Deposition of the Tapeats Sandstone (Cambrian) in central Arizona

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ABSTRACT

Grain size, bedding thickness, dispersion of cross-stratification azimuths, and assemblages of sedimentary structures and trace fossils vary across central Arizona; they form the basis for recognizing six facies (A through F) in the Tapeats Sandstone. Five of these (A through E), present in western central Arizona, are marine deposits containing the trace fossil *Corophioides;* several intertidal environments are represented. The association of large-scale cross-bedding (50 to 300 cm) that is characterized by compound cross-stratification, numerous reactivation surfaces, and herringbone patterns is typical of facies A and generally typical of the finer-grained, thinner-bedded facies B. The sedimentary structures and polymodal distribution of foreset azimuths common to facies A and B probably formed on intertidal sand bars during emergence and late-stage tidal runoff. Facies C consists of well-sorted sandstone, gently cross stratified or with continuous parallel stratification, and foresets tangential to the lower bedding surface. This facies generally occurs where the gradient of the depositional surface increases; it apparently was deposited on a beach by shoaling waves. Facies D and, to a lesser extent, the coarser-grained facies E are sandstones with trough cross-stratification, fining-upward cycles, abundant intercalated thin shale and sandstone, rare flaser bedding, and local bipolar distribution of foreset azimuths. Both facies are tidal flat deposits; facies D was probably produced by meandering tidal channels, whereas facies E was likely produced by migration of braided tidal channels. The sixth facies (F), present in eastern central Arizona, is an arkosic small-pebble conglomerate that lacks trace fossils; low dispersion of foreset azimuths and large-scale (1 to 11-m wide) cut-and-fill structure are typical. Facies F was deposited by bedload streams that transported coarse, poorly sorted sand and gravel westward to the intertidal flats.

INTRODUCTION

This paper describes the sedimentary structures, depositional environments, and stratigraphy of the Tapeats Sandstone in central Arizona. This area lies along the southern margin of the Colorado Plateau and includes most of the headwaters and northern reaches of the Verde and East Verde Rivers (Fig. 1).

Relatively thin Cambrian sandstone deposits such as the Tapeats are the basal coarse-grained facies of a transgressive sequence that extended along north-trending strandlines throughout the Rocky Mountain states (Lochman-Balk, 1971, p. 95–103). These sandstones are generally regarded as shelf deposits that formed on the craton east of the shallow, slowly subsiding Cordilleran miogeosyncline (McKee, 1945; Stewart, 1970; Lochman-Balk, 1971, p. 92). Fluvial deposits have not been recognized in these sandstones because, as Seeland (1969) suggested, they were probably extensively although (as shown here) not entirely reworked by the transgressing seas. In eastern central Arizona, fluvial deposits were accumulating contemporaneously with marine deposits in western central Arizona. The marine sandstones are overlain by fine-grained terrigenous deposits that formed in the deepening water of the transgressing seas. Further transgression resulted in deposition of calcareous and dolomitic sediments. In the Grand Canyon of northern Arizona, the deposits of the transgressing Cambrian sea are named the Tapeats Sandstone, Bright Angel Shale, and Muav Limestone. The Grand Canyon sequence has been the topic of many stratigraphic studies (in particular, see McKee, 1945), but comparable rocks in central Arizona have largely been overlooked.

Earlier studies of Cambrian strata in central Arizona were discussed elsewhere (Hereford, 1975). These studies were primarily mapping efforts, and because the Tapeats Sandstone is devoid of fossils (with the exception of trace fossils), they emphasized establishing the age and correlation of the basal sandstone (Tapeats). It is now clear that the basal sandstone in western central Arizona is the stratigraphic equivalent of the Tapeats Sandstone in the Grand Canyon (Krieger, 1959; 1965, p. 53–57). In central Arizona, the Tapeats ranges in thickness from 0 to 80 m and rests unconformably on older Precambrian rocks. East of Juniper Mesa (Fig. 1), the Bright Angel Shale and Muav Limestone are absent, and the Tapeats Sandstone is overlain unconformably by the Chino Valley Formation that was thought to be of probable Cambrian age when defined by Hereford (1975). The results of this study indicate that the Chino Valley Formation and Tapeats Sandstone are separated by a major stratigraphic break which probably spans several periods.

