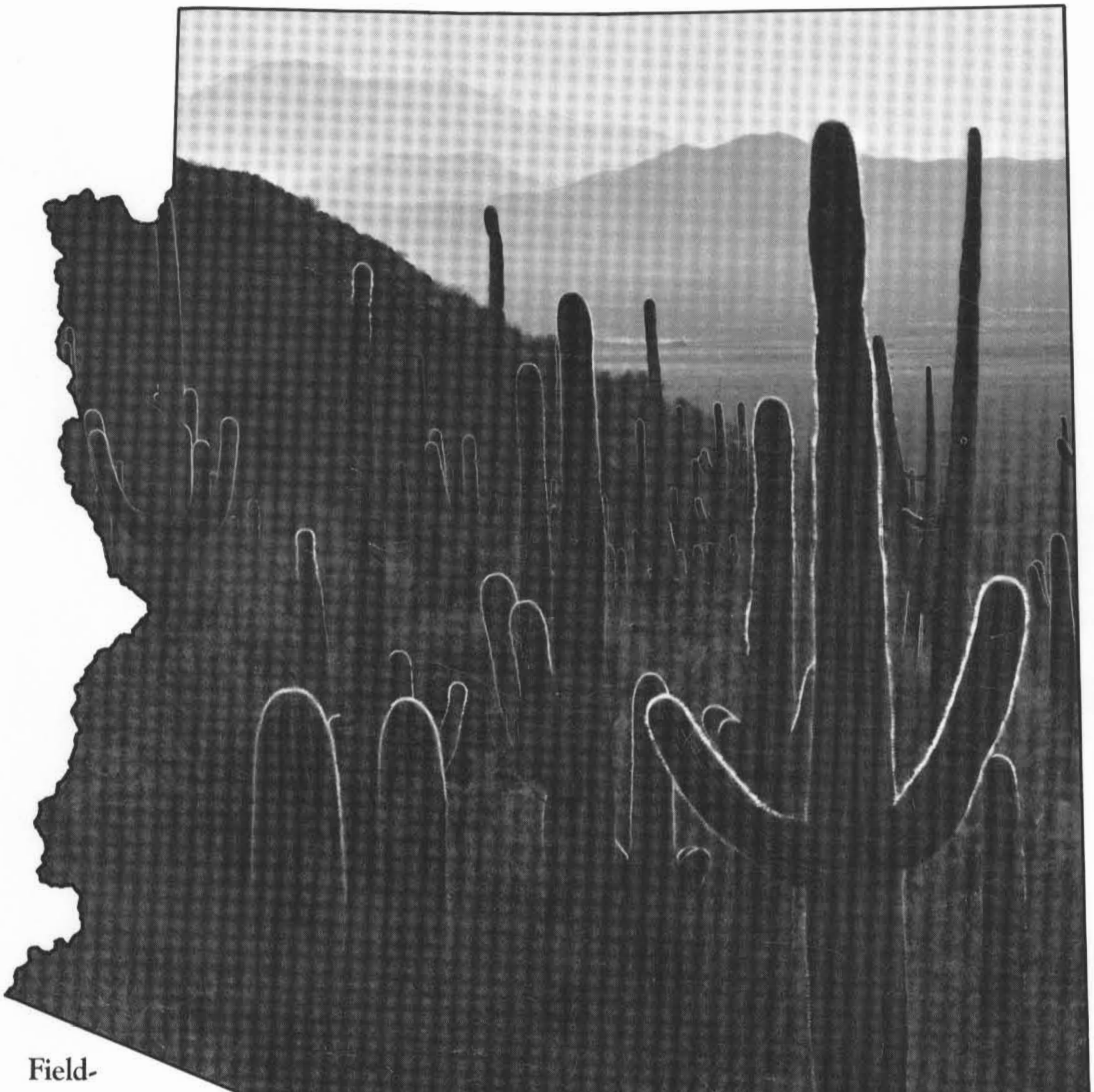
Geologic Diversity of Arizona and Its Margins: Excursions to Choice Areas



Field-Trip Guidebook
100th Annual Meeting
The Geological Society of America
Phoenix, Arizona, October 26–29, 1987

Arizona Bureau of Geology and Mineral Technology Special Paper 5

Field Guide to Sedimentary Structures in the Navajo and Entrada Sandstones in Southern Utah and Northern Arizona

David M. Rubin and Ralph E. Hunter
U.S. Geological Survey
Menlo Park, California 94025

INTRODUCTION

This field-trip guide describes the common sedimentary structures that occur in eolian sands. The outcrops that are described occur in the Navajo and Entrada Sandstones between the areas of Page, Arizona and St. George, Utah (Figure 1), but the sedimentary structures of these two sandstones are typical of most eolian deposits. The main part of the guide discusses the geologic setting and the origin of the various structures, and the road log discusses which structures are best displayed at selected outcrops.

Geologic Setting

The late Paleozoic and early Mesozoic were the peak times of eolian deposition in the western interior of the United States. On this field trip we will be looking at the Upper Triassic-Jurassic Navajo Sandstone and the Middle Jurassic Entrada Sandstone, the principal lower Mesozoic eolian deposits of the southwestern Colorado Plateau.

Kocurek and Dott (1983) have summarized the Jurassic paleogeography and paleoclimate of the western interior. During this time the southern Colorado Plateau lay at a paleolatitude of 10°-20° north, and paleonorth was a little westward of present north. Eustatic sea levels were relatively low through the late Paleozoic and early Mesozoic, and marine waters were much less extensive than during the early to middle Paleozoic and the Cretaceous. The climate was generally warm and dry. Winds, as interpreted from crossbedding, were from northerly directions ranging from the present northwest to present northeast (Poole, 1962; Kocurek and Dott, 1983).

The entire upper Paleozoic to lower Mesozoic section thickens and contains a greater proportion of marine beds toward the west. The axis of the depositional basin ran through the present Basin and Range province, an area from which Triassic and Jurassic deposits were later largely removed (Peterson, 1972; Kocurek and Dott, 1983). West of the basin axis were tectonic highlands of which little is known. The section thins and contains a greater proportion of nonmarine beds eastward toward the Transcontinental Arch and toward local uplands (the remnant Ancestral Rockies in Colorado and the Mogollon Highlands in southern Arizona). Sources of fluvial sediment were largely to the south and east of the depositional areas (Poole, 1961). Sources for the huge volumes of upper Paleozoic and Mesozoic eolian sand are poorly known, but a northern cratonic source on the Canadian Shield was probably important (Kocurek and Dott, 1983).

Stratigraphy

The Navajo Sandstone is the uppermost formation of the Glen Canyon Group, which also contains, in descending order, the Kayenta Formation and the Moenave Formation or its lateral equivalent, the Wingate Sandstone (Figure 2). The entire group may be Jurassic in age, although a Triassic age cannot be ruled out for the lower part (Peterson and Pipiringos, 1979). At the base of the group is a regionally traceable unconformity, designated the J-0 unconformity by Pipiringos and O'Sullivan (1978), that locally separates the Moenave Formation or Wingate Sandstone from the underlying Chinle Formation, of Triassic age.

The Moenave and Kayenta Formations are both composed largely of fluvial redbeds. The Moenave Formation contains the ledge-forming Springdale Sandstone Member at its top. Reddish-brown mudstone of floodplain origin is the dominant rock type in the Kayenta Formation, although fluvial sandstone and siltstone are also present. Thin intervals of eolian sandstone occur in the upper part of the Kayenta, and the contact with the overlying Navajo Sandstone is gradational (Middleton and Blakey, 1983). The fluvial sandstones of the Kayenta are less well sorted and mineralogically less mature than the eolian sandstones of the upper Kayenta and Navajo, and the fluvial crossbeds dip westward in contrast to the southeastward crossbed dips in the eolian sandstones in the upper Kayenta and lower Navajo (Middleton and Blakey, 1983). In the area around Kanab, Utah, the intertonguing between the Kayenta and Navajo is on a scale large enough that the tongues have been named; here the Lamb Point Tongue of the Navajo Sandstone is separated from the main body of the formation by the Tenney Canyon Tongue of the Kayenta Formation (Figure 2).

The Navajo Sandstone, like almost all the upper Paleozoic and Mesozoic eolian sandstones of the western interior, is composed largely of well-sorted, fine- to medium-grained sandstone. In the field-trip area the lower part has considerable evidence of the presence of water (wet-interdune deposits and early postdepositional deformation that required saturation of the sand) and has southeastward-dipping crossbedding, whereas the upper part has little evidence of wetness and has southwestward-dipping crossbedding (Marzolf, 1983). Accompanying these differences is a change from reddish in the lower part to light gray in the upper part, but this color change reflects diagenetic processes.

In the Zion area the Navajo Sandstone is separated from the overlying Temple Cap Sandstone by the J-1 regional unconformity, and the Temple Cap is in turn separated from the overlying Carmel Formation by the J-2 unconformity (Pipiringos and O'Sullivan, 1978; Peterson and Pipiringos, 1979). Eastward from Zion, the Temple Cap pinches out and the J-2 unconformity separates the Navajo from overlying formations (Figure 2). The upper part of the Temple Cap Sandstone is eolian sandstone, and the lower part is waterlaid sandstone, siltstone, and mudstone.

The Carmel Formation consists of sandstone, siltstone, mudstone, carbonate rocks, and evaporites. Many of the clastic rocks are reddish. The depositional environments ranged from dominantly marine on the west to dominantly tidal flat and sabkha on the east and south (Blakey and others, 1983). Farther to the southeast, in the Page area, the lower part of the Carmel intertongues with the eolian Page Sandstone (Peterson and Pipiringos, 1979). The crossbedding in the Page Sandstone dips to the south-southwest (Blakey and others, 1983).

