase in sandstone in the Shivwits Plateau area. This may reflect the increase in sandstone to the northwest, where the Hermit is believed to grade into the Queantoweap Sandstone (Blakey and Knepp 1988). Conglomerate is most abundant in the Sedona area. Current evidence indicates that it decreases in all directions from there.

The upper contact of the Hermit Formation is everywhere sharp and without gradation of any kind. Throughout Grand Canyon and into the Aubrey Cliffs, it is overlain by the Coconino Sandstone. Cracks 20 or more feet deep at the top of the Hermit frequently are filled with the overlying sandstone.

East of Seligman into the western Mogollon Rim region, a substantial sequence of strata not present in the Grand Canyon appears above the Hermit; these strata are defined as the Schnebly Hill formation. The regional relations shown in Fig. 9.5b are discussed in detail by Blakey (1980, 1990) and by Blakey and Knepp (1988). Contact with the overlying Schnebly Hill Formation in this area also is sharp, but it lacks the sandstone-filled cracks. Based on regional stratigraphic relations (Blakey 1996), the sharp contact is interpreted as a major regional unconformity, though sufficient paleontological evidence to confirm this hypothesis is not yet available.

PALEONTOLOGY

Both White (1929) and McKee (1982a) have examined the faunal and floral content of the Supai Group and the Hermit Formation. Overall, both stratigraphic units are sparsely and sporadically fossiliferous, but we have recorded a broad variety of fossil types.

Trace Fossils

Trace fossils—the trackways, burrows, impressions, resting marks, and feeding marks formed by organisms in sediment and preserved in the rock record—are ubiquitous throughout the Supai and the Hermit. McKee reported their presence, but no comprehensive report on trace fossils has yet appeared. Nondescript bioturbation is the most common form of trace fossil in the Supai Group. The swirls, crinkles, and irregular patterns associated with distinct burrows attest to its abundance. Rocks showing disturbed sediment range from structureless (complete bioturbation) to distinctly burrowed. Burrowed structures are common both in the Supai and the Hermit, most frequently in silty sandstone and carbonate units. Many types of burrows are present, but those occurring most frequently are cylindrical, smooth burrows several millimeters in diameter that are parallel, perpendicular, or oblique to the bedding. McKee also described trackways, mostly attributed to vertebrates, from both the Wescogame Formation and the Esplanade Sandstone. Other miscellaneous trace fossils, including invertebrate trackways, resting marks, bioturbation caused by plant roots, and possible feeding marks, are distributed throughout the Supai Group and the Hermit Formation. Clearly, this is a field with strong future research potential.

Fossil Invertebrates

McKee (1982a, Chapters E and F) described and pictured many of the invertebrates in the Supai Group. He reported a fairly rich and varied brachiopod fauna from the Watahomigi Formation. Marine fossils, including locally abundant fusulinids, occur sporadically throughout the Supai—especially in western Grand Canyon. However, the presence of abraded fossils in well-sorted, cross-stratified sandstone or limestone should not be used as a criterion for marine deposition because skeletal grains can be, and were, blown inland and incorporated into eolian deposits.

Flora

Plant remains are widespread but sparsely distributed throughout the Supai Group and the Hermit Formation. The comprehensive reports by White (1929) and McKee (1982a) list the location, stratigraphic horizon, and plant taxa found throughout the units, and Blakey (1980) reported plants fossils from the Hermit formation in the Sedona area. McKee (1982a, p. 100) found that although most of the plant remains have little stratigraphic value, they do suggest the presence of broad floodplains developed during times of regressing seas and semiarid-to-arid climates.

AGE AND CORRELATION

Introduction

McKee (1975; 1982a) provided us with the first detailed regional account of the age of the Supai Group. He based his age determinations primarily on brachiopods in the Watahomigi Formation and fusulinids scattered throughout the group. Using this information, he assigned to the formations of the Supai Group the following ages: Watahomigi Formation—Morrowan and Atokan; Manakacha Formation—Atokan; Wescogame Formation—Virgilian; Pakoon Limestone—Wolfcampian; Esplanade Sandstone—Wolfcampian (lower part of formation). The

age of the upper part of the Esplanade Sandstone is determined by its stratigraphic position.

White (1929, p. 40) assigned the Hermit formation an age of upper Lower Permian (Leonardian); McKee's work seems to confirm this. He used a fossil plant, *Callipteris arizonae*, found in the Bone Spring Limestone of Leonardian age in southeastern New Mexico, as his point of reference. If McKee is correct, the age of the Esplanade Sandstone must be no older than Wolfcampian and no younger than Leonardian.

In the Sedona area, the Esplanade Sandstone and the Hermit Formation lie several hundred feet below the Fort Apache Member of the Schnebly Hill formation. The Fort Apache Member has yielded conodonts of early late Leonardian age. This paleontologic evidence provides additional support for presuming an early Leonardian age for the Hermit Formation (Blakey and Knepp 1988).

Regional and Worldwide Correlation

In a 1988 study, Blakey and Knepp summarized the regional correlation of Pennsylvanian and Permian rocks of Arizona and adjacent areas. Based on age assignments presented above and using the worldwide geologic time scale of Van Eysinga (1975), they found that the Supai Group correlates with the Westphalian, Stephanian, and Sakmarian.

The Hermit Formation correlates with the Artinskian, which occurs in the standard Carboniferous and Permian units of Europe. In addition, both the Esplanade Formation and the Hermit Formation correlate with the eolian-bearing Rotliegendes of northern Europe.

ORIGIN OF DEPOSITS IN SUPAI AND HERMIT

Introduction

Interpretation of the depositional history of Pennsylvanian and Permian redbeds in the Grand Canyon region is plagued by a variety of problems—including sparsity of fossils, the complex cyclic nature of the strata, the fine-grained nature of some units, extensive diagenesis, inaccessibility of some outcrops, and some misconceptions concerning regional stratigraphy and depositional history.

Despite these problems, recent and detailed sedimentologic work (much of it previously unpublished) suggests the need to revise our interpretation of the depositional history of the Supai Group. Earlier studies suggested a fluvial, deltaic, beach, shallow-marine, or estuarine origin for Supai sandstone bodies. New evidence indicates that an eolian origin is much more likely. McKee (1982a) did not consider an eolian origin for any sections of the Supai Group. The presence of carbonate grains, including locally abundant fossil fragments, may have prevented his recognition of the eolian environment. Desert varnish and extensive diagenesis of some beds further obscure some important eolian characteristics.

Eolian Characteristics

Until fairly recently, eolian criteria consisted chiefly of the textural maturity of quartz sand, coupled with thick set size and a high-angle, cross-strata dip. Although many ancient eolian sandstones display these characteristics (the Coconino Sandstone, for example), these criteria are unreliable. Some eolian sandstones lack large-scale, high-angle sets. In some cases, sandstones with these characteristics are noneolian. Hunter (1977) found that the most reliable eolian indicators are not set size, angle of dip, or even cross-stratification, but the nature of the individual laminae and strata that compose the sandstone. Migrating and climbing wind ripples, regardless of their orientation, form the inversely graded laminae (generally less than several millimeters thick) that Hunter termed "climbing translatent strata." This strata type can form and be preserved in many parts of the eolian environmental—including on most parts of dunes, between dunes (interedunes), and on duneless areas of blowing sand termed sandsheets.

Many dunes contain one or more active slipfaces that are characterized by the periodic avalanching of loose sand. Hunter (1977) described deposits formed by this process as structureless strata (as much as several centimeters thick) at or near the angle of repose of dry sand (30–34°, though generally 24–26° in ancient sandstones because of compaction). He termed these "sandflow strata." Some small dunes, and most dunes without slip faces, may leave only climbing translatent strata in variously shaped sets with widely ranging angles of dip. Particularly common are trough-shaped sets several feet thick and tens of feet wide that were produced by crescentic-shaped dunes or troughlike scours (blowouts) on dunes or sandsheets. Many dunes with slip faces produce avalanche strata that fade out down the dune, where they are replaced by climbing translatent strata. This toe of the avalanche, which is termed "sandflow toe," is an easily observed eolian indicator.

Eolian sandstones typically consist of one or more facies that relate to their type and position within the eolian sand sea (erg). The central portion of most sand seas contains dunes and larger bedforms called draas. Marginal sand seas include small, possibly scattered dunes and sandsheets, and the most distal areas may consist of broad sandsheets, and the most distal areas may consist of broad sandsheets with few, if any, dunes. Sandsheets and small dune fields may filter into adjacent environments, such as fluvial plains, dry lake beds, and coastal plains. Facies formed in central sand–sea areas typically consist of cross-stratified sandstone comprised of sandflow strata and climbing translatent strata with interdune deposits of horizontal climbing translatent strata or various types of structureless to wavy-bedded sandstone or silty sandstone. Marginal sand–sea deposits typically display smaller sets of cross-stratified sandstone and sandsheet deposits. These consist of horizontal to irregular beds of climbing translatent strata. The most complete discussion of the distribution of all known ancient eolian sandstone units and their facies can be found in Blakey et al. (1988).

Controls on Deposition

Controls on deposition during Supai and Hermit time were varied and complex. They included a complicated tectonic regime, rapid rise and fall in sea levels, and an influx of sand from the north.

The tectonic elements that affected sedimentation during the late Paleozoic have been addressed in several recent papers (Blakey 1980, 1988; Blakey and Knepp 1988). Areas in which the crust subsides relatively rapidly have thicker deposits than do areas of less subsidence. The former tend to have more marine deposits and fewer redbeds. Areas of less subsidence are characterized by thin, red, terrigenous clastic deposits with few fossils and abundant unconformities. During the late Paleozoic and early Mesozoic in the Southwest, such deposits tended to form in fluvial, lacustrine (lake), and coastal-plain environments.

Areas of greater subsidence are characterized by deposits of limestone and thin sandstone and mudstone. They typically are grayish in color and fossiliferous. Marine and shoreline deposits dominate.

Geologists classify the resulting tectonic elements by their shape and their relation to adjacent elements. Circular to oblate areas with relatively high rates of subsidence are termed *basins.* Areas of intermediate subsidence rates covering broad areas are called *shelves,* and those more parabolic in shape are termed *embayments.* Arches are areas of little subsidence; typically, they interrupt brief periods of slight uplift. Areas dominated by uplift are referred to as upwarps or uplifts.