This study is restricted to data obtainable through field observations, such as the thickness and orientation of cross-stratification, description of physical and biogenic sedimentary structures, and visual estimates of grain size and mineral composition. The Tapeats in central Arizona is well suited to such an approach because of the relatively large scale of sedimentary structures and the coarse grain size of the sandstone. Furthermore, exposures are fairly continuous, although gaps occur, and most outcrops are accessible (at least at the scale of the bedding) to observation in three dimensions.

LITHOLOGY AND PHYSICAL SEDIMENTARY STRUCTURES

Lithologically, the Tapeats Sandstone is generally a very coarse-grained sandstone to granule conglomerate in which the composition ranges from quartzarenite to arkose. Grains are subrounded to rounded. Regardless of the degree of rounding, the absence of interstitial clay warrants a mature textural classification. Throughout central Arizona, the formation is predominantly a coarse-grained deposit; few shale or siltstone beds exceed 50 cm in thickness and, where present, are subordinate to sandstone. Cross-stratification is by far the most conspicuous type of sedimentary structure. In places, particularly near the center of the area, the bedding and lithology of the Tapeats closely resembles those described from modern braided streams, but trace fossils that are quite likely marine occur at many stratigraphic horizons.

Cross-stratification of the planar (both tabular and wedge-shaped) and trough varieties constitutes more than 95 percent of the bedding in the Tapeats Sandstone. Both varieties occur in relatively thick sets, although the planar type is consistently

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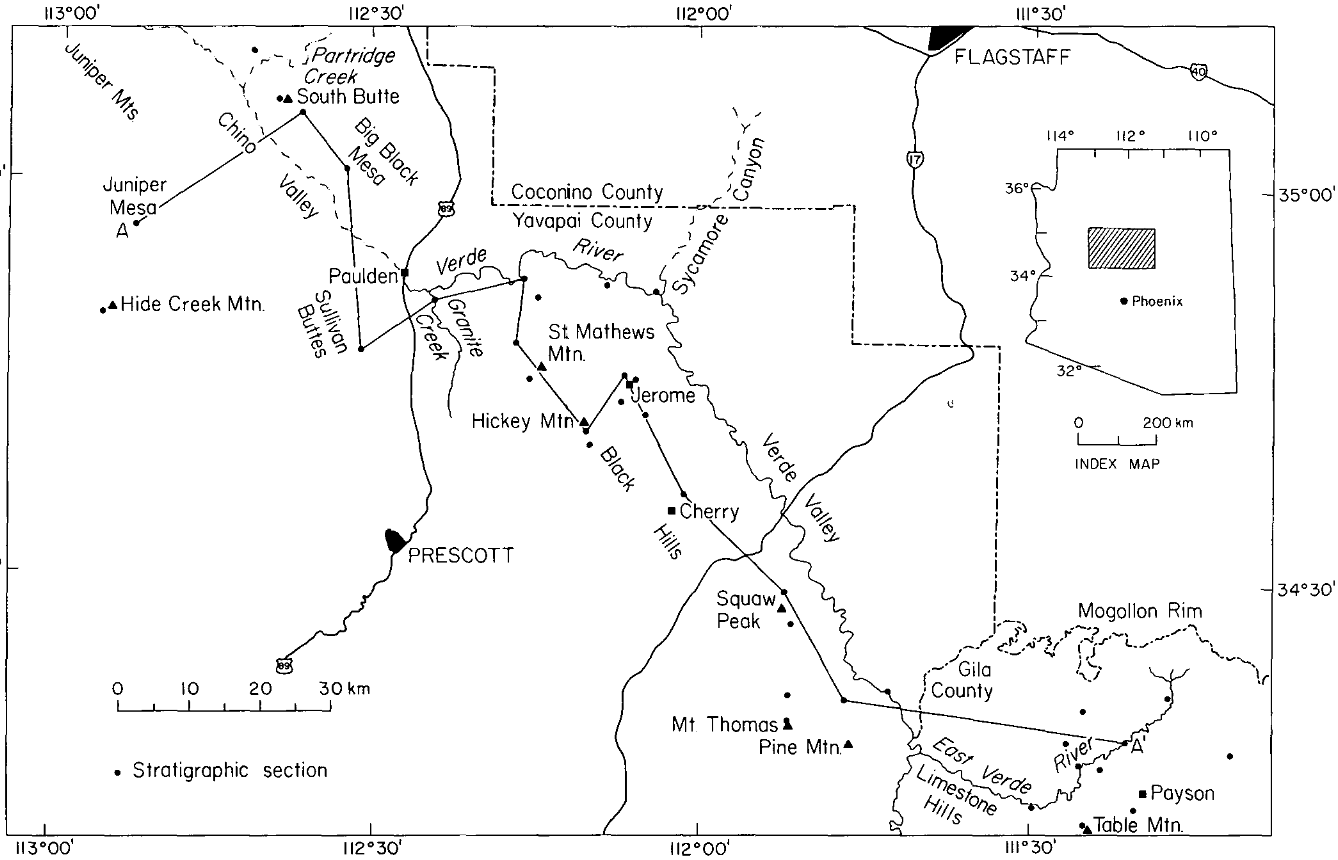