The Carmel is overlain by the Entrada Sandstone, a widespread unit of dominantly but not entirely eolian origin (Kocurek and Dott, 1983). In the field-trip area the Entrada is entirely eolian, and the crossbedding dips to the southwest. The Entrada is overlain unconformably by the Upper Cretaceous Dakota Formation in the field-trip area, although Jurassic formations younger than the Entrada appear a

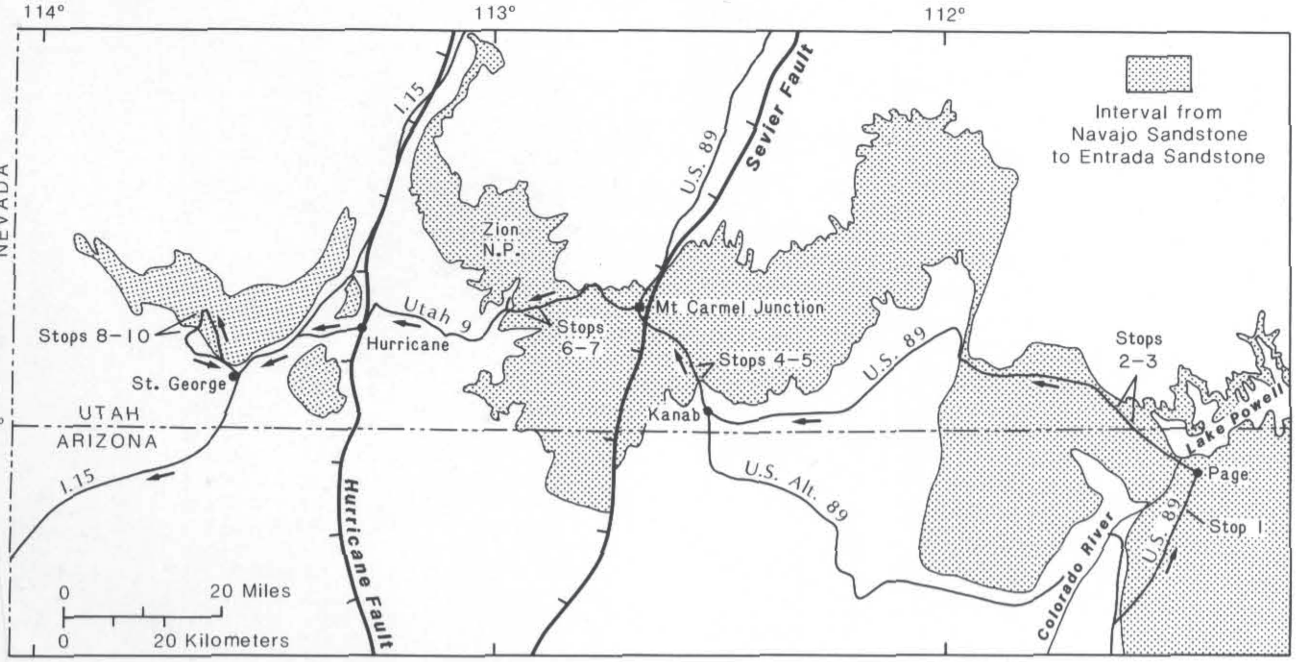


Figure 1. Map showing field-trip route, stop locations, and outcrop belt of the stratigraphic interval between the base of the Navajo Sandstone and the top of the Entrada Sandstone.

few miles to the east (Thompson and Stokes, 1970). To the west, between Page and Kanab, the Entrada pinches out and the Dakota rests on the Carmel Formation (Figure 2).

SEDIMENTARY STRUCTURES

Sedimentary structures in eolian sands fall into several categories: small-scale primary structures (fine structures) that record the depositional processes that operated at the sediment surface, larger-scale cross-stratification that records dune morphology and behavior, and miscellaneous structures such as trace fossils, deformational structures, or evaporite-related structures that record other postdepositional or syndepositional physical or chemical processes. The following discussion considers the origin and significance of these structures.

Fine Structures

Fine structures, as defined by Hunter (1985), are the smallest sedimentary structures, particularly the thinnest strata, that exist in a rock. No upper limit can be placed on the scale of these structures, but most strata that qualify as fine structures are less than a few centimeters thick. In general, strata that qualify as fine structures are without finer lamination internally. The classification of fine structures used here is genetic and, like all genetic classifications, cannot always be applied with confidence.

The three principal types of fine structure that are formed in dry sand are climbing-wind-ripple structures, grainfall lamination, and sandflow cross-stratification (Hunter, 1977; Fryberger and Schenk, 1981; Kocurek and Dott, 1981). Several types of fine structure form in moist to submerged interdune deposits, but only a few types have been studied sufficiently to be distinguished with confidence (Hunter, 1981).

Climbing-wind-ripple structures

Deposits formed by the climbing of wind ripples are common in most outcrops of the Navajo and Entrada Sandstones of the southern Colorado Plateau. This commonness is fortunate for the geologist concerned with the interpretation of depositional environments because the distinctive internal structure of such deposits provides the best single piece of evidence for an eolian origin of these formations. Excellent examples of climbing-wind-ripple structures can be seen in the Navajo Sandstone at the bottom of Water Holes Canyon (Stop 1) and in the Entrada Sandstone northwest of Page (Stops 2 and 3). To see the structures to best advantage, look for places where the outcrop surface bevels the bedding at a low angle.

Climbing-wind-ripple structures can form on any wind-rippled surface that is receiving net deposition. In the Navajo and Entrada Sandstones, climbing-wind-ripple structures occur most commonly in bottomset or toeset (basal-apron) deposits that intertongue upward with foreset (including slipface) deposits. In horizontal exposures of trough crossbedding, climbing-wind-ripple structures are best developed at the lateral margins of the troughs. Climbing-wind-ripple structures can make up complete sets of crossbeds, provided the slope angles were everywhere less than the angle of repose. In a few thin sets of crossbeds that represent nearly critically climbing small superimposed dunes (especially in the Entrada Sandstone at Stops 2 and 3), climbing-wind-ripple structures make up the preserved topset deposits.

Any climbing-ripple structure can be thought of as consisting of two components: translatent strata and rippleform strata, the latter including ripple-foreset cross-laminae (Hunter, 1977). In climbing-wind-ripple structures, however, the rippleform strata are generally not visible. A translatent stratum, or what Fryberger and Schenk (1981) call a ripple-produced stratum, is the deposit left by a single migrating ripple during its lifetime. Where the ripples climbed at an angle so low that the stoss sides of the ripples were eroded (in which case the angle of climb is called subcritical or stoss-erosional), the translatent strata have sharp, erosional contacts. The translatent stratification formed by stoss-erosionally climbing wind ripples long went unrecognized as having anything to do with ripples, for it rarely has any waviness suggestive of a rippled bed or any ripple-foreset cross-lamination suggestive of ripple migration (Figure 3). However, this kind of stratification can often be recognized (or at least be suspected) at a glance by its "pin-stripe" appearance, a small-scale cyclicity that is suggestive of varves (Figure 3). On closer inspection, the translatent strata can often be seen to be inversely graded, another good indicator of wind-ripple origin.

Translatent strata differ from other strata in that the imaginary time lines within them are not parallel to the bounding surfaces but rather intersect them; almost any set of crossbeds is a translatent stratum. Furthermore, when the translatent strata are of climbing type, the time lines cross from one layer into adjacent ones, meaning that

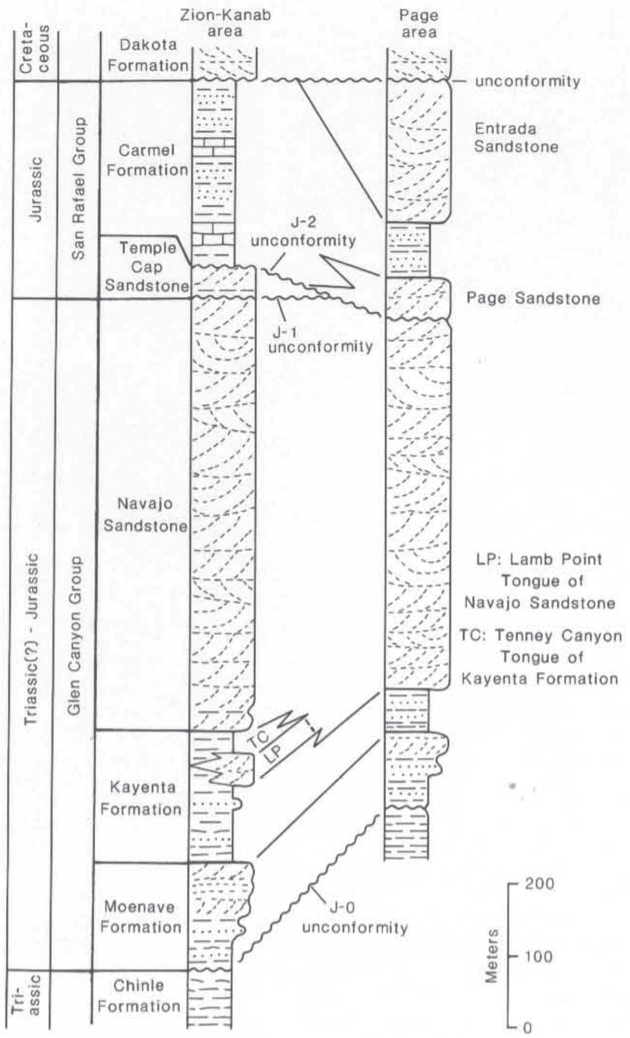


Figure 2. Columnar sections of Glen Canyon and San Rafael Groups in southwestern Utah (Zion-Kanab area) and northwestern Arizona (Page area).

all the strata in a set were forming at the same time. Of course, the law of superposition still holds along any vertical column.

The rarity of ripple-foreset cross-laminae within the translatent strata formed by climbing wind ripples derives from the nature of wind ripples. In wind ripples, in contrast to current ripples in water, the lee slopes dip so gently that avalanching does not occur, and consequently there are no miniature sandflow cross-strata defining the ripple foresets. Moreover, the thin lamination produced by wind gusts is naturally faint and tends to be blurred by the impacts of saltating grains. Although the rarity of ripple-foreset cross-laminae makes the translatent nature of the strata formed by climbing wind ripples difficult to recognize, this rarity helps in distinguishing climbing-ripple structures formed by wind ripples from those formed by current ripples.