During the formative period of the Supai and Hermit, northern Arizona was bisected by a northeast-trending element of relatively little subsidence, the Sedona arch (Fig. 9.10). To the northwest, the Grand Canyon embayment domi-

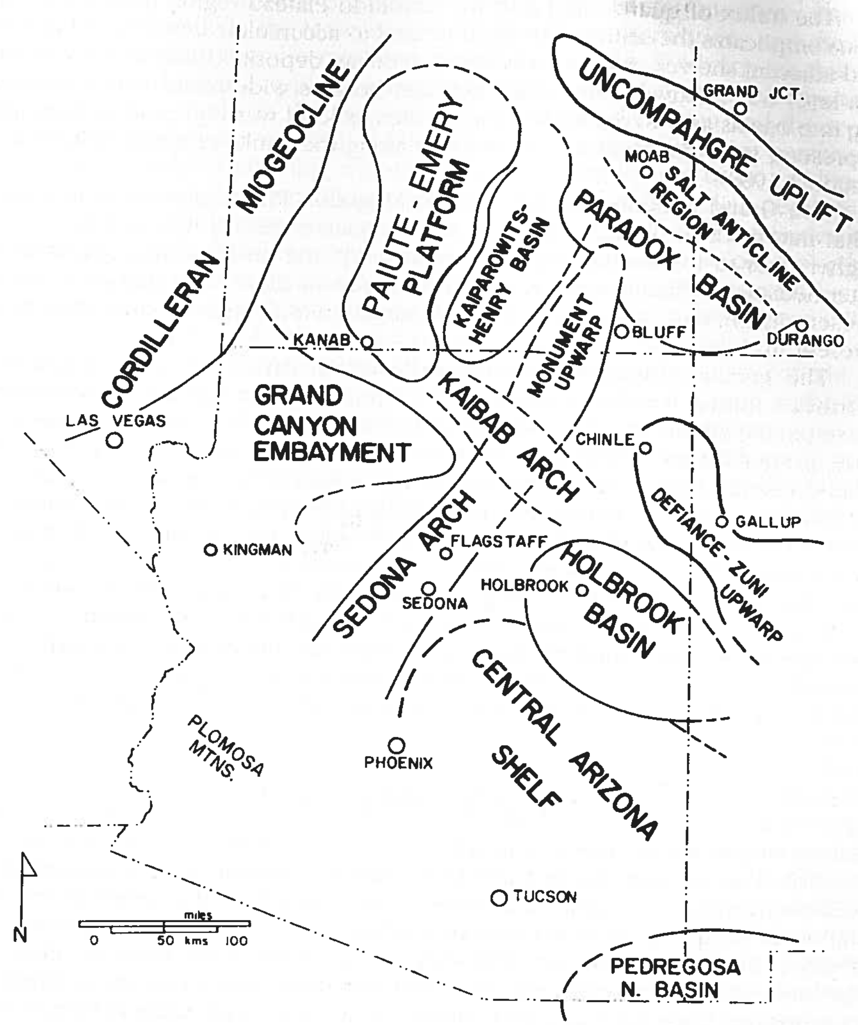

Figure 9.10. Map showing late Paleozoic tectonic elements of northern Arizona and vicinity.

nated the Grand Canyon region. East-central Arizona was the site of the Holbrook basin and the Mogollon shelf. Northeastern Arizona was dominated by the Defiance uplift during the Pennsylvanian or the Defiance arch in the Permian. To the north, in southeastern Utah, was the Paradox basin. During periods of relatively high worldwide sea level, the seas tended to encroach on Arizona from the west into the Grand Canyon embayment, from the east and south across the Mogollon shelf, and from the northeast into the Paradox basin. Although the uplifted areas (i.e., arches) received some sediment, much of it was nonmarine. Because great fluctuations of sea level (chiefly because of glaciation in the southern hemisphere) characterized the late Paleozoic, deposits tended to be cyclical in nature. Marine-dominated cycles occurred in basins, and continental-dominated cycles occurred across the arches. Movement on arches, uplifts, and basins altered a landscape already shaped by changes in sea level. Therefore, the resulting pattern of deposition is very complex.

The influx of quartz sand into the Colorado Plateau region from the north also complicates the setting. The sand tended to accumulate between the arches and adjacent shelves or basins as extensive eolian deposits (Blakey 1988). When sea level was relatively low, eolian deposits became widespread, often expanding into basins and across shelves; when the sea level was high, eolian deposits, if present, were confined to narrow belts along the flanks of arches (Chan and Kocurek 1988).

The Grand Canyon and the western Mogollon Rim region lay in a somewhat intermediate position between more negative areas to the northwest and higher ones to the southeast. Not unexpectedly, the resulting sedimentation is intermediate in character; cyclic sedimentation with more continental aspects is present in the east, whereas cyclic sedimentation with more marine aspects is present in the west.

The general structural grain of northwestern Arizona tended southwest–northeast during Supai and Hermit time. Shorelines and marine trends should have paralleled this pattern, and fluvial deposits should have flowed northwesterly, perpendicular to this grain. Indeed, marine and shoreline patterns do parallel the trend, but the sandstone bodies that McKee (1982a) suggested might be of fluvial origin show a paleocurrent reading to the south and southeast; this is as much as 180° from the expected trend. We have not documented the trends of Hermit Formation sandstone bodies.

Any one of the above controls on deposition was sufficient to produce a complicated depositional pattern. Together, they combined to produce a heterolithic record that varies abruptly, both vertically and laterally. The summary presented below must be considered preliminary. We have much to learn about the depositional history of the Supai Group and the Hermit Formation.

Depositional Environments

Recent studies tend to confirm the long-standing belief that the Supai Group was deposited on a broad coastal plain. Individual depositional settings ranged from shallow marine to continental. However, the recent identification of eolian-deposited sandstone units throughout the section necessitates a reinterpretation of Supai depositional history. Most of our knowledge of the Supai is based on the work of McKee (1982a, pp. 260, 261), who provided a general description of sandstone bodies in the Supai Group. He interpreted the large foreset planes in the Manakacha Formation as the fronts of large, subaqueous sand sheets or small Gilbert deltas in areas strongly affected by tides. Traditionally, geologists have interpreted sandstone bodies of the Wescogame Formation as having formed in a high-energy, fluvial environment. Marine bioclastic debris was thought to have been trapped in an estuarine setting.

McKee found the Esplanade the most difficult unit to interpret and offered a compromise of mixed fluvial, estuarine, and shallow-marine origins. It should be pointed out that although McKee's Supai publication appeared in 1982, the work actually started in the late 1930s and continued into the 1970s. Delays in publication created a large gap between the time when the work was written and the time it was published. Therefore, much of the state-of-the-art sedimentology of the late 1970s and the 1980s is not reflected in the publication; this is especially true of eolian sedimentology. McKee's monograph will continue to be a valuable source of information concerning the Grand Canyon Supai, but some of the interpretations, particularly concerning depositional environments, must be considered out of date.

Given the eolian criteria discussed earlier in this chapter, most cross-stratified sandstone units in the Manakacha and Wescogame formations and the Esplanade Sandstone now are regarded as eolian in origin. Evidence for this includes the fact that wind-ripple laminae (climbing translatent strata) are ubiquitous throughout these units in the eastern and central Grand Canyon and the western Mogollon Rim regions (Fig. 9.11). Most sets of cross-strata are composed chiefly or entirely of climbing translatent strata. There is little or no evidence, however, of avalanche-formed sand-flow strata. Studies of the eolian deposits in the Supai have not progressed enough to explain this, and several theories are possible. Kocurek (1986) reported that dunes with broad aprons (plinths) commonly do not preserve slipface deposits in the sedimentological record; longitudinal dunes, star dunes, and oblique dunes commonly have large aprons. Reversing dunes also tend to lack slipface deposits. Other late Paleozoic eolian deposits contain a high percentage of wind-ripple strata. They also contain at least locally abundant sand-flow strata (Blakey et al. 1988). But only in the Esplanade Sandstone of the Oak Creek Canyon area are sand-flow strata developed extensively in the Supai Group. Some cross-stratified sets (perhaps 40 percent in Grand Canyon and 60 percent in the western Mogollon Rim) have been so affected by diagenesis that the stratification type cannot be determined. A portion of these could contain sand-flow strata, and some sets could be noneolian in origin; however, close association with, and similarity to, sets with wind-ripple laminae strongly suggest that many or most unidentifiable sets formed as climbing wind-ripple deposits.

The geometry of eolian sandstone bodies in the Supai Group is somewhat unusual. Many well-known eolian sandstones, such as the Coconino Sandstone and the Navajo Sandstone, are mostly or entirely cross-stratified sandstone. Redbeds, conglomerate, and limestone are rare or absent. Blakey et al. (1988) have pointed out the problems of stereotyping eolian deposits as simple, pure sandstone bodies. Most are not. Individual eolian bodies range from irregular sheets to more common broad lenses, to wedges, and even small pods (Fig. 9.12). Bodies are separated from each other by red siltstone and mudstone, noneolian sandstone, limestone, and, rarely, conglomerate. Undoubtedly, this heterogeneous assemblage of lithologies and complex geometry is the chief reason that the Supai has not been interpreted previously as eolian in origin. Coupled with the high carbonate content and locally abundant sand-sized marine fossil grains in some eolian sandstone, the Supai eolian bodies are far removed from the "typical" eolian sandstone.

The origin of some noneolian deposits in the Supai Group is not clear. Bioturbated sandstone can form in any number of environments—including eolian,

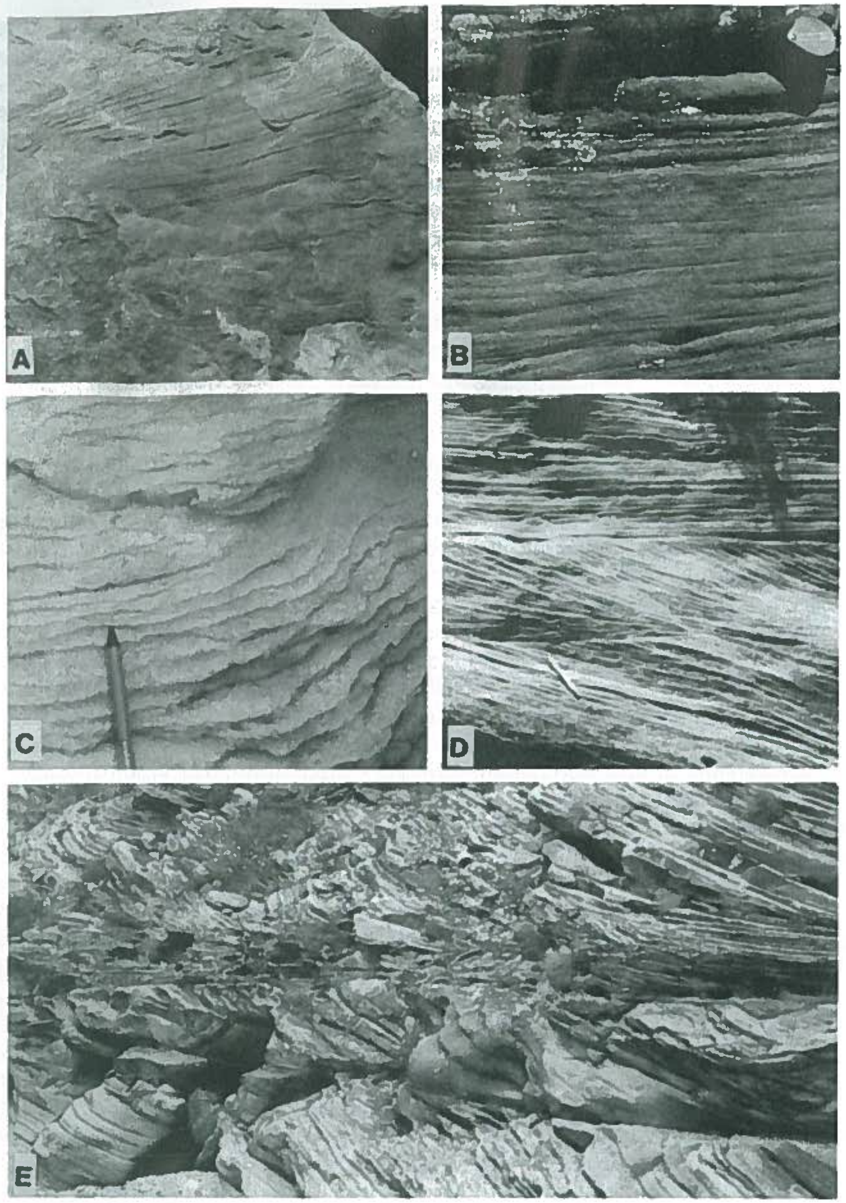

FIGURE 9.11. Eolian strata in Supai Group. (a–d) Eolian wind-ripple (climbing translatent) strata. (e) Eolian sand-flow (avalanche) strata.

fluvial, shoreline, and shallow marine. It is possible that this material has been reworked by organisms in a marine or continental setting. Horizontally laminated to low-angle, cross-stratified sandstone may have formed as plane-bed laminae in shallow streams or in beach swash zones, or as shallow-marine storm deposits. Fine-grained redbeds are particularly enigmatic. Their potential origin can range from fluvial flood plain to lacustrine and sabkha; through low-energy, shoreline (supratidal mudflats, lagoons); and into nearshore, shallow marine.