Figure 1. Central Arizona. The Tapeats Sandstone crops out along both sides of Chino Valley, in the Verde River canyon between Granite Creek and Sycamore Canyon, in the Black Hills, and along the base of the Mogollon Rim and the lower reaches of the East Verde River in the Payson area. A–A′ is the line of section for Figure 10.

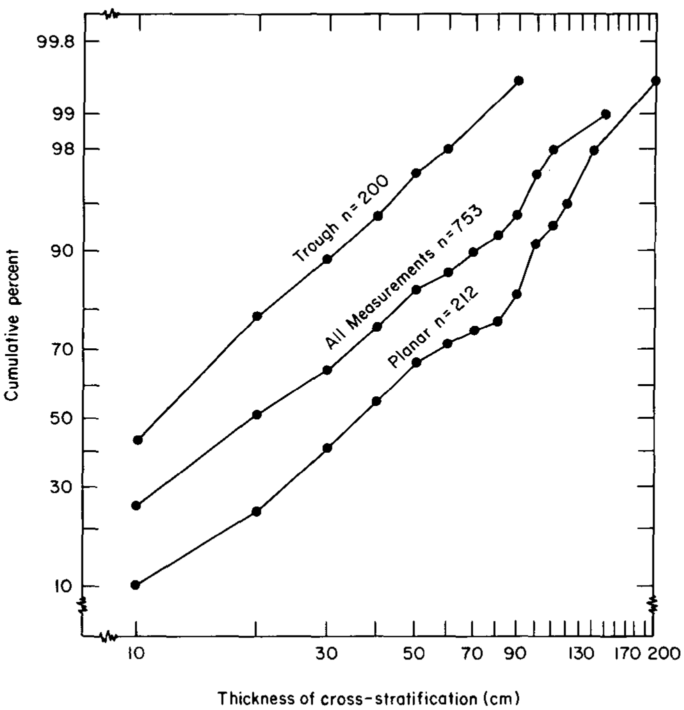

Figure 2A. Log probability plot of cross-stratification thickness.

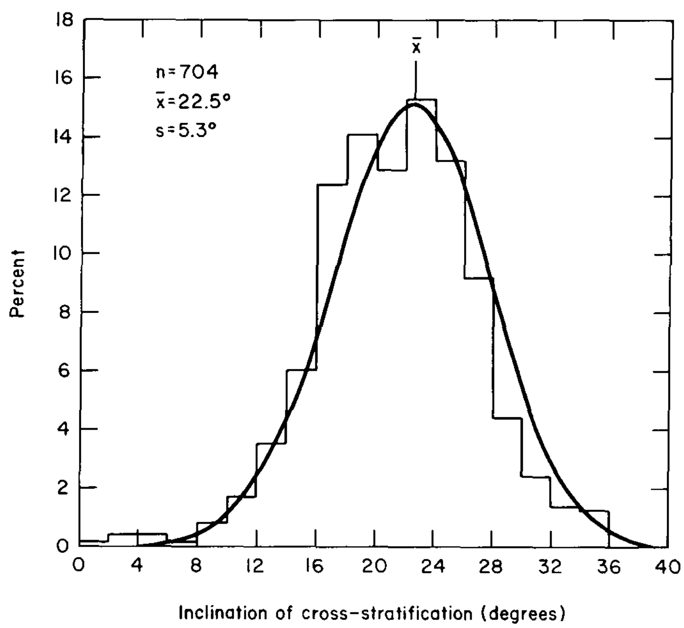

Figure 2B. Frequency histogram of cross-stratification inclination. The curve shows the fit between the normal distribution and the data.