Grainfall lamination

Grainfall lamination is formed when previously saltating grains pass into the relatively calm air in a zone of flow separation leeward of a dune crest and settle out onto the lee slope or basal apron in front of the dune. As they fall, the grains lose the forward momentum that was imparted to them by the wind and strike the surface so gently that the thin lamination produced by grain segregation during wind gusts is not destroyed. Grainfall lamination is distinguished from translatent stratification formed by climbing wind ripples by the

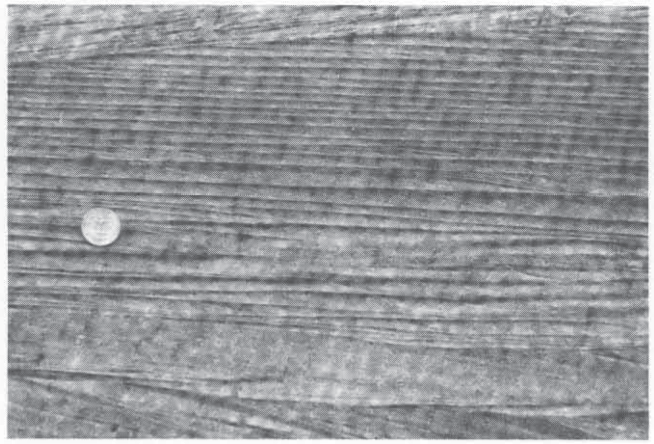


Figure 3. Climbing-wind-ripple structure (showing typical "pin-striped" appearance) in the Entrada Sandstone northwest of Page, Arizona (near Stop 3); quarter-dollar for scale. A few ripple-foreset cross-laminae are visible within some of the stoss-erosionally climbing translatent strata; they indicate that the ripples migrated toward the right. The 5-cm-thick layer with indistinct cross-laminae (near the bottom of the photograph) is a translatent stratum that probably was produced by a small dune. The faint lamination within this stratum is of grainfall type except for some climbing-wind-ripple translatent strata just above the base of the dune-formed translatent stratum.

extreme evenness of grainfall lamination and by its typical faintness and lack of small-scale cyclicity (pin-striping). Distinguishing grainfall lamination from plane-bed lamination, which is formed under conditions of intense saltation in strong winds, is probably impossible in hand specimens. Winds strong enough to produce plane beds are not common in modern dunes, however, especially on the moderately steep lee slopes where grainfall deposition is most common.

Grainfall lamination is uncommon in the Navajo and Entrada Sandstones of the field-trip area. Probably this is because of the generally large size of the dunes that formed these sandstones. On large dunes grainfall deposition may never reach the base of the slipface, and the grainfall deposits of the upper slipface tend to be destroyed by avalanching. Even if they escape destruction by avalanching, grainfall deposits of the upper slipface are usually destroyed in the course of dune migration. Preserved grainfall deposits in the field-trip area are generally found in thin sets of cross-strata formed on small dunes; foreset cross-strata of grainfall type are associated with topset deposits formed by climbing wind ripples in some thin sets of cross-strata in the Entrada Sandstone at Stops 2 and 3. A 1-m-thick set of cross-strata of grainfall type can be seen at Stop 5 (Figure 4). Grainfall or, less probably, plane-bed deposits also occur in some large-scale sets of concordant cyclic cross-strata in the Navajo Sandstone, as at Stop 7 (Hunter and Rubin, 1983).

Sandflow cross-strata

The large-scale, steeply dipping cross-strata that are generally thought of as classically eolian were formed by sandflows (noncoherent avalanches) on the slipfaces of tall dunes. In dip cross sections, sandflow cross-strata are most easily identified by their lack of curvature and by their distinctive toes or basal pinchouts (Figure 5), and in strike cross sections or horizontal exposures they are most easily identified by their thickness (typically a few centimeters) and lenticularity.

Two measurable characteristics of sandflow cross-strata are useful in quantitatively interpreting eolian deposits (Hunter, 1981). Because the original dip angles of sandflow cross-strata have a narrow range (the average dip angle being 32-33°), the present dip angle is a good indicator of the amount of compaction. A reduction in dip angle to 27°, a common amount of reduction in the upper Paleozoic and lower Mesozoic sandstones of the Colorado Plateau, indicates 19 percent compaction. And because the thickness of sandflow layers is positively correlated with slipface height (although the relationship has

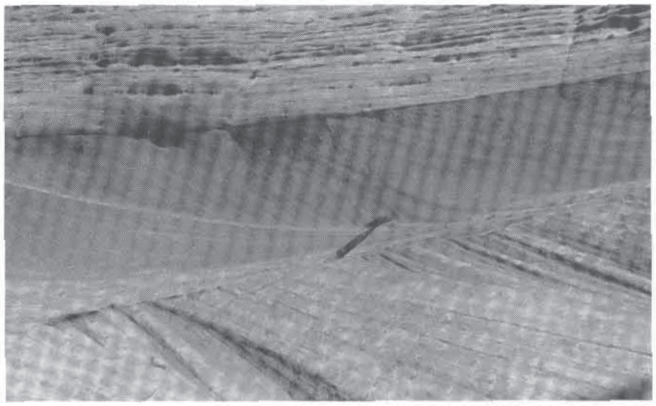


Figure 4. A 1-m-thick set of crossbeds of sandflow and grainfall type in the Navajo Sandstone north of Kanab, Utah (Stop 5). A wedge-shaped body of sandflow cross-strata occurs just above the hammer, but the rest of the set is composed of grainfall laminae identifiable by their faintness and extremely high tabularity (low lenticularity). Note difference in weathering style between the grainfall deposits and the underlying and overlying deposits, which were formed largely by climbing wind ripples.

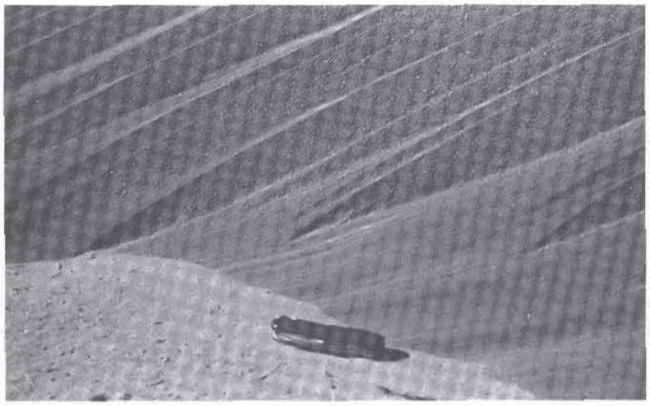


Figure 5. Toes (basal pinch-outs) of sandflow cross-strata in the Navajo Sandstone at Water Holes Canyon, Arizona (Stop 1). The sandflow cross-strata abut on and intertongue with basal-apron deposits in which the fine structure is probably grainfall lamination but may include intervals of climbing-wind-ripple translatent stratification.

much scatter), the sandflow cross-strata can give some indication of dune size; this relationship has recently been quantified by Kocurek (unpublished data).

Sandflow cross-strata are common in both the Navajo and Entrada Sandstones in the field-trip area. However, they are not common in all sets of cross-strata, and their rarity in some sets does not indicate a subaqueous origin for those sets.

Wet-interdune fine structures

Interdune areas can be dry, moist, or ponded. Where the sand is dry, the same fine structures that form on gently sloping dune surfaces can form. Where the sand is moist, the deposition of wind-blown sand involves the adhesion of grains to the surface, and the resulting fine structures are called adhesion structures (Kocurek and Felder, 1982). Where interdune ponds or streams occur, the various types of fine structure formed by oscillatory flow, unidirectional flow, and settling out in quiet water can form. Besides having distinctive structures, some interdune deposits differ texturally from associated dune deposits. The interdune deposits, although mostly sand, can have both coarser material (coarse sand and very fine pebbles) and finer material (silt and mud) than dune deposits.

The most common wet-interdune fine structure (or complex of fine structures) in the upper Paleozoic and Mesozoic eolian sandstones of the western interior is a type of stratification characterized by a small-scale irregular lenticularity or crinkly appearance (Figure 6). This lenticularity must reflect some combination of deposition on irregularly wavy surfaces and small-scale deformation soon after deposition. Many physical, chemical, and biological processes may have been involved (Hunter, 1981), but recent work in Arabian dune fields (Fryberger and others, 1983) suggests that the precipitation and dissolution of evaporite minerals within the sediment were the processes most important in giving the stratification its distinctive appearance. The best examples of this type of fine structure in the field-trip area are in the transition zone between the Navajo Sandstone and the underlying Kayenta Formation, for example, at Snow Canyon (Stop 10). Besides occurring in interdune deposits, the structure is common throughout the arid floodplain or sabkha deposits of the Kayenta Formation.

Easily identified adhesion structures are surprisingly rare in the upper Paleozoic and Mesozoic interdune deposits of the western interior, probably because of processes involving evaporite minerals, which can either prevent adhesion structures from assuming their typical form or alter the structures after they form. The only undoubted adhesion structures that we know of in the field-trip area are some thin layers characterized by quasiplanar adhesion lamination and

Figure 6. Wet-interdune or sabkha sandstone near the base of the Navajo Sandstone at Snow Canyon, Utah (Stop 10). Eolian crossbedding is subordinate in this part of the section, so that "sabkha" may be a more appropriate facies designation than "interdune." Note the irregularity of the bedding, much of which in this exposure is of deformational origin.

climbing adhesion ripples in dune deposits (not interdune deposits) of the Entrada Sandstone at Stop 3. The formation of adhesion structures at levels above the interdune troughs requires rainfall, not just the intersection of the water table by the troughs.

Crossbedding

Simple and compound crossbedding

Where fine structures are deposited on an inclined surface such as the flank of an eolian dune, the resulting deposit is an inclined bed called a crossbed. As a dune migrates, successive crossbeds are deposited on its advancing surface, thereby producing a set of crossbeds. Unless net deposition occurs while the dune migrates, the set of crossbeds will be thoroughly reworked by the next dune having an equally deep trough that migrates across the depositional site. In contrast, in an area undergoing net deposition, dunes will move upward (climb) relative to the generalized depositional surface (Allen, 1963; Brookfield, 1977; Rubin and Hunter, 1982). The result of this process of dune climbing (migration accompanied by net deposition) is a coset of crossbeds, shown in simplest form in row 1 of Figure 7.

Figure 7. Eight examples of dunes and crossbedding. The figure illustrates bedding produced by dunes with: simple (two-dimensional) and complicated (three-dimensional) morphology; simple (invariable) and complicated (variable) behavior; and transverse, oblique, and longitudinal orientations relative to the resultant sand transport direction (from Rubin, 1987b).