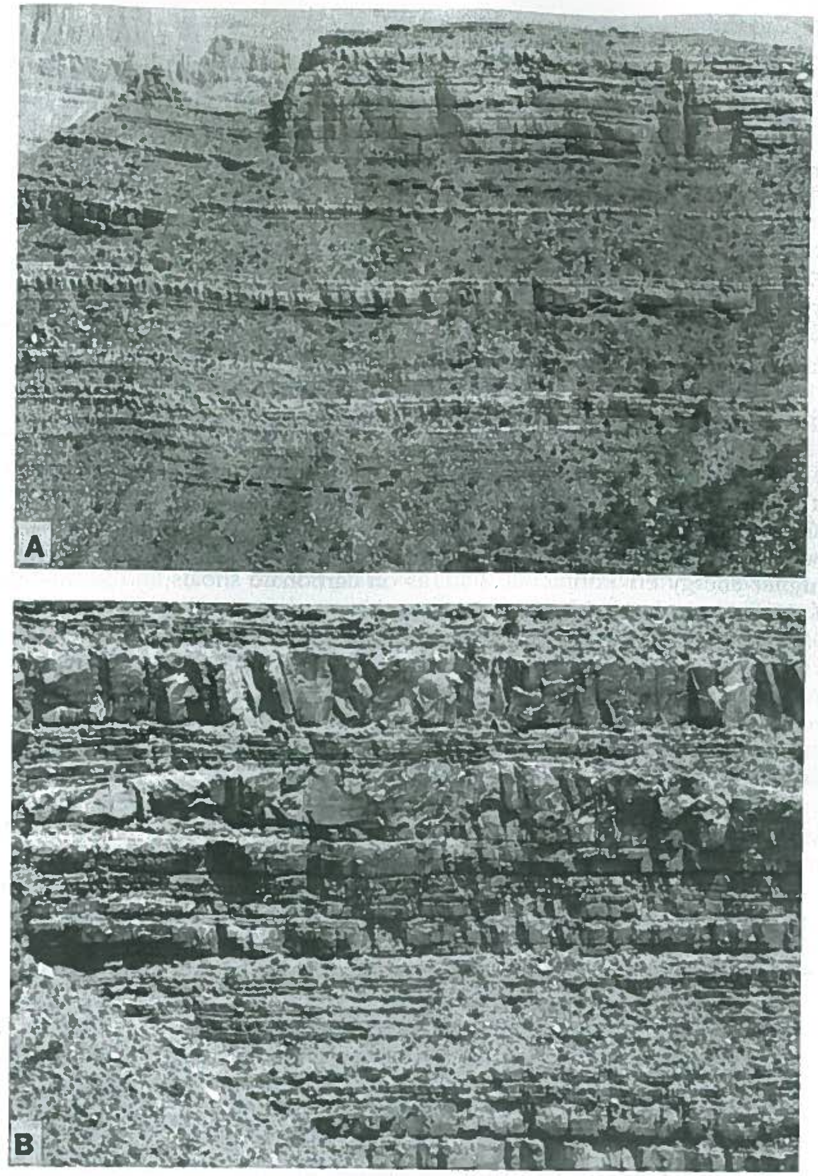

FIGURE 9.12. Eolian sandstone bodies in Supai Group. (a) Manakacha, Wescogame, and Esplanade along Hermit Trail. Dashed lines separate the formations. Most cliff units are composed of eolian sandstone. (b) Manakacha Formation on Hermit trail showing lens-like eolian sandstone bodies separated by thin sandstone and mudstone of uncertain origin. Thickest sand bodies are about 6 m thick.

Because the lower contacts of eolian sandstone bodies are sharp and probably disconformable (and upper contacts commonly are the same), stratigraphic positions an unreliable criteria for diagnosing the environmental of noneolian deposits. At most locations, a few sequences display the following cyclic pattern: sharp basal contact abruptly succeeded by cross-stratified, eolian sandstone; the eolian sandstone grades upward into bioturbated, micritic limestone that is overlain conformably by red mudstone, horizontally stratified sandstone, or a combination of these. Unfortunately, it cannot be determined whether these cycles are transgressive, regressive, or totally continental in origin. If transgressive, the most logical interpretation would be eolian, succeeded by low-energy shoreline, to local beach deposits. If regressive, the eolian likely would be succeeded by fluvial and lacustrine or sabkha deposits. If the cycle is continental (and not necessarily related to advance or retreat of the sea), the succession likely represents ingress and egress of fluvial and lacustrine environments into the eolian dune complex. Clearly, more work is needed on this topic.

Evaporite deposits in the Esplanade Sandstone, chiefly gypsum, could have formed in either coastal or continental sabkha environments. The evaporits are interbedded with redbed and eolian sandstone in the Toroweap area of western Grand Canyon.

We know now that not all carbonate units in the Supai Group are of marine origin; some arenaceous, cross-stratified limestones are of eolian origin. However, many Supai carbonates clearly are of shallow-marine origin. Micrite-rich, fossiliferous carbonates were deposited in open-marine, relatively low-energy environments (McKee 1982a, p. 345). Carbonate grainstone and packstone formed in higher-energy environments, such as on carbonate shoals and in tidal channels. Unfortunately, the detailed relations between marine limestone and eolian sandstone and arenaceous limestone are yet to be determined. McKee (1982a) has documented the abundance of marine carbonate in western facies of the Manakacha and Wescogame formations as well as Pakoon Limestone, but we have no complete study of their presumed interbedding with eolian sandstone. These relationships would be exposed best in the Shivwits Plateau region of the western Grand Canyon.

Documentation as to how far west eolian sandstone and arenaceous limestone extends also is nonexistent. Based on similarities to eolian strata that I have seen (photographs included in McKee 1982a), it seems likely that eolian strata exist in western Grand Canyon and northward along the Grand Wash Cliffs.

It is equally difficult to determine the depositional setting of the Hermit Formation. White (1929) used fossil flora, general lithology, and a regional setting to postulate deposition by sluggish streams on a broad, low-lying, arid coastal plain. Details of Hermit deposition in the Grand Canyon have yet to appear in the literature. Work currently in progress in the Sedona area is providing some detail of Hermit deposition for the western Mogollon Rim region. A brief examination of scattered Hermit outcrops in the Grand Canyon tends to confirm White's general conclusions.

This information, combined with more detailed work near Sedona, provides a few more specifics on the nature of Hermit streams. Several types of stream deposits have been identified. These include: narrow, channel-shaped bodies of limestone-pebble conglomerate that formed in local, arroyolike streams as well as broader sheets of conglomerate that probably formed in wider arroyos. The conglomerate was derived from caliche deposits on adjacent alluvial plains. Sheets of limy and nodular structureless-to-ripple-laminated sandstone were deposited as relatively unconfined stream deposits or as broad levee deposits. Trough, cross-stratified sandstone (with or without associated conglomerate) laid down in broad, gently dipping, irregular sheets (epsilon cross-stratification) represents the deposits of larger, more perennial, meandering streams. Red mudstone was deposited in low-lying areas as overbank deposits. Light-colored, large-scale, high-angle, cross-stratified sandstone (with climbing translatent strata) represents minor eolian dune deposits.

From the Sedona area westward to an area near Seligman, the types of deposits described above are distributed irregularly throughout the Hermit. In central and eastern Grand Canyon, geologists are able to identify all of these types. However, three general facies predominate.

The lower portion of the Hermit on both the Kaibab and Hermit trails consists of ledges and slopes of structureless and trough cross-stratified sandstone. Exposures there are not as good as some in the Sedona area, but the sand bodies are similar to the meandering stream deposits and unconfined stream deposits there. The bulk of the Hermit in eastern and central portions of the Grand Canyon consists of weak, ledge-forming, silty, faintly ripple-laminated sandstone and slope-forming mudstone. Generally, 10 to 15 cyclic alternations of these units are present. The ledges probably represent sluggish, shallow, perhaps unconfined stream deposits, and the slopes primarily are overbank deposits. Whether these cycles were formed by climatic fluctuation, by the migration of stream deposits, or by tectonic cycles is unknown.

Depositional Models

The high carbonate grain content, associated marine carbonates, and regional setting strongly suggest a coastal-plain setting for the eolian dune deposits in the Supai Group. Carbonate sand was derived locally by onshore winds from adjacent, bioclastic debris. The quartz sand was transported by wind and longshore currents into the Grand Canyon region from the north (Blakey et al. 1988). To the west was a broad, shallow, epicontinental sea, and to the east were low-lying coastal plains and low uplands (Fig. 9.13). The repetitive Pennsylvanian and Early Permian marine transgressions and regressions moved across this broad plain, but influxes of eolian sand formed obstacles to widespread transgression. During regression, the eolian dunes spread across the broad flats. The ensuing transgression interrupted eolian deposition and apparently prevented the development of large eolian sand seas (ergs).

The battle between dune and sea was repeated numerous times, and only a small fragment of the sedimentological record has been preserved. Sluggish and probably ephemeral streams continually washed fine-grained, terrigenous, and clastic debris into the area. The mud and fine-grained sand were trapped in lagoons, on tidal flats, and in river channels. The extremely arid climate provided a reflux action that pumped salty groundwater upward. This resulted in the precipitation of calcium carbonate and other minerals as deposits on the earth's surface. Some of these nodules were reworked later into conglomeretic deposits. Such a setting clearly was the site of a great deal of reworking of sediment.

Over time, the wind altered marine and fluvial deposits into dune deposits and eolian dust (loess). Rivers and advancing seas also reworked these eolian deposits. This intermixing and reworking of environments created an array of conditions that might seem incompatible within a given environment. For example, fluvial and tidal deposits might be better sorted than expected in one area because of the activity of the wind. Eolian deposits frequently display geometric patterns that are more typical of fluvial or shallow-marine deposits. The heterogeneity of much of the Supai is directly related to this complex intermixing of environments coupled with a broad range of available depositional material.