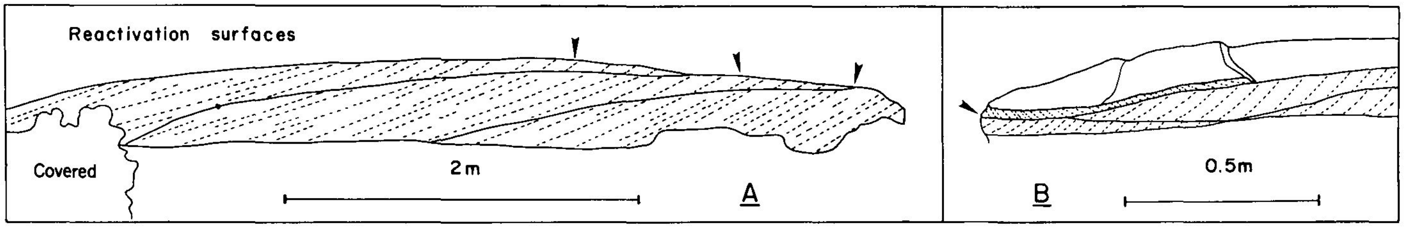

Figure 3. A. Three slightly convex reactivation surfaces (arrows) near South Butte. B. Perspective view of a reactivation surface (arrow) overlain by a thinly cross-stratified bed with reverse dip. Figures 3A and 3B were both traced from a photograph.

thicker. The sample distribution of bedding thickness and the inclination of cross-stratification are shown in Figure 2. Bedding thickness generally follows a lognormal distribution. The planar variety is bimodally distributed, having a secondary mode in the 80- to 110-cm interval (Fig. 2A). The maximum angle of foreset inclination is normally distributed with a minor mode in the lower angles (Fig. 2B). Beds are generally bounded by sharp contacts and ripple forms are poorly preserved. Foresets that alternate between poorly sorted granule conglomerate and very coarse sandstone are characteristic. Foresets range in thickness from less than 1 to about 10 cm.

Distinctive types or sequences of cross-stratification that reflect important characteristics of the transporting system are common in the Chino Valley area. Thick sets of cross-stratification are commonly cut diagonally (Fig. 3) by two or as many as ten planar or curved surfaces (Fig. 3A). Similar crosscutting surfaces are found in modern fluvial and marine deposits (Boersma, 1969; Collinson, 1970; Klein, 1970a) and have been called reactivation surfaces. These surfaces develop by the partial erosion of the avalanche faces of a bedform that has ceased to migrate. Numerous, closely spaced reactivation surfaces, such as those in Figure 3A, suggest alternating periods of erosion and deposition resulting from a periodically reversing transport system. In some places, the current that produced the reactivation surface also deposited a thin veneer of sediment on the surface that is commonly cross-stratified with foreset directions opposed to those beneath the reactivation surface (Fig. 3B). In a few examples, the deposits with opposed foreset azimuths were covered by renewed migration of the bedform. In one example, three reactivation surfaces and associated deposits with reversed dip were spaced at 1.7 and 2 m from each other within a single, otherwise laterally continuous cross-stratified set. These figures, assuming deposition under tidal conditions, provide a measure of the rate of bedform migration.

Large-scale sets of planar cross-bedding with thick foresets that are cross-stratified are common in Chino Valley. The foresets are cross-stratified at right angles or obliquely up the steep slip-face of the thick planar unit (Fig. 4). Every foreset is cross-stratified for its entire length. These beds can be traced for as much as 500 m and retain the same characteristics throughout. Williams (1971, p. 30) discussed a similar type of cross-stratification from ephemeral streams, but the internal organization of the foresets he described shows none of the remarkable repetition or continuity of those in the Tapeats Sandstone. Collinson (1968, p. 241, 247–250) discussed cross-stratified foresets from ancient deltaic deposits and proposed that they be termed "intrasets." These differ from the cross-stratified foresets in the Tapeats in that the entire foreset is not affected and the cross-strata tend to dip down the foreset. A type of large-scale cross-bedding with cross-stratified foresets, similar to those in the Tapeats, was described by Harms and others (1975, p. 51–57) who proposed the term "compound cross-stratification." Stratification comparable to the compound cross-stratification in the Tapeats is common on modern intertidal sand bars where it is produced during emergence by late-stage tidal runoff (Klein, 1970a, p. 1104). Compound cross-stratification has also been observed in ancient tidal sand bodies (Klein, 1970b; Swett and others, 1971). Beds with compound cross-stratification are the thickest found in the Tapeats; thicknesses between 1 and 3 m are not uncommon, with foresets as thick as 10 cm. If not planed by erosion, these beds, as well as many planar cross-stratified sets, are overlain by a relatively thin coset of trough cross-strata that commonly are oriented at right angles or opposite to the underlying set. Herringbone patterns are fairly common in Chino Valley and are likely produced by reversing currents.