TWO-DIMENSIONAL
Two-dimensional bedforms are straight and parallel in plan form; the flanks of the bedforms have the same strike at all locations. Two-dimensional bedforms produce two-dimensional cross-bedding: cross-bedding in which all foresets and bounding surfaces have the same strike. In plots showing the direction and inclination of dips of cross-beds and bounding surfaces, dips of all planes plot along a single straight line through the center of the plot.

INVARIABLE
Invariable bedforms are those that do not change in morphology or path of climb. Cross-bedding deposited by invariable two-dimensional bedforms has bounding surfaces that are parallel planes; their poles plot as a single point.

TRANSVERSE, OBLIQUE, AND LONGITUDINAL
Transverse, oblique, and longitudinal cross-bedding are not distinguishable unless bedforms are at least slightly three-dimensional (see below).

VARIABLE
Variable bedforms are those that change in morphology or path of climb. Variability causes dispersion in the inclination of bounding surfaces. Cross-bedding deposited by variable two-dimensional bedforms has bounding surfaces with a constant strike but with varying inclination; their poles plot as a straight line that parallels the line of cross-bed dips.

TRANSVERSE, OBLIQUE, AND LONGITUDINAL
Transverse, oblique, and longitudinal cross-bedding are not distinguishable unless bedforms are at least slightly three-dimensional (see below).

THREE-DIMENSIONAL
Three-dimensional bedforms are curved in plan form or have plan-form complexities such as scour pits or superimposed bedforms with a different trend from the main bedform; the strike of the flanks varies with location. Three-dimensional bedforms produce three-dimensional cross-bedding: cross-bedding in which foreset and bounding-surface strikes vary with location; dips of foresets do not plot along a single straight line through the center of polar plots.

INVARIABLE
Cross-bedding deposited by invariable three-dimensional bedforms has bounding surfaces that are trough-shaped; bounding-surface dips in a single trough (or in identical troughs) plot as a nearly straight line.

PERFECTLY TRANSVERSE
Plots of cross-bed and bounding-surface dips have bilateral symmetry; the axis of symmetry is the same for both plots; dip directions are distributed unimodally.

OBLIQUE, IMPERFECTLY TRANSVERSE, OR IMPERFECTLY LONGITUDINAL
Plots of cross-bed and bounding-surface dips do not have bilateral symmetry; cross-bed dips are asymmetrically distributed relative to bounding-surface dips.

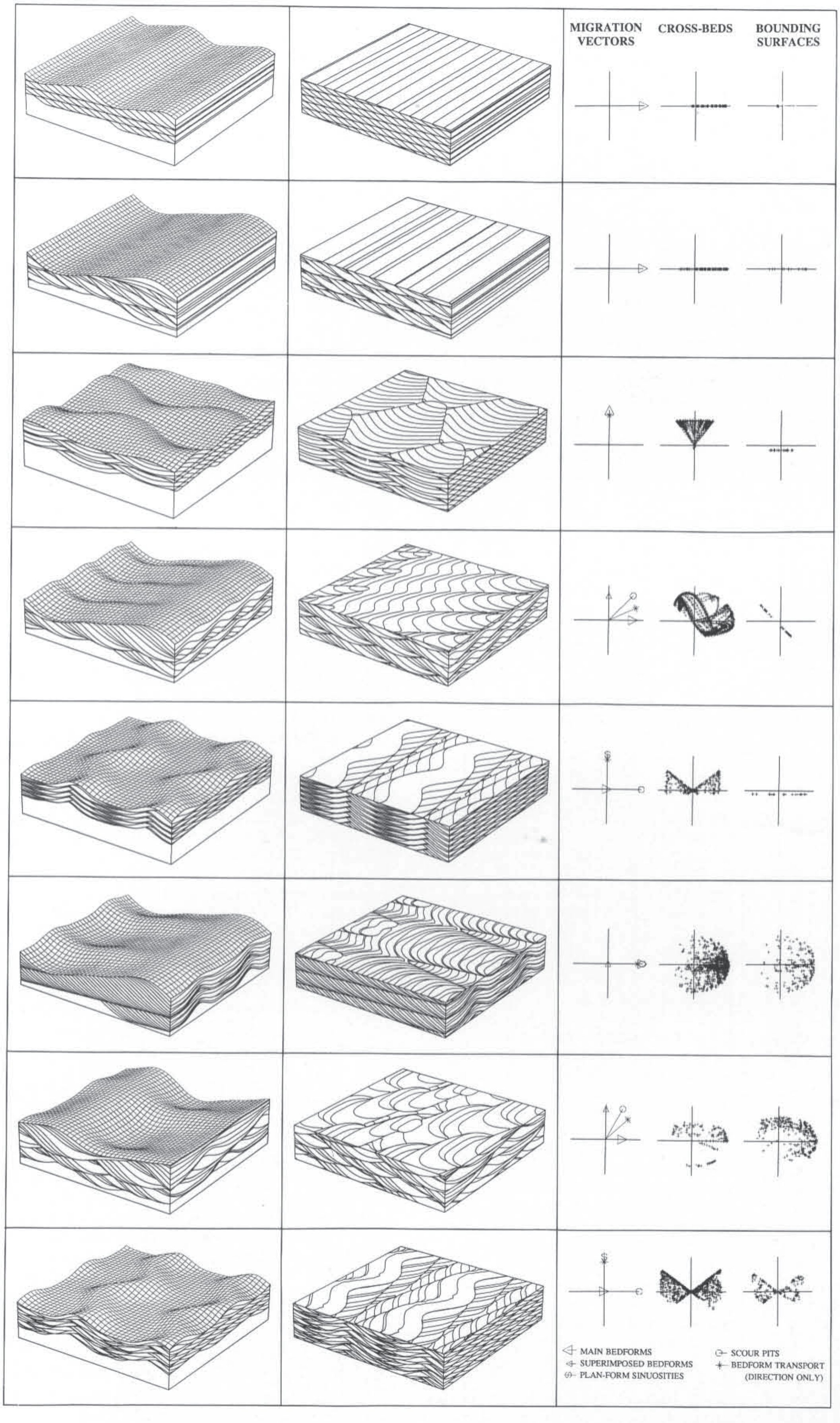
PERFECTLY LONGITUDINAL
Plots of cross-bed and bounding-surface dips have bilateral symmetry; dip directions may be distributed bimodally (as shown) or may be unimodal as a result of migration of the nose of the main bedform. Perfect longitudinality is evidenced by vertical accretion of bedforms; cross-beds dip in opposing directions on opposite flanks.

VARIABLE
Bounding surfaces have complex shapes produced by such processes as zig-zagging of scour pits; dips of bounding surfaces plot as scatter diagrams.

PERFECTLY TRANSVERSE
Same as perfectly transverse, invariable, three-dimensional cross-bedding.

OBLIQUE, IMPERFECTLY TRANSVERSE, OR IMPERFECTLY LONGITUDINAL
Same as oblique or imperfectly aligned, invariable, three-dimensional cross-bedding.

PERFECTLY LONGITUDINAL
Same as perfectly longitudinal, invariable, three-dimensional cross-bedding.


MIGRATION VECTORS
CROSS-BEDS
BOUNDING SURFACES
MAIN BEDFORMS
SUPERIMPOSED BEDFORMS
PLAN-FORM SINUOSITIES
SCOUR PITS
BEDFORM TRANSPORT (DIRECTION ONLY)

One of the most important goals of eolian sedimentology is to determine wind regimes and sand transport directions from crossbed geometry. The first step of this interpretive process is to reconstruct dune geometry and behavior from crossbed geometry. In general, dunes with simple geometry and simple behavior deposit simple crossbedding, whereas dunes with complicated geometry or complicated behavior deposit complicated crossbedding (Figure 7). Morphologic complications are caused by such features as sinuous, discontinuous, or irregular dune crestlines; scour pits in dune troughs; superimposed bedforms, peaks, spurs, or other topographic features; and morphologic differences between individual dunes in a dune field. Behavioral complications are produced by such processes as reversals in dune asymmetry or migration direction; random changes through time in the morphology of individual dunes; systematic changes in morphology of entire populations of dunes; splitting and merging of dunes; and creation, destruction, or changes in morphology or behavior of superimposed bedforms.

Where these behavioral and morphologic complications are absent, dunes that migrate in an area undergoing net deposition do not experience erosion on their lee slopes. Because of the lack of lee-side erosion, the sets of crossbeds deposited by these dunes lack internal erosion surfaces, and the crossbedding is defined as simple crossbedding. The simple crossbeds can have a narrow spread in crossbed dip directions (if the dune was relatively straight-crested, as in row 1 of Figure 7) or a wide spread in dip directions (if the dune was sinuous, as in row 3). In contrast, where dune morphology and behavior are sufficiently complicated, erosion can occur on the lee side of a dune either during erosional episodes (such as when the dune reverses) or at erosional sites (such as on the stoss sides of dunes superimposed on the lee side of the main dune). These erosional episodes or migrating topographic features with erosional sites produce internal erosion surfaces within the set of crossbeds deposited by the migrating dune (Figure 7, rows 2 and 4-8); such sets of crossbeds are defined as compound crossbeds (Harms and others, 1975; Rubin and Hunter, 1983). Behavioral and morphologic complications can occur together, producing crossbedding that can be extremely difficult to interpret (Figure 7, rows 6-8).

The effects of morphologic complications can often be distinguished from those caused by behavioral complications. In general, changes in dip direction indicate morphologic complications, whereas changes in dip inclination that are not accompanied by changes in dip direction indicate behavioral complications (Figure 7). (The exception to this generalization is where straight-crested dunes migrate directly down the lee slope of a larger straight-crested dune; the result of this morphologic complication is a set of crossbeds that vary in inclination but not direction.) More thorough discussion of this topic is given by Hunter and Rubin (1983), Rubin and Hunter (1983), and Rubin (1987a). In the following discussion we present examples of crossbedding deposited by dunes with behavioral complications (Figs. 8 and 9), morphologic complications (Figs. 10-12), and both kinds of complications (Figure 13).

Behavioral complications caused by annual wind cycles

Because dune morphology is controlled largely by wind conditions, changes through time in the wind regime can be expected to produce changes in dune morphology, provided that the wind conditions do not fluctuate so rapidly as to be filtered out by slowly responding dunes. Many of the relatively random changes in geometry of crossbeds in a set are undoubtedly produced by random fluctuations in dune geometry (in response to random flow fluctuations). The most noticeable and interpretable structures that are produced by flow fluctuations, however, are the cyclic crossbeds produced by cyclic fluctuations in dune morphology or fine structure, both of which can form in response to annual or other regular fluctuations in wind strength or direction (Stokes, 1964; Hunter and Rubin, 1983).