Hermit stream deposits do not fit readily into established fluvial models. The broad, low-gradient, arid plain was crossed by several kinds of streams, probably both perennial and ephemeral. Local coarse conglomerate indicates that

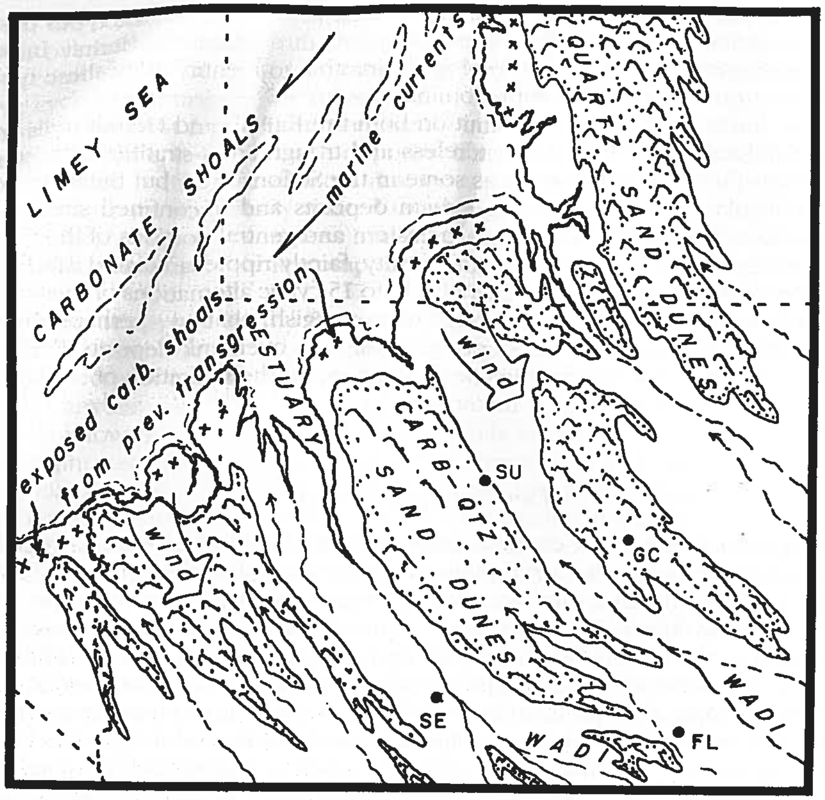

FIGURE 9.13. Map showing general depositional model and hypothetical paleogeography of eolian units in Supai Group. Map depicts region during regressive sequence. Carbonate sand is derived from exposed, older carbonate shoals; quartz sand is fed into the region from the north. Note state lines and abbreviations for Supai, Grand Canyon, Seligman, and Flagstaff.

streams occasionally had sufficient energy to transport larger clasts. Widespread and small dune fields were present on the floodplain, but the great eolian faucet from the north had been shut off temporarily.

STRATIGRAPHIC ANALYSIS

The incredible range of Supai depositional environments and the processes within them are only generally understood. Clearly, many problems remain and some may never be solved. Patterns of transgression and regression have yet to be documented. Hermit deposition is poorly understood, primarily because of a lack of regional and local data. Despite these problems, the ideas, models, and conclusions offered here permit a preliminary analysis of Supai and Hermit depositional history.

Morrowan History

The lower two-thirds of the Watahomigi formation was deposited in shallow-marine and low-energy shoreline environments (McKee 1982a). The Morrowan portion of the Watahomigi consists generally of a lower, fine-grained, siliciclastic sequence and an upper, fossil-bearing limestone. Poor exposures and a lack of reliable criteria prevent detailed analysis of the lower slope. A basal conglomerate suggests that continental or shoreline processes reworked weathered Redwall Limestone. It is possible that the overlying mudstone accumulated in low-energy, shoreline environments or on a broad coastal plain. A rise in sea level (or subsidence in the Grand Canyon region) permitted a period of widespread, clear-water, carbonate deposition. Local intercalated mudstone suggests that there were periods of influx by fine clastics. These may have been caused by fluctuating sea levels. A widespread erosion surface at or near the top of the limestone suggests a period of erosion near the close of Morrowan time (McKee 1982a, p. 160).

Atokan History

The upper Watahomigi Formation consists of thin limestone, redbeds, and minor sandstone. This suggests an advance of the sea into the Grand Canyon area. However, a strong influx of eolian material from the north initiated a trend of eolian deposition that would continue periodically for over 150 million years (Blakey et al. 1988). During a major regression, quartz and carbonate sand blew across the landscape. It was derived locally from exposed, recently deposited, shallow-marine materials. Quartz sand from the north was transported across the Paiute and Emery arches and was funneled into the Grand Canyon embayment (Fig. 9.13). Here, the sand was deposited in coastal dunefields. We are not certain whether the eolian sand was deposited in a broad coastal erg or in a series of small dunefields. However, eolian deposits were planed off by marine transgressions and perhaps minor fluvial events. The next regression saw a return to eolian deposition. Sharp bases to eolian sandstone units suggest a rapid change to eolian conditions.

DesMoinesian and Missourian History

No rocks of these ages have been reported from the Grand Canyon or the western Mogollon Rim regions. This period of time apparently is represented by the unconformity between the Manakacha and Wescogame formations. Whether sediments of Middle Pennsylvanian age were deposited and later removed or were never deposited extensively in the region is unknown.

Virgilian History

The pattern of deposition established during Atokan time was renewed during deposition of the Wescogame Formation. Repeated transgression and regression caused numerous depositional cycles. The widespread Wescogame cliff unit, believed to be mostly eolian in origin, may represent the development of a large erg, or of several ergs, across the region (Fig. 9.13). Redbeds at the top of the Wescogame Formation may have formed during a period of decreased eolian influx.

Another period of erosion followed. This is indicated by the unconformity at the presumed Pennsylvania–Permian boundary. Apparently, a lowering of the sea level caused an incision of arroyolike streams into the underlying Wescogame Formation.

Wolfcampian History

During this period, the sea advanced into the region at least as far east as the central Grand Canyon. The Pakoon Limestone formed in a variety of clear-water, shallow-marine environments. Farther east, coastal-plain and minor eolian deposits are represented in the lower slope unit of the Esplanade Sandstone. Later in the Wolfcampian, a vast blanket of eolian sand spread southward across the Colorado Plateau region. Forming the Cedar Mesa Sandstone in southeastern Utah, the Queantoweap Sandstone in southwestern Utah, and the Esplanade Sandstone in northern Arizona, eolian deposits of Wolfcampian age blanketed much of the Southwest (Blakey et al. 1988).

Interruptions in eolian deposition in the Esplanade Sandstone may have been caused by a rise in sea level, influxes of fluvial activity, or both. East of the Sedona arch, eolian deposits became sparse to absent, and a period of fluvial deposition dominated the landscape.

As eolian conditions waned, fluvial deposits spread westward into the Grand Canyon and the western Mogollon Rim region. Some fluvial channels were incised into the underlying eolian sandstone, but in many locations a general transition to fluvial conditions is evident.

Leonardian History

Most likely beginning in the Late Wolfcampian and continuing into the Leonardian, low-energy, fluvial conditions developed. Based on the distribution of the Hermit and coeval deposits in Utah, Colorado, New Mexico, and eastward into the High Plains, geologists have concluded that this marked a period of extremely widespread fluvial deposition. Deposits farther east were left behind by streams that had steeper gradients and carried coarser debris. As the streams crossed the broad, low-lying coastal plain, they carried fine material and locally derived carbonate clasts. A few scattered dune fields persisted into this period of time, but in general the great eolian episode was brought to a temporary halt.

Eolian deposition was renewed in the Leonardian, first with the development of eolian deposits in eastern and central Arizona in the De Chelly Sandstone and the Schnebly Hill Formation and then across most of the northern and central Arizona, including the Grand Canyon region in the Coconino Sandstone.

SUMMARY

Some of the conclusions concerning the depositional history of the Supai Group and Hermit formation are preliminary and probably will alter with further work. However, one aspect seems clear: Eolian depositional processes played an important role in the formation of the Supai Group and parts of the Hermit formation. The Atokan eolian deposits in the Manakacha formation may represent the very earliest record of the vast eolian systems that were to dominate the Southwest until the Late Jurassic.

• 10 •

Coconino Sandstone

Larry T. Middleton, David K. Elliott, and Michael Morales

INTRODUCTION

The Early Permian Coconino Sandstone is one of the moist conspicuous formations in the Grand Canyon. The high-angle, sweeping sets of cross-stratification, which record the southerly advance of very large dunes, are visible at great distances throughout the area. Ironically, few geologists, except for McKee (1933b) and Reiche (1938), have studied the Coconino. These Sahara-like dunes were part of an enormous desert that once extended north into Montana.

During the last 10 years, a number of studies on modern and ancient sand seas, or ergs, have increased greatly our understanding of eolian bedform dynamics. At the same time, these studies have facilitated the development of facies models for eolian depositional systems. It is possible now to apply the results of these studies of stratification styles to the Coconino Sandstone and to discuss possible dune morphologies and migration paths. Additionally, we can describe trace fossils in the Coconino and relate them to the physical environments of deposition. This provides us with a more comprehensive reconstruction of the erg environments (e.g., substrate conditions). Although the cliff-forming nature of the formation can present certain logistical problems, the exceptional three-dimensional exposures in the Grand Canyon afford the sedimentologist and ichnologist the opportunity to examine the details of one of the more classic and interesting eolian deposits on the Colorado Plateau.

Because the results reported here are preliminary, it is our hope that they will spur more detailed studies of lateral and vertical facies relationships in the Coconino Sandstone, as well as initiate more effort at documenting the variety and distribution of the trace fossil assemblages.

REGIONAL STRATIGRAPHIC RELATIONSHIPS

The Middle Leonardian Coconino Sandstone crops out throughout much of the southern Colorado Plateau south of the Utah–Arizona border. The most exceptional exposures are in the Grand Canyon and the Marble Canyon areas and along the Mogollon Rim south and southeast of Flagstaff. Darton (1910) named the formation for exposures in Coconino County, Arizona. Although

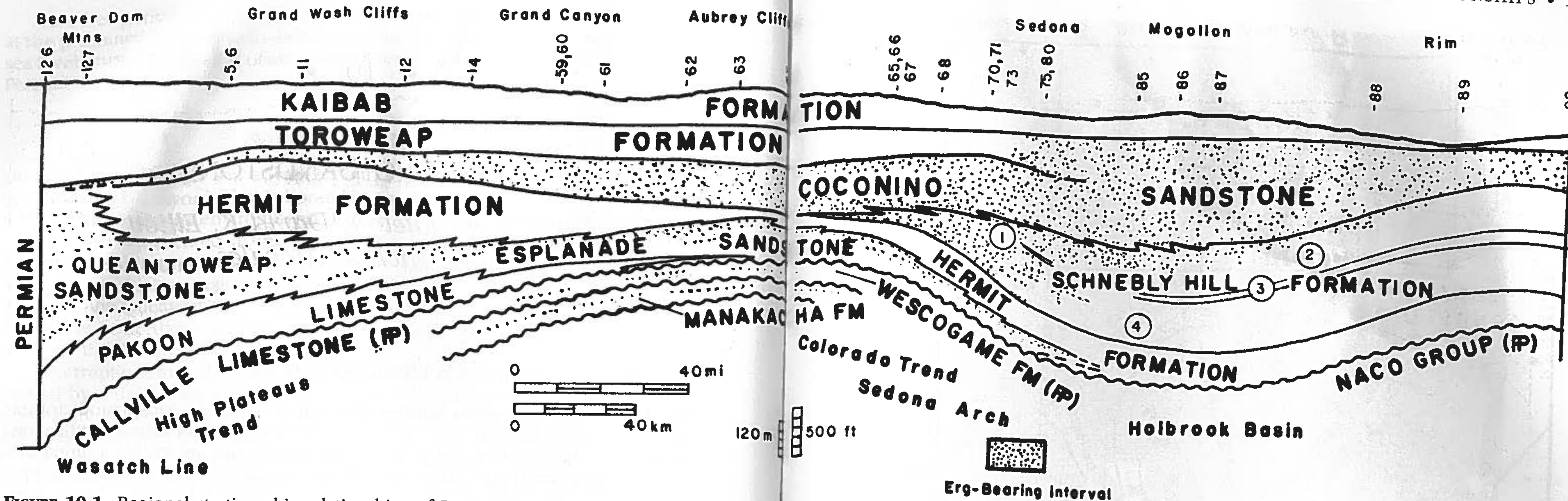

FIGURE 10.1. Regional stratigraphic relationships of Permian strata along the southern margin of the Colorado Plateau. (After Blakey and Knepp, 1988.)

there have been a number of controversies regarding Permian stratigraphy in northern Arizona, recent studies have established some regional stratigraphic relationships (Fig. 10.1) (Cheevers and Rawson 1979; Rawson and Turner-Peterson, 1980; Blakey and Middleton, 1983; Blakey and Knepp 1988; Blakey 1996).