ORIENTATION AND DISPERSION OF CROSS-STRATIFICATION

The relevant statistical parameters of more than 800 orientation measurements are summarized in Table 1. The localities are arranged (Table 1) as they occur from west to east. The measures of dispersion (L and s columns) indicate a pattern of decreasing variability down the respective

Figure 4. Compound cross-stratification in a large-scale planar bed. The foresets in the right half of the photograph dip to the left at 24°. In the left half of the photograph, the foresets are exposed along strike, showing the internal cross-stratification. Scale = 15 cm in length.

TABLE 1. ORIENTATION STATISTICS OF CROSS-STRATIFICATION IN THE TAPEATS SANDSTONE

	Locality	n	$\overline{\theta}$* (°)	L† (%)	s§ (°)	s^2	X^2p (d.f.)# (%)	R.T.S.**	Totals n††	Totals $\overline{\theta}$ (°)	Totals L (%)	Totals s (°)	Totals s^2	Totals X^2p (d.f.) (%)	Totals R.T.S.
I	1. Partridge Creek	50	230	35.1	82.9	6874	97.8 (9)	2×10^{-3}	307	245	33.1	85.20	7259	96.6 (9)	3×10^{-15}
	2. South Butte	88	255	32.9	85.4	7299	86.8 (9)	7×10^{-5}							
	3. Big Black Mesa	41	270	38.9	78.7	6199	41.4 (9)	2×10^{-3}							
	4. Sullivan Buttes	63	253	27.2	92.5	8548	99.3 (9)	9×10^{-3}							
II	5. Verde River	38	280	43.0	74.4	5541	94.7 (7)	9×10^{-4}	240	247	40.8	76.7	5886	83.6 (9)	4×10^{-18}
	6. St. Mathews Mountain	47	253	43.3	74.1	5496	86.7 (7)	1×10^{-4}							
	7. Hickey Mountain	47	177	46.9	70.5	4971	94.8 (9)	3×10^{-5}							
	8. Jerome	76	264	44.1	73.3	5375	36.6 (9)	4×10^{-7}							
III	9. Sycamore Canyon	41	232	62.0	56.0	3139	96.6 (5)	1×10^{-7}	296	256	51.4	66.1	4370	37.4 (9)	1×10^{-34}
	10. Cherry	35	311	50.1	67.4	4538	75.4 (6)	2×10^{-4}							
	11. Squaw Peak	37	293	44.3	73.1	5346	76.8 (4)	7×10^{-4}							
	12. Mt. Thomas	80	253	58.8	59.0	3487	55.8 (6)	1×10^{-12}							
	13. East Verde River	84	218	72.4	46.0	2120	93.8 (4)	8×10^{-20}							

* Average direction computed by the vector summation method described by Curray (1956, p. 118).

† Magnitude of the resultant vector, a measure of dispersion.

§ Circular standard deviation (Mardia, 1972, p. 24):

$$s = \frac{180}{\pi}\left(-2 \log_e \frac{L}{100}\right)^{\frac{1}{2}}.$$

A test of the goodness of fit with the circular normal distribution. Expected frequencies were computed by numerical integration. Deviations from the expected values are approximately distributed as χ^2 with $(k-3)$ degrees of freedom (Mardia, 1972, p. 123). Probabilities greater than 90 percent indicate that the data do not fit the circular normal distribution.

** Rayleigh test of significance (Curray, 1956, p. 125). Values larger than 5×10^{-2} are taken to mean that the data are randomly distributed on the interval 0°–360°.

†† Includes scattered measurements that could not be assigned to the localities on the left.