At least two kinds of cyclic crossbedding produced by fluctuating flow occur in the Navajo and Entrada Sandstones: conformable bedding cycles produced by cyclic changes in fine structure (Figure 8) and discordant bedding cycles produced by cyclic fluctuations in dune asymmetry or dune height (Figure 9). A fluctuating-flow interpretation is preferred for these examples of cyclic crossbedding because the crossbeds and internal bounding surfaces in the sets dip toward the same direction, thereby demonstrating that the dunes had a relatively simple two-dimensional morphology. Additional examples of crossbedding produced by fluctuating flow are given by Hunter and Rubin (1983). The cyclicity in the Navajo Sandstone is interpreted to be annual (Stokes, 1964; Hunter and Rubin, 1983) primarily because unusually strong winds would have been necessary to have caused the observed dune migration in shorter time cycles.

Morphologic complications caused by topographic features

Like crossbedding produced by fluctuating flow, the crossbedding produced by dunes with superimposed dunes or other morphologic complications can be either random (produced by random topographic features) or cyclic (produced by trains of similar superimposed dunes, spurs, or crestline sinuosities). Migration of the superimposed features is often accompanied by erosion on the stoss side followed by deposition on the lee side. The cyclic passage of such topographic features deposits cyclic sets of crossbeds that are grouped in a coset deposited by the main dune.

Two kinds of cyclic, complicated-morphology, compound crossbedding are conspicuous throughout the Navajo and Entrada Sandstones along the route of this field trip: (1) structures formed by small dunes migrating across the lee slopes of larger dunes (Figure 10), and (2) structures formed by dunes with scour pits that migrated along the troughs of the main dunes (Figure 11). The two kinds of structures listed above are geometrically related because intersecting troughs of the main dunes and the superimposed dunes form topographic depressions that behave geometrically like scour pits, and migration of the superimposed dunes can cause these depressions to migrate along the trough of the main dune. Regardless of the morphologic details, alongcrest migration of superimposed features causes a cyclic rotation in dip direction of crossbeds within the set (Figure 12).

Recognition of structures deposited by superimposed dunes is important, not only because such structures are useful for reconstructing in detail the original dune morphology, but also because the migration direction of the superimposed features relative to the main dune is an important indicator of whether the main dunes were transverse, oblique, or longitudinal. Superimposed features can be expected to move dominantly in one alongcrest direction if they are superimposed on an oblique or longitudinal dune. In contrast, on a transverse dune, superimposed features can be expected to show a less consistent preferred alongcrest migration (Rubin and Hunter, 1985). In the Entrada Sandstone in the vicinity of Page, Arizona, the compound crossbedding indicates that the superimposed bedforms had a pronounced tendency to migrate in an alongcrest direction (to the right of the migration direction of the main dunes). This characteristic indicates that the main dunes in this region were not aligned transverse to flow, and may have been more nearly longitudinal than transverse. The hypothesis of a relatively longitudinal trend of Entrada dunes is supported by observations to the east of the field-trip area (in northwestern New Mexico), where sand-body surface relief defines lineations (buried dunes) that trend northeast-southwest (Vincellete and Chittum, 1981), approximately parallel to the local sand transport direction inferred by Poole (1962) and Tanner (1965).

Combined complications

As might be expected, dunes having complicated morphology and complicated behavior can produce extremely complicated crossbedding. Examples of one of the more cyclic structures deposited by such dunes are the "zigzags" produced by lee-side spurs and scour pits that reverse their asymmetry and direction of alongcrest migration (Figure 13). This kind of behavior can be expected of dunes that are transverse to the long-term resultant transport direction and are subject to winds that have a reversing crest-parallel component of flow. Such reversals could be expected to cause opposing flanks of lee-side spurs to experience erosion and deposition alternately.

Miscellaneous Structures

Deformational structures

Several types of structures formed by deformation of unconsolidated sediment are found in the Navajo and Entrada Sandstones of

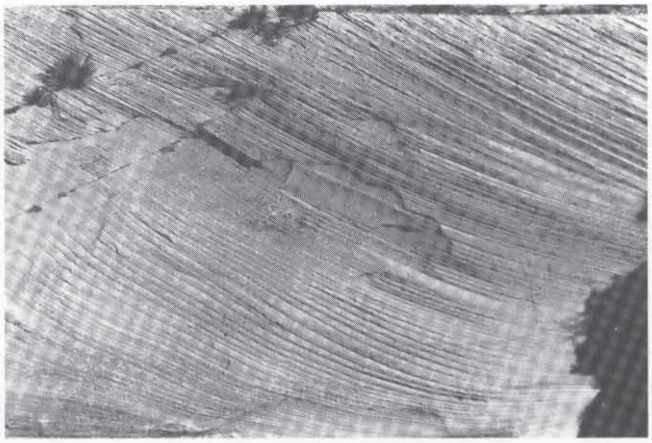


Figure 8. Conformable cyclic foresets in a 10-m-thick set of crossbeds in the Navajo Sandstone, Zion National Park, Utah (Stop 7). The cyclicity is due to an alternation of grainfall deposition and deposition by wind ripples and is interpreted to be the result of an annual cycle of fluctuations in wind speed or direction. In this particular set of crossbeds, the grainfall deposits are more resistant to erosion than the climbing-wind-ripple structures. The mean distance of dune advance represented by individual cycles is 0.3 m.

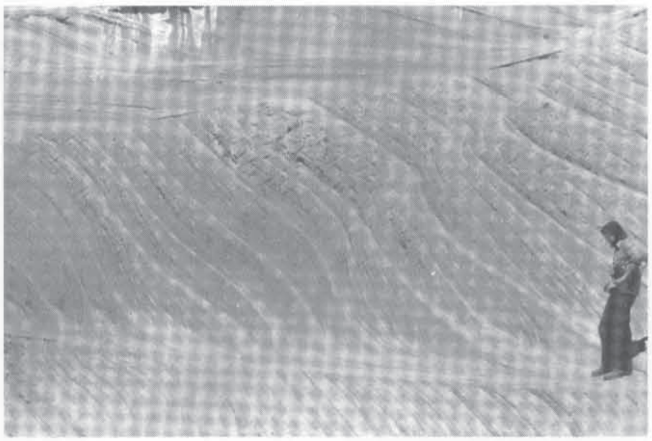


Figure 9. Cyclic compound crossbedding (discordant cyclic foresets) in a set of crossbeds in the Lamb Point Tongue of the Navajo Sandstone, north of Kanab, Utah (Stop 5). The crossbeds are largely of sandflow type. After each period of dune advance, the slipface was eroded back. During each period of slipface erosion, a wedge-shaped body of sand (light-colored) was deposited at the base of the slipface. The mean distance of dune advance represented by individual cycles is 1.5 m. The cyclicity, which must have involved fluctuations in wind direction, is interpreted to have been annual.

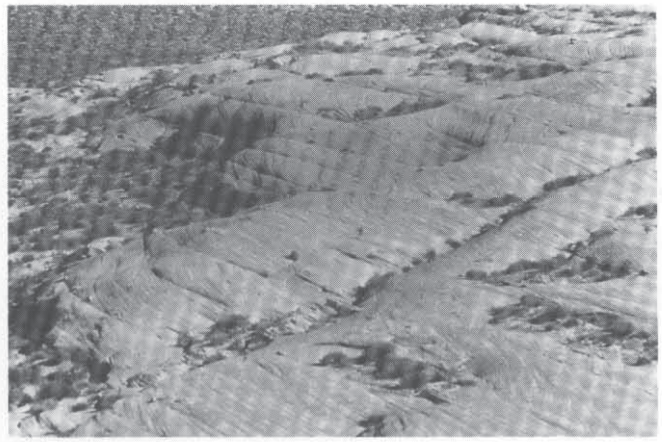


Figure 10. Compound crossbedding deposited by alongslope-migrating superimposed dunes; Entrada Sandstone near Stops 2 and 3. The main dune that deposited this entire coset migrated from left to right, and the superimposed dunes migrated toward the viewer.

Figure 11. Scalloped crossbedding produced by scour pits that migrated along the trough of the main dune; Navajo Sandstone in Zion National Park, Utah.

the field-trip area. The structures range in scale (fold amplitude or thickness of deformed zone) from 1 cm or less to 10 m or more.

The smallest deformational structures include some of the irregularities in the stratification of wet-interdune deposits (described under "Fine Structures") and structures formed by the sliding of coherent or semicoherent sediment masses down slipfaces. The latter, whose formation has been studied experimentally by McKee and others (1971), are not common; most of the avalanches were sandflows by the time they came to a stop, even if they started as slides.

Decimeter-scale contortions are common in flat-bedded intervals in the transition zone between the Navajo Sandstone and the underlying Kayenta Formation, for example, at Snow Canyon (Figure 6; Stop 10). We interpret these structures to have formed by a combination of loading and fluid escape. Some of the flat beds are certainly waterlaid, and even the eolian deposits probably became water-saturated after slight burial. The uneven loading due to the migration of sand dunes over the flats (a "rolling-pin" mechanism) then induced compaction and fluid escape.

Zones of contorted bedding as thick as 10 m or more are common in parts of the Navajo Sandstone, for example, in the lower part of the formation and in the Lamb Point Tongue north of Kanab (Stops 5 and 5A). Another example is well exposed in the walls of Water Holes Canyon (Figure 14; Stop 1). Structures of this type have been studied by Sanderson (1974), Doe and Dott (1980), and Horowitz (1982). All these workers agree that the contortions formed while the sand was saturated with water but that the deposits are eolian, not subaqueous. An origin below the level of the dune troughs is thereby indicated.

One characteristic of the deformation poses a paradox if the deformation took place below the level of the dune troughs. The overturning of crossbeds at the bases of deformed zones indicates shearing by an overriding sediment mass that moved approximately in the direction of crossbed dip (Figure 14). How could dune migration direction have controlled the direction of shearing below the level of the dune troughs? One proposed interpretation capable of resolving this paradox is that of Horowitz (1982), which involves earthquake-induced liquefaction centered beneath an interdune trough and the consequent collapse of the dune impinging on the trough. A second possible explanation is that the overriding dunes, which may have been as much as several hundred meters in height (Rubin and Hunter, 1982), acted like huge rolling pins that deformed the sediment in the underlying beds. More study of these structures is needed before their origin is fully understood.