In the Grand Canyon region, the Coconino Sandstone disconformably overlies the Permian (Wolfcampian–Leonardian) Hermit Formation. Throughout the canyon, the Coconino is overlain by, or intertongues with, the Permian Toroweap Formation. In easternmost localities where the Toroweap is not present, the Kaibab Formation conformably overlies the Coconino. The Coconino Sandstone grades eastward into the Glorieta Sandstone in western New Mexico. To the south, along the Mogollon Rim, it intertongues with the Schnebly Hill Formation and is transitional into the overlying Toroweap Formation or is sharply overlain by the Kaibab Formation.

Regional structural features (Fig. 10.2) control the marked thickness variations in the Coconino Sandstone. In the western part of the Grand Canyon near the Grand Wash Cliffs, the formation is 65 feet (20 m) thick, but it thins progressively to the west. The Coconino thickens in the central part of the canyon and is over 600 feet (183 m) thick near Cottonwood Creek. Eastward and northward, the Coconino thins to 57 feet (17 m) in Marble Canyon before pinching out in the vicinity of Monument Valley. The Coconino, likewise, wedges out in southwestern Utah.

McKee (1933b, 1974) reported a maximum thickness of nearly 1000 feet (305 m) near Pine, Arizona, along the Mogollon Rim (Fig. 10.2). This rapid southerly thickening probably is related to increased subsidence along the southern boundary of the Coconino depositional basin. The lithologies, textures, and sedimentary structures in the Toroweap Formation are similar to those of the Coconino toward the south. Hence, some of the thickening might reflect the presence of Toroweap equivalents in the upper part of the Coconino.

Blakey and Knepp (1988) proposed that the Sedona arch (Fig. 10.2) controlled facies distributions, as well as thickness variations, in several of the late Paleozoic units in northern Arizona. This seems likely because the Coconino thickens across and to the east of the Sedona arch, and it is in this area that the Toroweap Formation undergoes facies changes into the Coconino Sandstone.

ICHNOLOGY

Introduction and Preservation

Because body fossils have not been reported from the Coconino Sandstone, invertebrate and vertebrate trace fossils (Fig. 10.3) remain the only evidence of the organisms that lived in the Coconino desert. They are also the only data on which to base the studies of fossil diversity, morphology, and paleoecology. In particular, trace fossils have been important evidence in the debate on the depositional environment of the Coconino Sandstone. Work by Brady (1939, 1947), McKee (1944, 1947), and Brand (1979) has shed a considerable amount of light on the formation of the traces though differences in interpretation still remain.

Brady (1939, 1947) carried out a series of experiments with small vertebrates and invertebrates and pointed out similarities between the fossil traces and those formed by modern scorpions, millipedes, and isopods. In addition, he found that in wet or slightly moist sand, no trace was left by the scorpions he was working with but that they made clear impressions in dry sand. As will be discussed, work by Sadler (1993) has expanded on these studies.

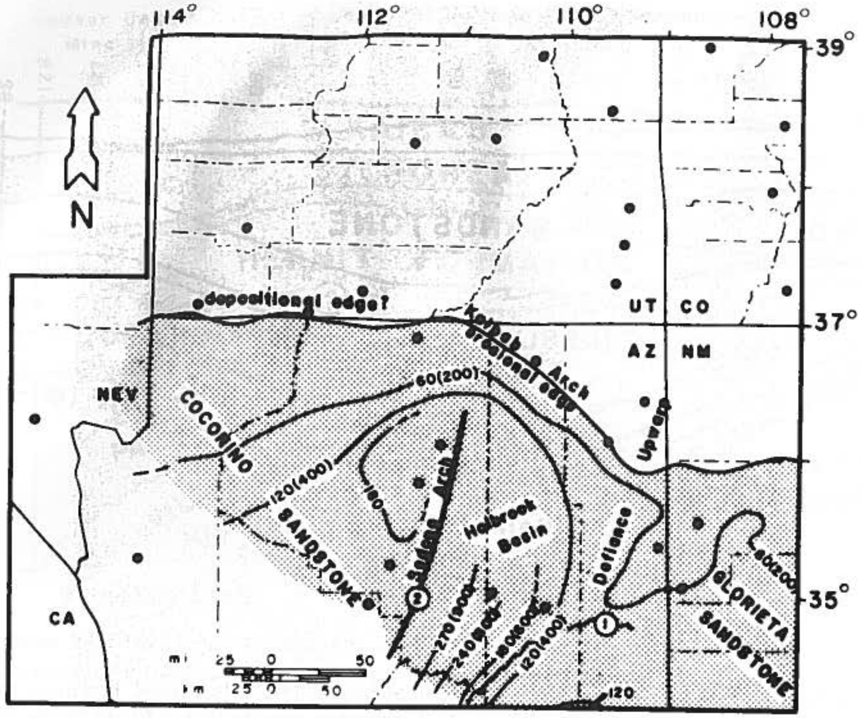

FIGURE 10.2. Isopach map of the Coconino Sandstone illustrating thinning away from the Sedona arch. (After Blakey and Knepp, 1988.)

In the 1940s, McKee conducted a series of experiments designed to duplicate the Coconino tracks. He filled a long trough with sand that formed a hill in the center. A variety of small vertebrates and invertebrates then were induced to walk along the sand and over the hill. By varying the slope of the hill and the moisture content of the sand, McKee was able to test the trace-forming capabilities of a variety of animals in a number of different environments. He tested the animals on a variety of surfaces—including dry sand, damp sand, saturated sand, and even sand that had been soaked and then allowed to dry. He found that only the largest animals tested (chuckwalla lizards) were able to make tracks in wet sand or crusted sand, and even then, the tracks were not as clear as those formed in dry sand. Smaller animals, such as millipedes and scorpions, were unable to make tracks in wet sediment and left clear traces only in dry, loose sand. He also found that at slopes below 27°, both uphill and downhill tracks were likely to be preserved. Avalanching, however, tended to destroy tracks on steep slopes.

Based on previous studies of conditions necessary to preserve surface features in dune sands, McKee (1945) suggested that the tracks in the Coconino Sandstone were formed initially in loose, dry sand that was dampened subsequently before being covered. The mists and fogs that intermittently are present in areas of coastal sand dunes would provide a suitable means of dampening the surface. Recent work by Brand (1978, 1979), however, suggests that clear traces with morphologies similar to those found in the Coconino Sandstone can be generated on sandy surfaces submerged in standing water. It may be, there-

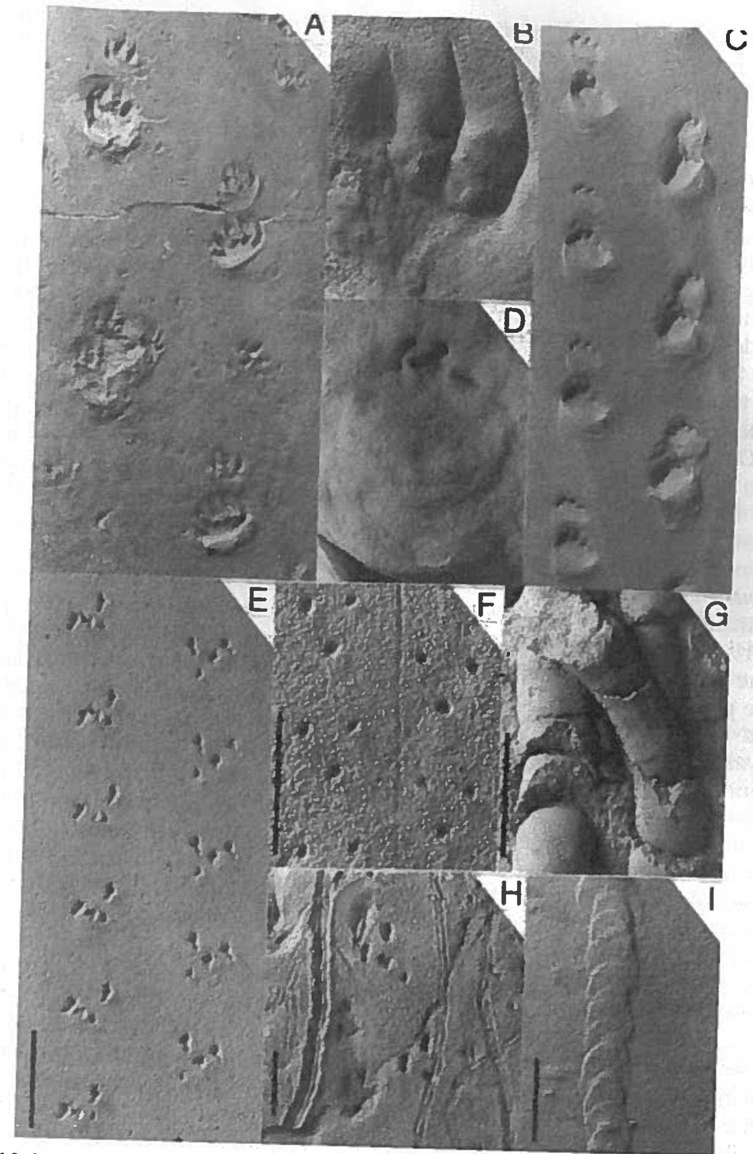

FIGURE 10.3. Vertebrate and invertebrate trace fossils from the Coconino Sandstone. (All specimens numbers refer to the Museum of Northern Arizona, Flagstaff, Geology Collection.) (a) *Laoporus* sp. (V3470); notice the four toes on the forefeet prints and the five toes on the hindfeet prints. (b) *Baryopus tridactylus* (V3360). (c) *Laoporus* sp. (V3472); notice the sand humps behind the tracks, indicating that the animanl walked uphill, pushing back loose dune sand. (d) *Baropus* sp. (V 3388). (e) Spider track (N3089). (f) *Paleohelcura dunbari* (N3666). (g) *Scolecocoprus cameronensis* (N3707). (h) *Diplodichnus biformis* (N3657). (i) *Scolecocoprus arizonae* (N3655).

fore, that a study of trace morphology alone is not sufficient to distinguish between sand surfaces exposed to subaqueous or subaerial conditions. Further experimentation may help to resolve this problem.