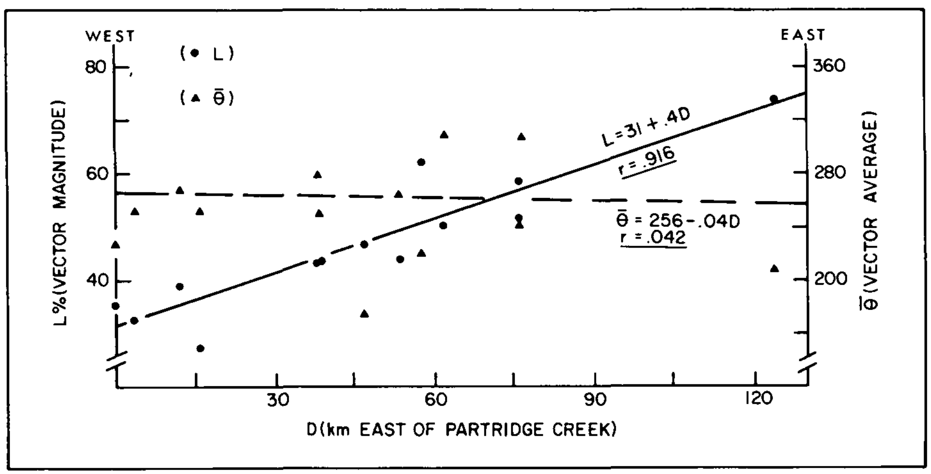

Figure 5. Vector magnitude (L, left ordinate) and vector average ($\overline{\theta}$, right ordinate) of cross-stratification azimuths plotted against the distance east of Partridge Creek.

columns; in contrast, the measure of location or average direction ($\bar{\Theta}$ column) shows no significant change down its column. The values of L and s fall into three groups as shown by I, II, and III in Table 1. The grouping implies that the measures of dispersion (and location) were derived from separate populations. An easily computed nonparametric test (Hollander and Wolfe, 1973, p. 114–117) was used to test the hypothesis of separate populations against the alternative that the measures of dispersion and location can be considered as a combined sample from two populations (one population for each measure). Taking either L or s as the appropriate measure of dispersion yields nearly identical results — there is virtually no possibility that the samples were derived from the same population. The measure of location, however, can be considered (with probability of 99.7 percent) as a combined sample from one population.

A rigorous mathematical approach requires that each of the three populations of dispersion be treated separately. Figure 5, however, shows that a linear function conveniently describes the relation between distance measured in an easterly direction and dispersion; as the data in Table 1 suggest, dispersion decreases in an easterly direction. Moreover, there appears to be no relation between distance and average direction (Fig. 5). Throughout central Arizona, the average direction of sediment transport was west-southwest, but the dispersion about this direction can be expected to decrease in an easterly direction.

The frequency distributions of foreset azimuths are shown on Figure 6. Six of these are not satisfactorily described by the circular normal distribution (CND; column 6 in Table 1 and Fig. 6); the distributions, however, are not random, and an average direction exists (Table 1, R.T.S. column). Deviations from the normal distribution (assuming no sampling bias) are caused by an excessive number of observations in one or more class intervals. At localities 1 and 2, the modes occur between 90° and 120° on either side of the mean direction. Modes at 180° to the dominant direction are less common (loc. 5, Fig. 6 and Table 1). The foregoing discussion applies to the area west of Jerome; east of Jerome, the distributions that differ from the CND are characterized by one mode at right angles or a mode within one class interval of the average direction (Table 1; Fig. 6).

Frequency distributions with modes at right angles to the prevailing direction (best developed in groups I and II, Table 1) are similar to distributions shown by Bluck (1974, p. 540) from a modern braided river; such polymodal distributions with high directional variability might be regarded as common to this type of fluvial