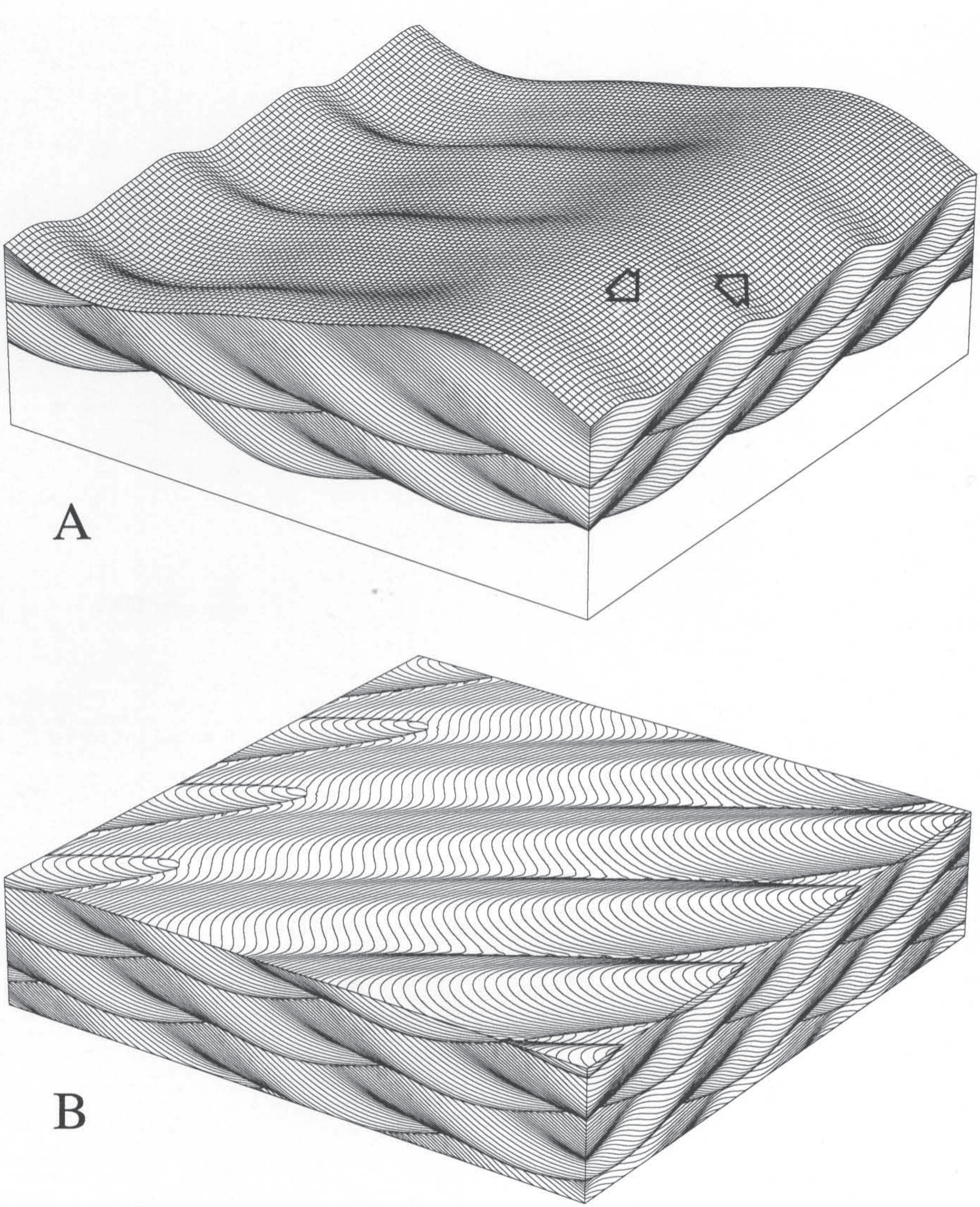


Figure 12. Computer-graphics model of the origin of scalloped crossbedding formed by superimposed dunes.
A. Dune morphology and vertical sections. The main dune is migrating left to right, and the superimposed dunes are migrating away from the viewer. Intersections of the troughs of the two sets of dunes produce scour pits, and the cyclic passage of these scour pits through the left vertical section produces the scallops.
B. Horizontal and vertical sections. Deposition causes the scour pits to migrate upward through the horizontal section, producing the "fingertip" structures.

Figure 13. Zigzag structure formed by a dune with reversing lee-side spurs; Lamb Point Tongue of the Navajo Sandstone, Kanab Creek, Utah (Stop 4). The spur that deposited this structure reversed back and forth across the outcrop plane while the main dune migrated toward the viewer.

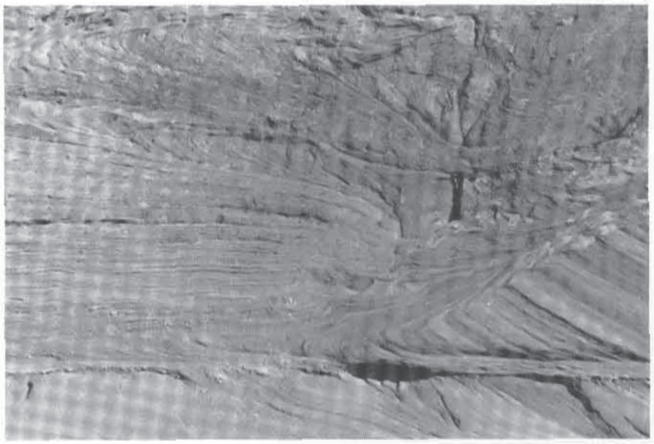


Figure 14. Large-scale deformational structure in the Navajo Sandstone at Water Holes Canyon, Arizona (Stop 1). The orientations of the drag folds indicate that the overriding sediment masses moved in the same direction as the crossbeds dip.

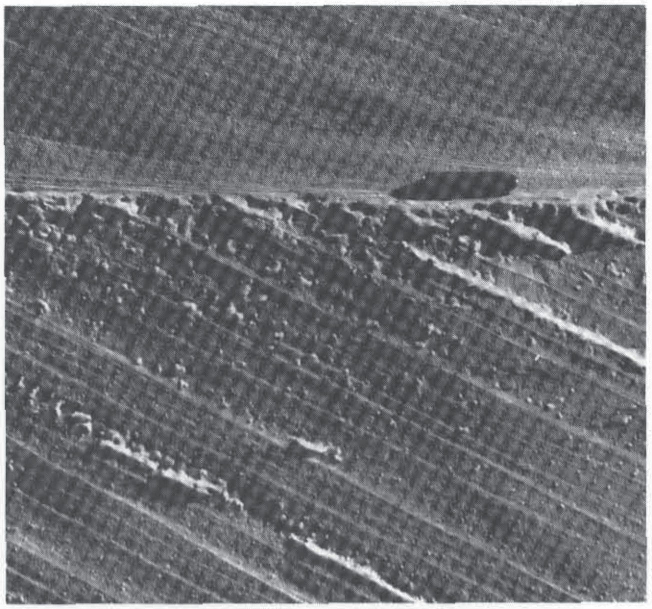


Figure 15. Nodules interpreted to have formed by replacement of evaporites; Navajo Sandstone at Water Holes Canyon, Arizona (Stop 1). The nodules are most abundant near the top of the set, which is the location where evaporation would have occurred if the dune troughs had been wet.

Trace fossils

Although trace fossils are far from common in eolian sandstones, except perhaps in wet-interdune deposits, they are known (Ahlbrandt and others, 1978) and indeed are fairly common in a few beds of the Navajo Sandstone in the field-trip area. We have no doubt that the beds in which the trace fossils occur are eolian. At least three kinds of trace fossils have been seen in dune-foreset or basal-apron deposits at Water Holes Canyon (Stop 1).

Small (diameter 0.5-1.0 cm), densely clustered, diversely oriented, unlined tubes filled by structureless sand occur near the base of a set of crossbeds at the bottom of the canyon. These trace fossils are probably intrastratal trails produced by an organism that lived on the lower lee slopes of dunes. Penetrating into the same set from its upper bounding surface are roughly vertical tubes about 10 cm in diameter filled by sand, some of which is stratified. These tubes evidently stood as open burrows and probably required damp sand to prevent collapse. The set of crossbeds in which these trace fossils occur, unlike so many other sets at the locality, has no carbonate nodules. The lack of these nodules, which are interpreted to have formed by replacement of evaporites, suggests that the moisture was provided by unusually fresh water, which was evidently attractive to organisms.

Near the rim of the canyon is a set of crossbeds (or several laterally unconnected sets at nearly the same stratigraphic level) that appears more weathered than sets above and below and that is penetrated from its upper bounding surface by mostly vertical tubes as much as 4 m long. The tubes have cores, probably the fillings of originally open burrows, that are 2-4 cm in diameter and that are surrounded by more resistant, iron-oxide-rich linings. Near the top of the set, some of the vertical tubes connect with horizontal tubes, some of which are cork-screwed. These burrows, which resemble the burrows of some modern crustaceans, evidently opened onto a surface that remained exposed long enough for the sand to become weathered. In places, wet-interdune deposits are preserved above the surface.

Evaporite structures

Centimeter-diameter carbonate-cemented nodules are abundant at many outcrops of the Navajo Sandstone and at some outcrops of the Entrada Sandstone. The nodules have several characteristics that suggest that some of them formed by replacement of evaporite crystals or nodules. In some sets of crossbeds that contain the nodules, the nodules are concentrated near the bottom or top of the set (Figure 15). Both of these sites are inferred to be favorable for evaporite precipitation, because both tend to be near the interface between air and interstitial waters at the time of deposition. Water can rise into lower foresets and evaporate, thereby depositing evaporites that are concentrated near the base of a set of crossbeds; and water that evaporates from a dune trough may deposit evaporites in the top of the underlying set. In the Entrada Sandstone east of the field-trip area (near Gallup, New Mexico) some nodules that are concentrated at the base of a set have preserved the shape of the evaporite crystals that were replaced (Figure 15), demonstrating that at least these nodules formed by replacement of evaporites. The origin of the more typical spherical or irregular nodules that appear to be randomly distributed within sets of crossbeds may also be related to evaporites, but such an origin is not conclusively demonstrated.