Invertebrate Trace Fossils

In 1918, Lull described trace fossils from the Coconino Sandstone that he collected near the Hermit Trail in the Grand Canyon (Lull 1918). Subsequently, C.W. Gilmore of the U.S. National Museum carried out more detailed work in the same area. Gilmore published his results in a series of papers (Gilmore 1926, 1927, 1928), describing both vertebrate and invertebrate tracks. In these studies, he described 10 genera and 17 species, of which the following five genera and species clearly were invertebrate: *Mesichnium benjamini*, a trail consisting of two parallel lines of footprints with a median row of suboval, regularly spaced depressions; *Octopodichnus didactylus*, a trail consisting of alternating sets of impressions in groups of four; *Paleohelcura tridactyla*, a trail consisting of alternating sets of three prints with a median tail drag; and *Triavestiga niningeri*, a trail composed of paired, longitudinal grooves. Gilmore ventured little opinion as to the identity of the trail-formers beyond suggesting that *Octopodichnus* and *Paleohelcura* showed some similarity to tracks made by modern crustaceans.

Through many years, Brady continued work on these tracks (Brady 1939, 1947, 1949, 1955, 1959, 1961). He described new traces, provided new insights into the conditions necessary for their formation, and identified the trace-formers. He pointed out (1939, 1947) that the *Paleohelcura* trails were very like those formed by the modern scorpion (*Centruroides*) in dry sand when the temperature was about 60°F (15°C). He also showed, however, that with variations in temperature and surface conditions, the same animal could leave a variety of traces—sometimes impressing two, three, or four feet on each side and leaving no tail drag, an intermittent one, or a complete impression of the tail. Based on this information, he showed that a variety of trails in the Coconino Sandstone were simply variations on *Paleohelcura*. It is clear from this work that two of the ichnogenera described by Gilmore (1926), *Triavestiga* and *Mesichnium*, represent variations on *Paleohelcura* and do not merit the status of separate ichnogenera. In addition, Brady recognized one new ichnospecies of *Octopodichnus*, *O. minor*.

Brady (1947) was able to show that some two- and three-grooved trails were very similar to those formed by modern millipedes as they move up and down sand slopes. He named these *Diplopodichnus biformis* (Fig 10.3H) and pointed out that the traces attributed by Gilmore (1926) to *Unisulcus* were identified incorrectly. The traces attributed to *Unisulcus* are slender, smooth, and wormlike and should be included in *Diplopodichnus*. Trails similar to those formed by the modern isopod *Oniscus* he named *Isopodichnus filiciformis*. However, he discovered subsequently that the name was already in use, and he renamed the ichnogenus *Oniscoidichnus* (Brady 1961). Traces that Brady attributed to oligochaete fecal pellets he named *Scolecocoprus cameronensis* and *S. arizonensis* (Fig. 10.3g and 10.3i). He described them as consisting of an overlapping series of pellets and attributed the sometimes grooved appearance of the lower surface to traces left by spines on the ventral surface of an oligochaete. It is more probable that these represent the infillings of burrows, the spreite representing what was thought to be the rounded ends of the fecal pellets. In 1961, Brady also described an additional species of *Paleohelcura*, *P. dunbari* (Fig. 10.3f).

Table 10.1. Invertebrate Ichnospecies from the Coconino Sandstone

Diplopodichnus biformis (Brady 1947)
Mesichnium benjamini (Gilmore 1926)
Octopodichnus didactylus (Gilmore 1926)
O. minor (Brady 1947)
Oniscoidichnus (Isopodichnus) filiciformis [Brady 1949 (1947)]
Paleohelcura tridactyla (Gilmore 1926)
P. dunbari (Brady 1961)
Scolecocoprus cameronensis (Brady 1947)
S. arizonensis (Brady 1947)
Triavestiga niningeri (Gilmore 1927)

Alf (1968) carried out the most recent work on invertebrate traces from the Coconino Sandstone, describing a trail composed of alternating sets of four prints (Fig. 10.3e). These clearly were different from *Octopodichnus*, which consists also of alternating sets of four prints, and Alf was able to show from experiments with modern tarantulas and wolf and trapdoor spiders that the most likely trace-former was a spider. He did not name the trace, however, though clearly it is a separate ichnogenus. The Coconino Sandstone's invertebrate trace fossils are listed in Table 10.1.

Traces similar to those reported from the Coconino Sandstone have been reported from other late Paleozoic and Mesozoic eolian sequences (Ekdale and Picard 1985; Sadler 1993). In a study of trace fossils from the Permian De Chelly Sandstone in northeastern Arizona, Sadler (1993) examined tracks similar to those in the Coconino Sandstone and conducted experiments using modern arthropods as a means of identifying the possible track makers. The De Chelly Sandstone is correlative to the lower part of the Coconino Sandstone, and like the Coconino is eolian in origin (Blakey and Knepp 1989). Sadler described four ichnospecies assigned to the ichnogenera *Paleohelcura* and *Octopodichnus* based on comparisons with similar traces in the Coconino Sandstone in Grand Canyon. Sadler's study involved experimental studies using scorpions and spiders to demonstrate that not only did these arachnids produce tracks similar to those found in both formations but that the trackways provided evidence of direction of movement and information concerning substrate moisture conditions.

Sense of movement in these two ichnogenera is indicated by a bifurcation of impressions that open in the direction of travel. In Figure 10.3 (f and h) the direction of movement is toward the top of the page. Sadler's study also showed that determination of true sense of movement requires more detailed and more quantitative analyses of the tracks because several trails exhibited a backward orientation to the bifurcation angle.

The consistency of the substrate as a requirement for preservation of tracks has long been a question. Early work by McKee (1947) indicated that preservation only was likely in dry cohesionless sand deposited on gentle slopes or horizontally. Sadler's work supported McKee's contention that preservation potential was greatest for tracks going up dune slip faces due to the problems of track obliteration by grain flows during down-slope movement of animals. Sadler (1993) demonstrated that track preservation was likely under these conditions but also showed that a damp sand surface also was conducive to preservation of tracks made by scorpions and spiders and that increased moisture content of dune sands can result in preservation of down-slope-directed tracks. Preservation of tracks made by arachnids in sands that are damp necessitates subsequent infilling by sand deposited by grainfall or wind-ripple migration.

Desert sands become damp and therefore somewhat cohesive for a variety of reasons. Moisture content of desert sands is of course low because most deserts typically receive less than 150 mm of mean annual rainfall. Modern settings such as the Namib Desert in western Africa have varying moisture contents as a function of their proximity to onshore coastal winds, to basin margin and basin-central fluvial systems, and also to the magnitude of water table rise during wetter seasons (Lancaster 1984, 1989, 1995). Studies by Hasiotis and Bown (1992) of coastal dune systems have shown that the abundance of insects and arachnids increases landward due to an increase in vegetative diversity and substrate stability.

As has been shown in several studies (Briggs et al. 1984; Sadler 1993), a single trackway can exhibit a myriad of track morphologies obviously generated by the same animal. More experimental studies utilizing insects and arachnids as well as more careful morphometric analyses of fossil trackways is clearly warranted and needed if definitive identification of the track maker is to be attained.

Vertebrate Trace Fossils

Lull (1918) produced the first description of tracks attributed to tetrapod vertebrates in the Coconino Sandstone. The tracks consist of forefoot prints with four toes and hindfoot impressions with five toes. Both of these tracks are relatively small, 0.5–1.5 inches (1.5–3.0 cm) in length, and show claw marks. Lull assigned the tracks to two ichnospecies, *Laoporus schucherti* and *L. noblei*. The rest of the vertebrate trace fossils found in the Coconino Sandstone were described by Gilmore in the late 1920s. On the basis of tracks similar to those described by Lull (though different in detail), Gilmore named the following ichnotaxa: *Agostopus matheri* and *A. medius*; *Allopus? arizonae*; *Amblyopus pachypodus*; *Baropezia eakini*; *Baropus coconinoensis* (Fig. 10.3d); *Barypodus palmatus*, *B. metszeri*, and *B. tridactylus* (Fig. 10.3b); *Dolichopodus tetradactylus*; *Nanopus merriami* and *N. maximus*; and *Palaeopus regularis*. Gilmore (1926) also redescribed *Linopus? coloradensis* and referred it to the ichnogenus *Laoporus*.

Baird (1952 and in Spamer 1984) and Haubold (1984) reevaluated the vertebrate tracks from the Coconino Sandstone. Baird (1952) decided that *Allopus? arizonae* and *Baropus coconinensis* are junior synonyms of *Baropezia eakini*. Later, he transferred *Nanopus maximus* to *Barypodus metszeri* and attributed the ichnospecies *N. merriami* and *Dolichopodus tetradactylus* to the ichnogenus *Laoporus*. However, Haubold (1984) maintained the generic distinction of *Dolichopodus*. He also had doubts about the taxonomic validity of *Amblyopus pachypodus* and considered *Agostopus*, *Barypodus*, *Nanopus*, and *Palaeopus* to be junior synonyms of *Laoporus* (Fig. 10.3a and 10.3c). This view may be a case of excessive taxonomic lumping, however, because Baird (1952 and in Spamer 1984) has not indicated such sweeping synonymies. To resolve this and related problems, a thorough taxonomic revision of the Coconino Sandstone's vertebrate tracks is needed. A list of all the original names given to vertebrate trace fossils from the Coconino Sandstone is given in Table 10.2.

Table 10.2. Original Names of Vertebrate Ichnospecies from the Coconino Sandstone

Agostopus:	*A. matheri* (Gilmore 1926)
	A. medius (Gilmore 1927)
Allopus?:	*A.? arizonae* (Gilmore 1926)
Amblyopus:	*A. pachypodus* (Gilmore 1926)
Baropezia	*B. eakini* (Gilmore 1926)
Baropus:	*B. coconinoensis* (Gilmore 1927)
Barypodus:	*B. palmatus* (Gilmore 1927)
	B. metszeri (Gilmore 1927)
	B. tridactylus (Gilmore 1927)
Dolichopodus:	*D. tetradactylus* (Gilmore 1926)
Laoporus:	*L. schucherti* (Lull 1918)
	L. coloradensis (Gilmore 1926)
	L. noblei (Lill 1918)
Nanopus:	*N. maximus* (Gilmore 1927)
	N. merriami (Gilmore 1929)
Palaeopus:	*P. regularis* (Gilmore 1926)

EOLIAN DEPOSITIONAL SYSTEMS

Although we have known about the eolian origin of the Coconino Sandstone for a long time, there are comparatively few data on the details of dune types and distribution. Nor is there documentation of small-scale stratification features within the larger co-sets of cross-stratification that characterize the formation. Within the last decade, a number of studies have examined ancient and modern eolian deposits. Geologists have developed criteria and facies models by which we can interpret these sediments. In the process, we have learned more about depositional processes and the geomorphic features of eolian sand seas, or ergs. When this information and these models are applied to the Coconino Sandstone, a better understanding of the genesis and evolution of the Coconino erg will emerge.