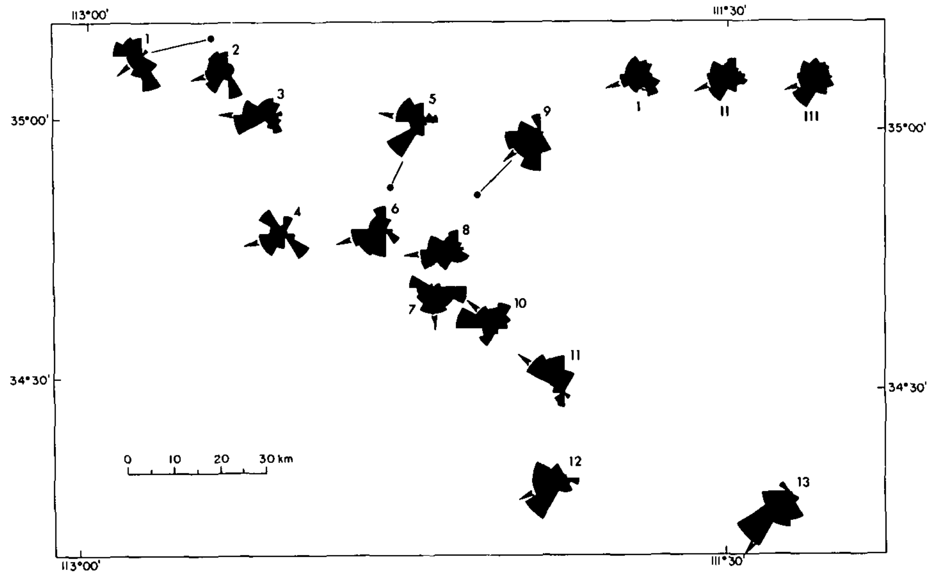

Figure 6. Frequency distributions of cross-stratification azimuths. The statistical parameters at each of the 13 numbered localities are given in Table 1. The frequency distributions in the upper right are the totals from localities 1–4, 5–8, and 9–13, respectively (see Table 1). Width of class interval is 30°.

deposit (Ore, 1964; Williams, 1971; Smith, 1972). Other studies of modern and Pleistocene braided rivers, however, have yielded low variance unimodal distribution (Williams and Rust, 1969; McDonald and Banerjee, 1971; Costello and Walker, 1972). The conditions favoring preservation of directional elements deviating significantly from the downslope direction are poorly understood (Allen, 1967; Banks and Collinson, 1973; Smith, 1973). This further complicates the problem of interpreting the shape of frequency distributions and the variability of directional elements in fluvial as well as marine systems. Apparently, Allen's (1967, p. 78) contention that straight channels should preserve only those structures that closely coincide with the channel is not yet refuted by studies of modern fluvial systems, and unimodal, low-variance distributions could result when the preserved directional structures of braided stream deposits are sampled.

The variability of cross-stratification azimuths of groups I and II (Table 1) lies within the range commonly found in ancient marine rocks, whereas the variability of group III lies within the range of ancient fluvial rocks (Potter and Pettijohn, 1963, p. 89). Polymodal distributions of the type in groups I and II could be expected to develop on intertidal sand bars by deflection of tidal currents around irregularities on the depositional surface (Klein, 1970a).

BIOGENIC SEDIMENTARY STRUCTURES

The trace fossil *Corophioides* (Häntzschel, 1962, p. W189) is found at most outcrops of the Tapeats Sandstone south of the east Verde River and northwest of Mount Thomas. Figure 7 shows the occurrence of trace fossils.

Rarely is *Corophioides* exposed in vertical section (Fig. 8A); more commonly, it occurs on bedding planes as the eroded end of the U-shaped tube (Fig. 8B) which has been referred to *Arenicoloides* (Häntzschel, 1962). For practical purposes, the two trace fossils can be considered as one that hereafter will be referred to as *Corophioides*. *Corophioides* and other trace fossils occur through a wide range of grain sizes not uncommonly as coarse as granule or small-pebble conglomerate. In general, the size of the burrow varies with grain size; the largest burrows occur in the coarsest deposits.

Corophioides has morphologic affinities with U-shaped burrows constructed by several types of polycheates living on modern tidal flats (Smith, 1893; Bather, 1925; Richter, 1926; Ager and Wallace, 1970). Furthermore, *Corophioides* has generally been recognized as an indicator of marine conditions, especially an intertidal high energy environment, or has been found in rocks known by independent criteria to be of marine origin (Richter, 1926; MacKenzie, 1968; Hecker, 1970; Heinberg, 1970; Sellwood, 1970; Land, 1972). The presence of *Corophioides* seems to be a reliable indicator of marine or intertidal deposition. Trails and horizontal burrows are not common, possibly because of unfavorable exposure or removal by erosion. Alpert (1974, p. 144) illustrated trails similar to those in the Tapeats from Cambrian rocks in California which he concludes resemble