DAY 1

Mileage

0.0 Road log begins at intersection of U.S. 89 and U.S. Alternate 89, 24 miles south of Page, Arizona. Take U.S. 89 north towards Page. Ascend Echo Cliffs, passing through section from Moenkopi Formation at base of cliff to Navajo Sandstone at top.

12.7 Begin descending hill, passing outcrops of Page Sandstone and upper part of Navajo Sandstone for next 1 mile.

17.8 STOP 1. Navajo Sandstone at Water Holes Canyon. This stop is on property of the Navajo Nation. Persons following this route should obtain permission to stop here from the Navajo Nation Minerals Department, Window Rock, Arizona. It is worth this extra effort to stop at this site, because the structures are exceptionally well displayed. Allow several hours to follow the hike and to examine the structures described. Park on north side of canyon; enter gate on east side of highway; walk east along canyon rim until reaching a location where north wall of canyon becomes less steep; climb down to canyon floor. Continue east along the canyon floor, noting the following features: climbing wind-ripple structures (including some with preserved foreset cross-laminae); sandflows (visible in vertical sections in the canyon walls and in a few horizontal sections on the floor); two kinds of trace fossils; nodules — inferred to be replaced evaporite nodules — that in some beds are concentrated above and below bounding surfaces; small subsets of crossbeds deposited by superimposed dunes; scalloped crossbedding formed by fluctuating flow; and a zone of deformed bedding that includes almost the entire section in the canyon wall. At the site of the deformed zone, climb the north wall of the canyon and continue east along the rim to inspect the conformable fluctuating-flow bedding cycles. The following features are visible along the rim on the return route to the highway: a thick bed containing burrows that weather out of the rock (just below the elevation of the rim) and horizontal sections through trough-shaped sets of crossbeds — some of the sets are cyclically filled, first from one direction, then from another.

Continue north along U.S. 89 towards Page.

22.9 South turnoff to Page (Loop 89) on right. If time permits, continue north along U.S. 89. Page Mesa on right (east) is held up by Page Sandstone.

24.2 North turnoff to Page (Loop 89) on right.

24.7 Overlook on south side of Glen Canyon (gorge of Colorado River). Turn right on gravel road.

25.2 ALTERNATE STOP 1A. Contact of Page Sandstone and Navajo Sandstone. Park along road and walk 0.3 mile northeast to small mesa. Contact is near base of mesa, marked by large polygonal cracks at top of Navajo. Return to U.S. 89.

25.7 Turn right (north) onto U.S. 89, cross bridge over Glen Canyon.

26.2 ALTERNATE STOP 1B. Visitor center at Glen Canyon Dam. Navajo Sandstone exposed in walls of canyon. Note zone of large-scale contorted bedding. Return to Page for night.

DAY 2

Retrace route from Page to visitor center at Glen Canyon Dam.

0.0 Leave Glen Canyon Dam driving on Lakeshore Drive toward Wahweap Marina. Follow road north along west side of Lake Powell. Navajo Sandstone and Page Sandstone crop out along road for first 2 miles. Entrada Sandstone forms white cliffs on north side of Lake Powell.

4.3 Wahweap Marina.

6.8 Intersection of U.S. 89 and Lakeshore Drive. Cross U.S. 89, enter parking lot of Lake Powell Motel.

6.9 STOP 2. Entrada Sandstone on hillside in back of Lake Powell Motel. Park in the motel lot, and walk up the hill toward the closest outcrops. Note the climbing wind-ripple structures. Continue walking up the hill, and note that the crossbeds generally dip toward the right (toward the west or southwest), whereas the bounding surfaces that separate the crossbeds dip into the outcrop (roughly toward the south). This coset of beds was deposited by a large dune that migrated toward the south or southwest while superimposed bedforms migrated toward the west or northwest.

7.0 Turn left (west) onto U.S. 89 toward Kanab, Utah.

12.3 Turn left (south) onto dirt road. Close gate behind you. Continue to pulloff.

12.8 STOP 3. Entrada Sandstone in mesa walls. Note the excellent examples of fine structures, including climbing adhesion ripples, and note the scalloped crossbedding (on east side of dirt road) that was produced by scour pits that migrated toward the west, while the main dune migrated toward the south. Return to U.S. 89.

13.3 Turn left (west) onto U.S. 89 toward Kanab. Several outcrops of Entrada Sandstone along highway for next 7 miles.

29.7 Entrada Sandstone in cut on left (south) side of highway. North of highway for next 2 miles are outcrops of white Entrada Sandstone and underlying red beds of Carmel Formation.

35.9 For next 0.7 mile, pass downsection through steeply tilted Navajo Sandstone. Continuing on to Kanab, note Kayenta Formation and lower part of Navajo Sandstone in Vermilion Cliffs on right side of highway.

75.3 Intersection of U.S. 89 and U.S. Alternate 89 in Kanab, Utah. Turn right (north), continuing on U.S. 89 toward Mount Carmel Junction.

78.5 Thin eolian sandstone in Kayenta Formation on right (east) side of highway.

79.8 Gradational contact between red beds of Kayenta Formation and overlying Lamb Point Tongue of Navajo Sandstone.

80.6 Turn right (east) onto gravel road into Kanab Canyon.

81.4 STOP 4. Zigzagging spur in Lamb Point Tongue of Navajo Sandstone. Park along road. This structure is shown in Figure 13. Measure dip directions of crossbeds and bounding surfaces within these zigzags, and note that the beds dip roughly toward the east and south (on the northeast and southwest flanks of the southeastward-plunging spur). The spur shifted back and forth laterally while the main dune migrated toward the southeast.

Continue along gravel road.

81.7 Turn around at driveway to private property; return to U.S. 89.

82.8 Turn right (north) onto U.S. 89 toward Mt. Carmel Junction.

83.2 STOP 5. Compound crossbedding in the Lamb Point Tongue of the Navajo Sandstone. Structure is shown in Figure 9. Park along road. Note the cyclic character of these foresets. They are interpreted to be annual layers, and they document the net rate of dune advance (1.5 m/year). Continue northward on U.S. 89.

84.2 For next 0.2 mile, pass upsection through Tenney Canyon Tongue of Kayenta Formation.

84.6 ALTERNATE STOP 5A. Zone of large-scale contorted bedding near base of main body of Navajo Sandstone. Park along highway. Exposures on both sides of road. The top of the deformed zone is at least locally a shear surface, not an erosional surface. Continue northward on U.S. 89.

85.1 Exposure of large-scale scalloped crossbedding in Navajo Sandstone in canyon wall on left (west) side of highway.

91.0 South turnoff on left (west) to Coral Pink Sand Dunes. The dunes are 12 miles southwest of the highway and worth the drive if the wind has been strong enough to obliterate the unsightly offroad-vehicle tracks that usually mar this otherwise beautiful area.

91.5 North turnoff to Coral Pink Sand Dunes.

92.0 Cross Sevier fault. Navajo Sandstone on east side, Carmel Formation on downthrown (west) side of fault.

95.0 Mt. Carmel Junction. Turn left (west) on Utah Highway 9. Note cut through oolitic limestone in Carmel Formation on right (north) side of road 0.2 mile past intersection.

102.5 Stratigraphically highest point on road between Mt. Carmel Junction and Zion National Park. Upper Cretaceous Tropic Shale and Dakota Formations overlie Carmel Formation unconformably.

106.0 For next 0.5 mile, pass eolian sandstone outcrops of Temple Cap Sandstone. Next outcrops after that are of Navajo Sandstone.

106.7 On right (north) side of road is a transverse exposure of a trough set in which foresets are cyclic (annual layers).

107.4 Entrance to Zion National Park.

107.6 Outcrop north of road has trace fossils and a trough set with zigzagging infilling (not visible from road).

108.3 ALTERNATE STOP 6A. Parking area providing view of Checkerboard Mesa (Navajo Sandstone). Walk along road toward mesa to view very thick (more than 30 m) set of crossbeds south of road and east of mesa. Return to vehicle and continue westward on Utah 9.

108.9 STOP 6. Parking area provides view eastward of transverse section of concave-downwind foresets in a tabular set in Navajo Sandstone. The variable strike of the foresets indicates that the dune was curved in plan form, whereas the planar lower set boundary indicates that the dune trough was relatively uniform in elevation (i.e., no scour pits in the trough). Continue driving westward.

109.6 STOP 7. Parking area for exposures of conformable cyclic foresets in the Navajo Sandstone (Figure 8). Horizontal exposures reached by walking 0.1 mile back (eastward) along highway, then going south from road. Vertical exposures are visible from a point 0.2 mile west of parking area, looking at opposite wall of canyon. Return to vehicle and continue westward.

110.1 Valley-bottom exposure of compound crossbedding deposited by downslope-climbing dunes (not visible from road).

111.5 ALTERNATE STOP 7A. Parking area provides view northward of coset of cross-strata exposed on hillside (Navajo Sandstone). Walk back (north and east) 0.4 mile to see large-scale scalloped crossbedding on southeast side of same hill.

113.3 ALTERNATE STOP 7B. Parking area provides view eastward of coset of cross-strata (Navajo Sandstone) with bounding surfaces that rise southward (in dip direction of cross-strata). A few meters above road level on north side of road is an outcrop of flat-bedded wet-interdune sandstone. Walk westward 0.1 mile along road to trailhead north of road. About 0.3 mile along trail is horizontal exposure of desiccation-cracked interdune deposits.

113.4 Enter 1.1-mile-long tunnel.

114.8 First of five hairpin turns. Next 2 miles provide several good views of Navajo Sandstone exposed in canyon walls.

115.8 Stratigraphic level of thin eolian sandstone in Kayenta Formation. The sandstone forms a resistant ledge visible on opposite side of canyon.

116.7 Exposure of subaqueous climbing ripples in Kayenta Formation.

117.4 Cross bridge over Pine Creek. Springdale Sandstone Member of Moenave Formation forms resistant ledge just above road level on both sides of bridge. The sandstone is fluvial.