The Coconino Sandstone is composed of fine-grained, well-sorted, and rounded quartz grains and minor amounts of potassium feldspar. The cement is primarily silica in the form of quartz overgrowths. These textural and mineralogic characteristics are compatible with an eolian environment in which sediment transport involves numerous grain-to-grain collisions. These collisions result in the mechanical destruction of less stable grains and in winnowing by the wind. As McKee (1979) correctly pointed out, however, these characteristics do not substantiate conclusively a wind-blown origin. Although paleocurrent trends suggest a northern source for this sand, we cannot identify the source(s) of such a large quantity of quartz in the Coconino, as well as in correlative units to the north such as the Weber Sandstone in Utah and the Tensleep Sandstone in Wyoming and Montana.

Primary sedimentary structures in the Coconino Sandstone include small- to large-scale [up to 66 feet (20 m) thick] planar-tabular and planar-wedge cross-stratification (Fig. 10.4), compound cross-stratification, horizontal stratification, ripple marks (Fig. 10.5), and raindrop impressions (Fig. 10.6). Deformation features also occur and consist of small, pull-apart structures and a variety of slump-related features (Fig. 10.7). Small-scale stratification comprises wind-ripple lami-

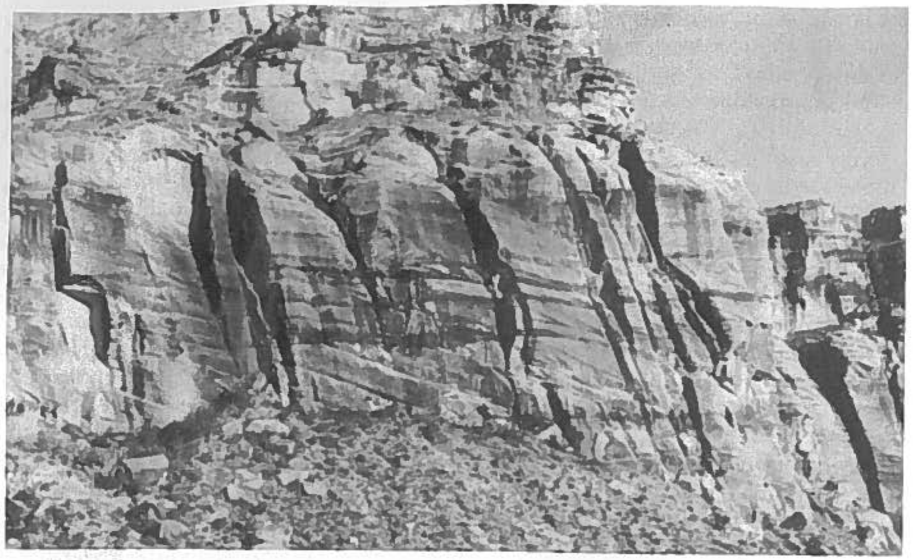

FIGURE 10.4. Large-scale planar-tabular cross-stratification in the Coconino Sandstone in Hualapai Canyon. Contact with overlying Toroweap Formation is at slope above the large-scale cross-strata.

FIGURE 10.5. Ripple marks striking down large foreset of cross-strata. This orientation of ripple crests is common throughout the Coconino Sandstone and indicates wind flow and sediment transport across the surfaces of the larger bedforms. Lens cap for scale.

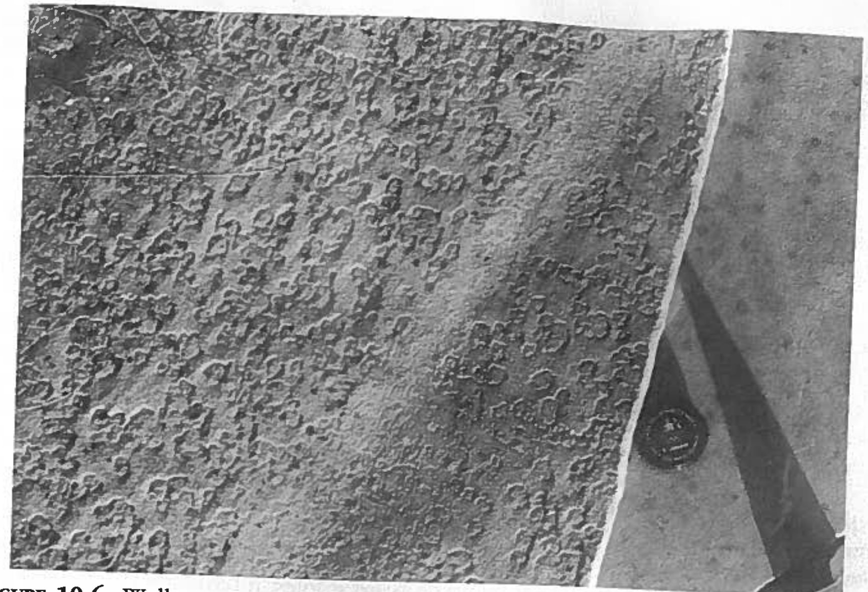

FIGURE 10.6. Well-preserved raindrop impressions on foreset on large set of cross-strata. The irregular margins of the circular pits are somewhat raised on the down dip side. Lens cap for scale.

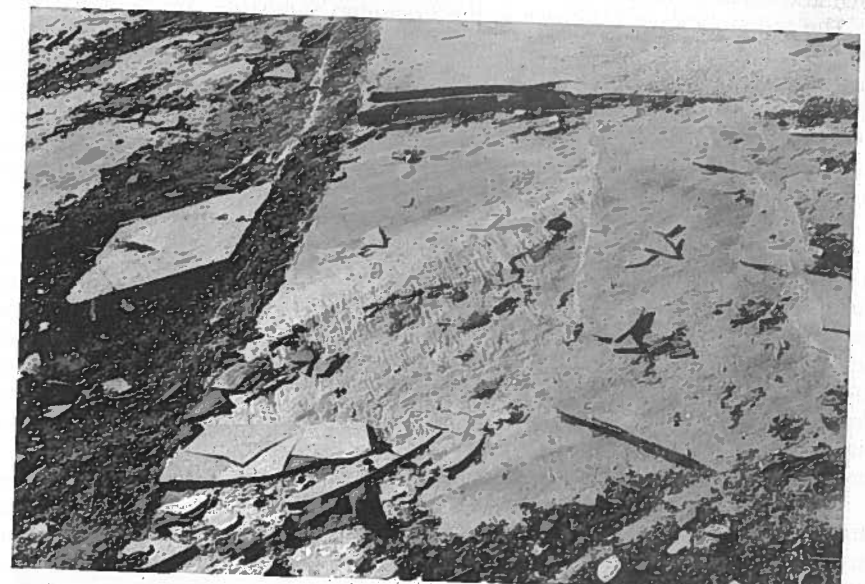

FIGURE 10.7. Pull-apart (detachment structures on slipface of gently dipping foresets. These features probably represent small-scale avalanching of semicohesive sand. Lens cap for scale.

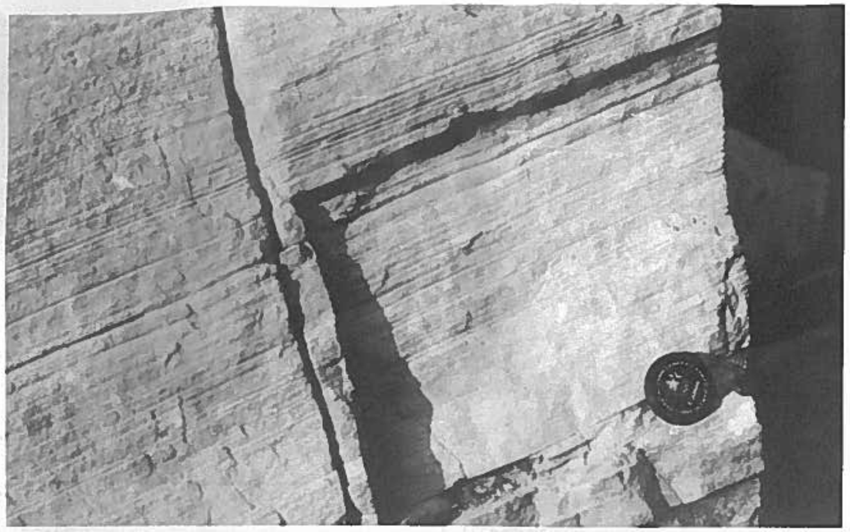

FIGURE 10.8. Thin, evenly continuous laminae that compose the bulk of the foreset deposits. These laminae represent climbing wind ripples.

nations and sandflow strata and minor grainfall laminae. Collectively, these features, together with facies geometries, support an eolian interpretation and can be used to characterize bedform morphologies as well as variations in depositional and substrate conditions.

The most obvious structures in the Coconino Sandstone are the thick sets of cross-stratified sandstone (Fig. 10.4). Internally, these strata exhibit thin (less than 1 cm thick) laminae that are continuous for considerable distances along the foreset of the cross-stratification (Fig. 10.8). These laminae conform to the shape of the foresets: planar where the foresets are straight and becoming concave upward on the tangential foresets. In most cases, these laminae contain no internal structures. Grain size, though quite uniform within the laminae, exhibits a slight inverse grading.

Recent studies, most notably that of Hunter (1977), demonstrate that these structures are the products of wind-ripple migration as they climb over the upwind and downwind sides of dunes and in interdune areas. For the most part, the wind ripples in the Coconino Sandstone moved transversely across the lee side of the dunes. This cross-dip path of migration is caused by secondary air currents that parallel the strike of the lee side of simple and complex dunes (or *draas*). The low height-to-wavelength ratio of the wind ripples as measured in plan view exposures of many foresets is consistent with those recorded from modern coastal and inland dunes (McKee 1979).

Thin, down-foreset, tapering laminae occur in many of the thicker cross-stratified sets. These deposits rarely exceed 3 inches (7.5 cm) in thickness and lack any noticeable grain-size grading and internal stratification. Wind-ripple laminae typically surround these wedge-shaped units. McKee et al. (1971) and Hunter (1977) have shown that these features form by the avalanching of loose sand on dune slipfaces. These "sandflow" strata form either by a loss of grain cohesion in a descending avalanche sheet or by scarp recession (Hunter 1977). In the latter case, sand moves down and away from the detachment area as the scarp it-

self migrates up and across the slipface. Whichever the mechanism, these strata represent the avalanching of loose, dry sand.

Other "detachment structures" in the Coconino Sandstone, however, suggest wetter substrate conditions. McKee et al. (1971) demonstrated that detachment and down-slope movement of moistened, somewhat cohesive sand produced steep-sided, pull-apart structures. Possible causes for the cohesiveness of the sand are difficult to verify, but wetting by rain and/or dew or early cementation are obvious possibilities. Our studies to date have demonstrated that cementation patterns generally are uniform throughout these strata, and it seems likely, therefore, that rainfall or dew caused the increase in substrate cohesiveness.

We can gain a great deal of information concerning the types of eolian bedforms in the Coconino erg by analyzing larger-scale features such as the thick sets and co-sets of cross-stratification and the truncation surfaces that bound or are contained within these sets. The Coconino Sandstone contains both simple sets of cross-strata and also complexly cross-stratified packages. Each is related to differences in dune types and/or flow processes associated with dune migration.