117.9 Intersection of Utah 9 and park road. Turn right on park road into Zion Canyon.

120.8 Zion Lodge. Stop for night.

DAY 3

Retrace route to intersection of park road and Utah Highway 9.

0.0 Turn right (west) on Utah 9.

0.0 ALTERNATE STOP 7C. Zion National Park visitor center. Exhibits; books and maps for sale.

0.6 Booth at west entrance to Zion National Park. Chinle Formation crops out at road level between here and Springdale.

2.5 Pass through Springdale.

5.5 Shinarump Member of Chinle Formation forms resistant ledge at road level. Member is composed of fluvial sandstone and conglomerate. Beyond here road traverses Moenkopi Formation.

23.5 Begin descending hill, passing outcrops of Kaibab Limestone of Permian age.

24.5 Pass through La Verkin, continuing southward on Utah 9. Road follows Hurricane fault for 2 miles from here to Hurricane.

33.5 Outcrops of Moenkopi Formation near axis of Virgin anticline.

35.5 Intersection of Utah 9 and Interstate 15. Turn left (south) on Interstate 15.

42.5 Take exit from Interstate 15 to Utah 18, passing through St. George and continuing northward toward Snow Canyon State Park.

49.8 ALTERNATE STOP 8A. Turn right on road into Winchester Hills subdivision. Continue 1.2 miles and park along road. View transverse cross section of large-scale trough crossbedding in upper part of Navajo Sandstone. The trough-shaped sets were produced by scour pits that migrated out of the plane of the outcrop. Dune geometry is inferred to have been similar to that shown in row 3 of Figure 7. Return to Utah 18.

49.8 Turn right (or continue northward) on Utah 18.

51.4 STOP 8. Intersection of Utah 18 and road through Snow Canyon State Park. Note unusually large scale of crossbedding in upper Navajo Sandstone east of Utah 18, and note Quaternary basalt flows. Turn left onto road through park.

53.1 STOP 9. Large-scale scalloped crossbedding in Navajo Sandstone (visible toward west). Structure is shown in Figure 11. Continue south along road through park.

54.2 Relatively thin sets of crossbeds in lower part of Navajo Sandstone.

54.5 STOP 10. Interbedded flat-bedded sandstone and crossbedded sandstone near base of Navajo Sandstone. These beds record the transition from subaqueous deposition in the Kayenta Formation to the eolian deposition in the Navajo Sandstone. Note several structures that are inferred to have been produced by the wetness of the sediment: deformational structures, crinkly lamination, and nodular appearance of bedding. Park along road and walk east to east wall of canyon. Return to vehicle and continue on road through park.

58.3 Turn left at town of Ivins.

59.8 Turn left (east) on old U.S. 91 toward St. George.

64.7 Rejoin Utah 18 at outskirts of St. George. End of road log.

ACKNOWLEDGMENTS

Ron Blakey (Northern Arizona University), Hugh McLean, and Rick Stanley (both at the USGS, Menlo Park) reviewed an earlier version of this manuscript.

REFERENCES CITED

Ahlbrandt, T. S., Andrews, Sarah, and Gwynne, D. T., 1978, Bioturbation in eolian deposits: Journal of Sedimentary Petrology, v. 48, p. 839-848.

Allen, J. R. L., 1963, Asymmetrical ripple marks and the origin of water-laid cosets of cross-strata: Liverpool and Manchester Geological Journal, v. 3, p. 187-236.

Blakey, R. C., Peterson, Fred, Caputo, M. V., Geesaman, R. C., and Voorhees, B. J., 1983, Paleogeography of Middle Jurassic continental, shoreline, and shallow marine sedimentation, southern Utah, *in* Reynolds, M. W., and Dolly, E. D., eds., Mesozoic paleogeography of west-central United States: Society of Economic Paleontologists and Mineralogists Rocky Mountain Section, Paleogeography Symposium 2, p. 77-100.

Brookfield, M. E., 1977, The origin of bounding surfaces in ancient aeolian sandstones: Sedimentology, v. 24, p. 303-332.

Doe, T. W., and Dott, R. H., Jr., 1980, Genetic significance of deformed cross bedding--with examples from the Navajo and Weber Sandstones of Utah: Journal of Sedimentary Petrology, v. 50, p. 793-812.

Fryberger, S. G., Al-Sari, A. M., and Clisham, T. J., 1983, Eolian dune, interdune, sand sheet, and siliciclastic sabkha sediments of an offshore prograding sand sheet, Dhahran area, Saudi Arabia: American Association of Petroleum Geologists Bulletin, v. 67, p. 280-312.

Fryberger, S. G., and Schenk, Christopher, 1981, Wind sedimentation tunnel experiments on the origins of aeolian strata: Sedimentology, v. 28, p. 805-821.

Harms, J. C., Southard, J. B., Spearing, D. R., and Walker, R. G., 1975, Depositional environments as interpreted from primary sedimentary structures and stratification sequences: Society of Economic Paleontologists and Mineralogists Short Course No. 2, Dallas, Texas, 161 p.

Horowitz, D. H., 1982, Geometry and origin of large-scale deformation structures in some ancient wind-blown sand deposits: Sedimentology, v. 29, p. 155-180.

Hunter, R. E., 1977, Basic types of stratification in small eolian dunes: Sedimentology, v. 24, p. 361-387.

—— 1981, Stratification styles in eolian sandstones—some Pennsylvanian to Jurassic examples from the western interior U.S.A., *in* Ethridge, F. G., and Flores, R. M., eds., Recent and ancient non-marine depositional environments—models for exploration: Society of Economic Paleontologists and Mineralogists Special Publication No. 31, p. 315-329.

—— 1985, Subaqueous sand-flow cross-strata: Journal of Sedimentary Petrology, v. 55, p. 886-894.

Hunter, R. E., and Rubin, D. M., 1983, Interpreting cyclic crossbedding, with an example from the Navajo Sandstone, *in* Brookfield, M. E., and Ahlbrandt, T. S., eds., Eolian sediments and processes: Developments in Sedimentology 38, p. 429-454.

Kocurek, Gary, and Dott, R. H., Jr., 1981, Distinctions and uses of stratification types in the interpretation of eolian sand: Journal of Sedimentary Petrology, v. 51, p. 579-595.

—— 1983, Jurassic paleogeography and paleoclimate of the central and southern Rocky Mountain region, *in* Reynolds, M. W., and Dolly, E. D., eds., Mesozoic paleogeography of west-central United States: Society of Economic Paleontologists and Mineralogists Rocky Mountain Section, p. 101-116.

Kocurek, Gary, and Fielder, Gordon, 1982, Adhesion structures: Journal of Sedimentary Petrology, v. 52, p. 1229-1241.

McKee, E. D., Douglass, J. R., and Rittenhouse, Suzanne, 1971, Deformation of lee-side laminae in eolian dunes: Geological Society of America Bulletin, v. 82, p. 359-378.

Marzolf, J. E., 1983, Changing wind and hydrologic regimes during deposition of the Navajo and Aztec Sandstones, Jurassic (?), southwestern United States, *in* Brookfield, M. E., and Ahlbrandt, T. S., eds., Eolian sediments and processes: Developments in Sedimentology 38, p. 635-660.

Middleton, L. T., and Blakey, R. C., 1983, processes and controls on the intertonguing of the Kayenta and Navajo Formations, northern Arizona; eolian-fluvial interactions, *in* Brookfield, M. E., and Ahlbrandt, T. S., eds., Eolian sediments and processes: Developments in Sedimentology 38, p. 613-634.

Peterson, Fred, and Pipiringos, G. N., 1979, Stratigraphic relations of the Navajo Sandstone to Middle Jurassic formations, southern Utah and northern Arizona: U.S. Geological Survey Professional Paper 1035-B, 43 p.

Peterson, J. A., 1972, Jurassic System, *in* Mallory, W. W., ed., Geologic atlas of the Rocky Mountain region: Rocky Mountain Association of Geologists, Denver, Colorado, p. 177-189.

Pipiringos, G. N., and O'Sullivan, R. B., 1978, Principal unconformities in Triassic and Jurassic rocks, western interior United States: U.S. Geological Survey Professional Paper 1035-A, 29 p.

Poole, F. G., 1961, Stream directions in Triassic rocks of the Colorado Plateau, *in* Geological Survey Research 1961: U.S. Geological Survey Professional Paper 424-C, p. C139-C141.

—— 1962, Wind directions in late Paleozoic to middle Mesozoic time on the Colorado Plateau, *in* Geological Survey Research 1962: U.S. Geological Survey Professional Paper 450-D, p. D147-D151.

Rubin, D. M., 1987a, Formation of scalloped cross-bedding without unsteady flows: Journal of Sedimentary Petrology, v. 57, p. 39-45.
—— 1987b, Cross-Bedding, Bedforms, and Paleocurrents: Society of Economic Paleontologists and Mineralogists, Concepts in Sedimentology and Paleontology, v. 1, Tulsa, Oklahoma, 187 p.
Rubin, D. M., and Hunter, R. E., 1982, Bedform climbing in theory and nature: Sedimentology, v. 29, p. 121-138.
—— 1983, Reconstructing bedform assemblages from compound crossbedding, *in* Brookfield, M. E., and Ahlbrandt, T. S., eds., Eolian sediments and processes: Developments in Sedimentology 38, p. 407-427.
—— 1985, Why deposits of longitudinal dunes are rarely recognized in the geologic record: Sedimentology, v. 32, p. 147-157.
Sanderson, I. D., 1974, Sedimentary structures and their environmental significance in the Navajo Sandstone, San Rafael Swell, Utah: Brigham Young University Geological Studies, v. 21, p. 215-246.
Stokes, W. L., 1964, Eolian varving in the Colorado Plateau: Journal of Sedimentary Petrology, v. 34, p. 429-432.
Tanner, W.F., 1965, Upper Jurassic paleogeography of the Four Corners region: Journal of Sedimentary Petrology, v. 35, p. 564-574.
Thompson, A. E., and Stokes, W. L., 1970, Stratigraphy of the San Rafael Group, southwest and south-central Utah: Utah Geological and Mineralogical Survey Bulletin 87, 53 p.
Vincellete, R.R., and Chittum, W.E., 1981, Exploration for oil accumulations in Entrada Sandstone, San Juan Basin, New Mexico: American Association of Petroleum Geologists Bulletin, v. 65, p. 2546-2570.