The most common types of cross-stratification in the Coconino Sandstone are planar-tabular and planar-wedge sets (Fig. 10.4). These consist of foresets with an average dip of 25 degrees and have a sharp or tangential basal contacts. McKee (1979) reported foreset lengths of up to 80 feet (24 m), but most are less than 40 feet (12 m) long. In general, the foresets do not contain any smaller cross-beds and are composed primarily of wind-ripple and subordinate sandflow laminae. Paleocurrent trends reported by Reiche (1938) and McKee (1979) are to the southeast and southwest. Our studies corroborate these earlier findings of a unimodal path of dune migration.

FIGURE 10.9. Stacked sets of planar tabular cross stratification separated by horizontal to gently dipping first order bounding surfaces. Concave-upward scouring of dune surface. Hualapai Trail, western Grand Canyon.

In most cases, these simple cross-bedded sets are underlain and overlain by horizontal to gently dipping planar surfaces (Figs. 10.4 and 10.9). These surfaces typically can be traced hundreds of meters along the outcrop. In all cases, these surfaces truncate the foresets of these thick, cross-bedded sets. Thin beds of horizontally laminated, fine-grained sandstone and coarse-grained siltstone overlie these surfaces in some localities. Typically, however, the surfaces separate large-scale cross-beds.

Although the origin of these features in ancient sandstone is controversial, the fact that planar surfaces truncate the cross-strata argues strongly that these are erosional features. Two major hypotheses put forth to explain the genesis of these laterally persistent surfaces involve the role of groundwater in desert basins and the phenomenon of migrating and climbing bedforms.

Stokes (1968) proposed that these extensive planes were the direct result of groundwater rise in dunes and subsequent wind erosion. According to this hypothesis, deflation of the dry, cohesionless sand above the saturation horizon resulted in the generation of a nearly planar surface. Subsequent migration of other dunes over these areas caused a vertical juxtaposition of cross-strata, such as that which occurs in the Coconino Sandstone and numerous other eolian sandstones on the Colorado Plateau. The net result of a repetition of this process during periods of basin subsidence were stacked sets of thick cross-strata separated by multiple truncation planes.

A major problem with the groundwater theory as applied to the Coconino Sandstone involves the nature of the water surface and the extensiveness of these surfaces. McKee and Moiola (1975) have shown that rather than being a horizontal surface, the upper level of the saturation horizon commonly is irregular and, in many instances, mimics the topography of the bedform. Deflation of the overlying dry sand in these cases would not result in a horizontal surface. It also seems rather improbable that this sand would cover a large enough area to create an extensive planation surface, notwithstanding the presence of interdune lows.

A more plausible hypothesis for the genesis of these first-order bounding surfaces in the Coconino Sandstone involves the downwind migration and climb of interdunal areas and dunes. Brookfield (1977) referred to these as first-order bounding surfaces and attributed their origin to the passage of large, complex eolian bedforms or draas. The planation surface most likely is caused by the truncation of downwind dunes or draas by migrating interdune areas. Interdune areas are zones of deflation that bevel the next downwind dune as the sand migrates downwind. Inherent in this model is the requirement that there is a net downwind climb to the bedforms. Rubin and Hunter (1982) have demonstrated the viability of bedform climb as a mechanism of generating these features. In a regional study of the Jurassic Entrada Sandstone, Kocurek (1981) proved that first-order surfaces can develop in this manner.

Another possibility that may explain the formation of these widespread surfaces involves climatic fluctuations. Talbot (1985) argues that basin-wide climatic changes can result in generation of these regional surfaces based on studies of geomorphic changes that have occurred along the southern margins of the Sahara Desert in northern Africa. In particular, this hypothesis suggests that periods of climatic shifts from arid to more humid periods during the Quaternary resulted in generation of broad areas where dunes were stabilized and eolian processes were relatively ineffective. In this model, stabilization of dunes would occur, erosion would ensue, and ultimately featureless areas would be common. Talbot documented this phenomenon in the southern Sahara where widespread areas are essentially devoid of large-scale dunes and planation surfaces are extensive. If preserved, these surfaces would represent regional bounding surfaces and are clearly indicative of widespread erosion.

There is scant evidence of fluvial and/or marine influences on development of the Coconino sand sea. These interactions would be expected around the margins of sand seas where, for example, rivers influence dune field buildup and migration (Middleton and Blakey, 1983; Langford, 1989; Langford and Chan, 1989; Herries, 1993) and strandline migrations modify coastal dune complexes (Chan and Kocurek, 1988). This is explicable if most of the Coconino Sandstone records deposition in interior ergs far removed from fluvial and coastal systems. There is no evidence of extradunal sand sheet deposition (Kocurek and Nielson, 1986).

The paucity of interdune deposits in the Coconino Sandstone is puzzling. Interdunes are common geomorphic elements of sand seas. A possible reason for their scarcity in the Coconino is that climatic fluctuations resulted in prolonged periods of stasis when bedform stabilization was protracted and interdune buildup was minimal. Also, it is possible that during bedform migration and climb, little deposition occurred in the interdunal corridors. Thus, the only record of interdunes would be the regional bounding surfaces.

Paleocurrent readings are strongly unimodal toward the south. This, coupled with the fact that the sets are internally simple, suggests that the bedforms that produced these cross-strata probably were relatively straight crested and that they were oriented transverse to the prevailing winds. McKee (1979) reported that in some localities the "curved shapes of small barchan types are recognizable." The majority of the dunes of the Coconino erg appear to have been of these two morphologies.

Although the aforementioned sets are characteristic of the Coconino, in many places the formation exhibits a complex system of bounding surfaces (Fig. 10.9) and extremely complicated sequences of internal erosional surfaces and stratification styles (Fig. 10.10). The concave-upward, bounding surfaces (Fig. 10.9) probably formed as the result of erosion—either by wind and the subsequent infilling of the eroded surface or by scouring associated with the migration of the next upwind dune–interdune couplet. Rubin and Hunter (1983) and Blakey et al. (1983) have documented the occurrence of these scoured surfaces in the late Paleozoic and Mesozoic sandstones on the Colorado Plateau.

Figure 10.10 illustrates some of the more complex geometries in the Coconino Sandstone. Low-angle surfaces truncate higher-angle sets of cross-stratification through this large set. These surfaces, which dip in the downwind direction, are over- and underlain by smaller sets. Brookfield (1977) has termed these low-angle, dipping surfaces second-order bounding surfaces "because they are truncated by the more extensive first-order surfaces."

These complex sets represent the deposits of large-scale, eolian draas (Wilson 1972a, b). Draas, which are common in most modern ergs, are characterized by smaller dunes migrating on both the stoss and lee sides of the larger bedform. Although it is possible that the second-order bounding surfaces represent periods of erosion of the lee face, it is more likely that these surfaces were formed by the migration of dunes down a draa that lacked a well-developed slipface.

Draas can be oriented either transverse or parallel to the prevailing wind direction. In the absence of plan view exposures that clearly show the geomorphic shape of the bedform, it is necessary to examine the orientation of the smaller-scale cross-stratification relative to the dip of the major bounding surface

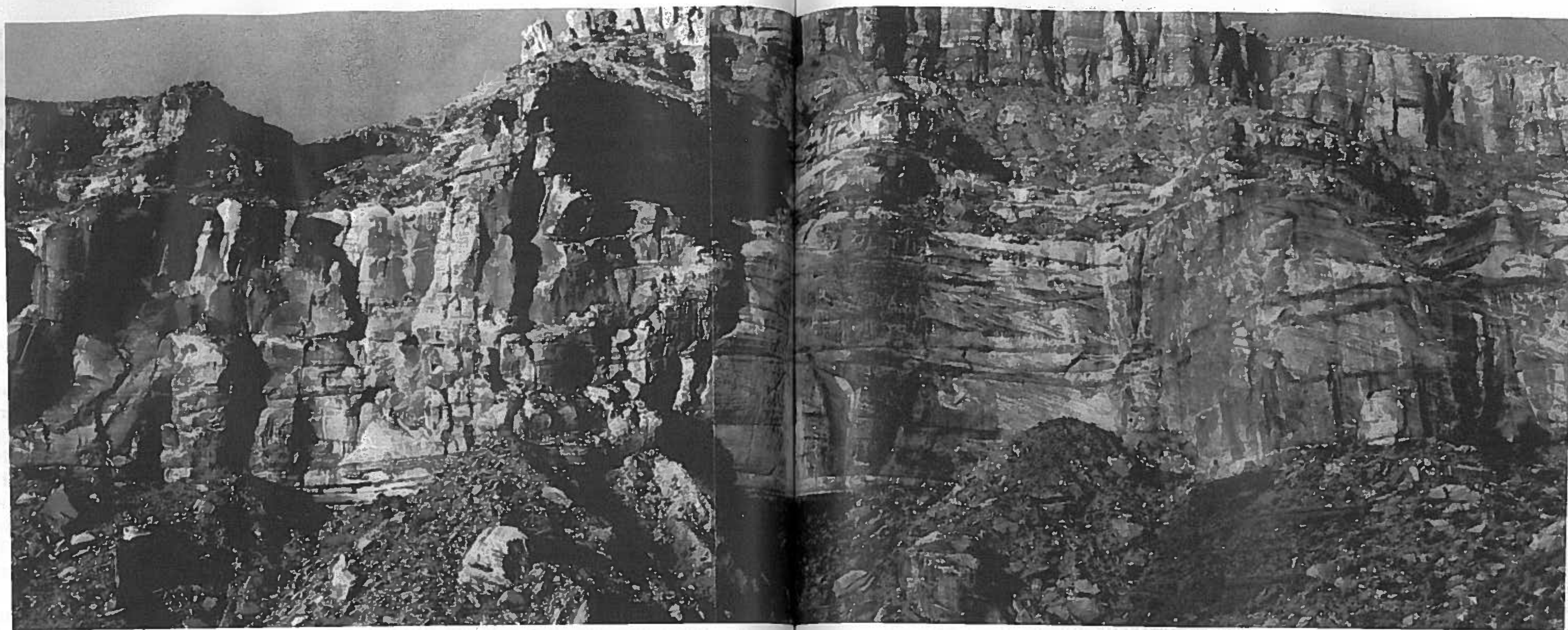

Figure 10.10. Sets of complexly cross-stratified sandstone exhibiting both first- and second-order bounding surfaces and intraset cross-stratification. Hualapai Trail, western Grand Canyon.

(Rubin and Hunter 1985). In draas oriented parallel to the main wind direction (and, therefore, long-term sand-transport direction), the smaller-scale sets should exhibit current directions parallel to the draa set. In all sections of the Coconino Sandstone examined, the internal sets are oriented parallel to the dip of the master surfaces. Thus, it appears that flow-transverse bedforms also formed these sets.

Of minor importance in the Coconino Sandstone are sets of trough cross-stratification. These sets, typically less than a meter thick, are filled by wind-ripple laminae. Winds that scoured the surface probably formed the troughs. The pit itself then was filled by wind ripples migrating into the scour during periods of reduced wind strengths. Blakey et al. (1983) have proposed a similar origin for these units in the underlying Schnebly Hill Formation in Oak Creek Canyon.

SUMMARY

The sediments and trace fossils of the Coconino Sandstone record the advance and passage of a major eolian sand sea. Both invertebrate and vertebrate traces indicate that the substrate was, for the most part, very dry, though it is clear that light rainfall or dew periodically moistened the dune surfaces. Although we know that the erg was characterized by both simple and complex bedforms that apparently were oriented approximately transverse to the main direction of sand transport, we have much to learn about the distribution of the bedforms and the dynamics of the Coconino erg.