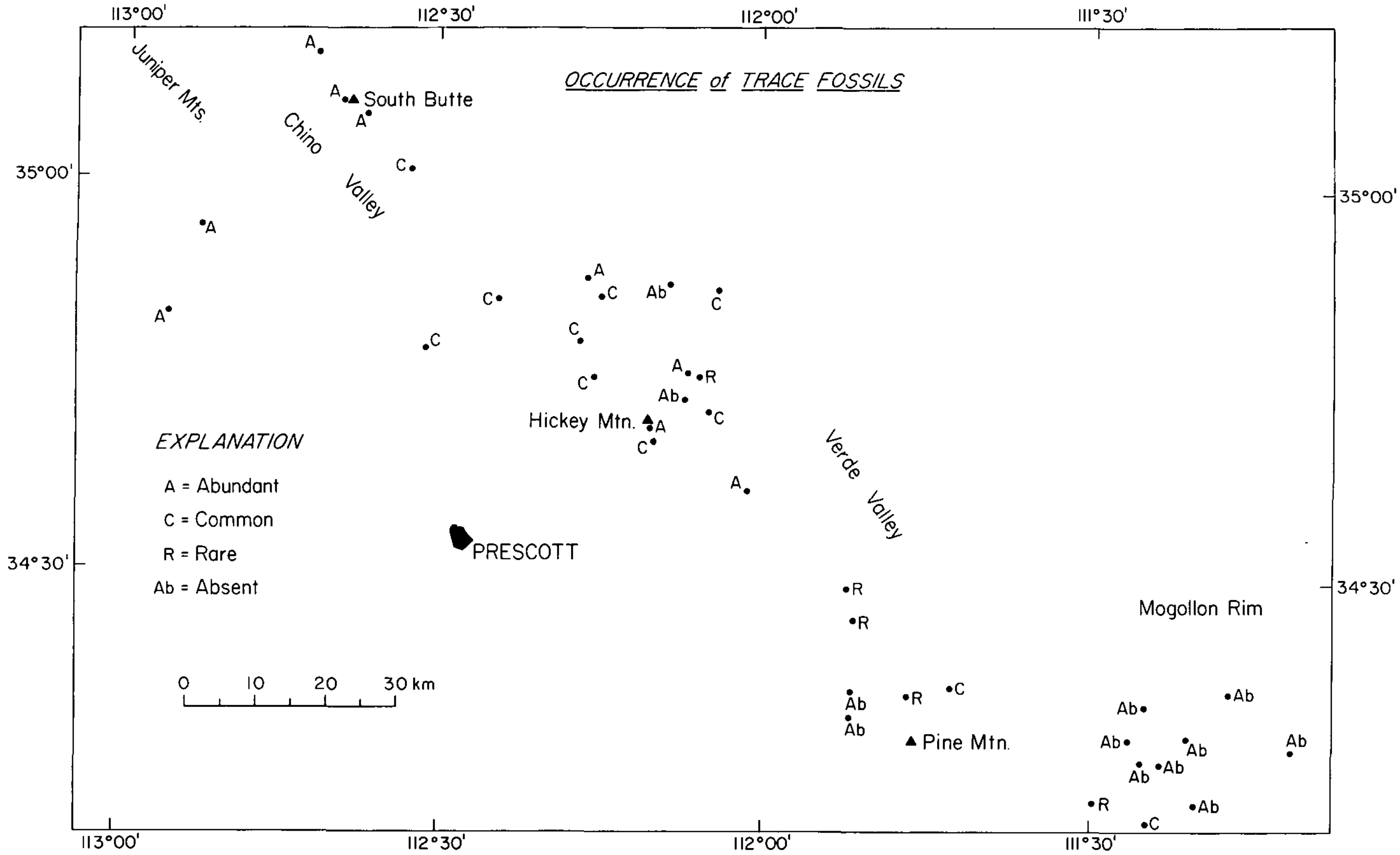

Figure 7. Occurrence of trace fossils in central Arizona. Abundant = fossils found on every exposed bedding plane; common = most beds contain fossils; rare = fossils found with difficulty; absent = no fossils found regardless of exposure.

trails produced by gastropods living on modern tidal flats.

STRATIGRAPHY

Lithofacies

Variations in the types of bedding, grain size, bedding thickness, occurrence of trace fossils, and topographic expression occur in a systematic pattern and provide the basis for recognizing six facies (A to F) within the Tapeats Sandstone. The distinguishing characteristics of the six facies and the location of typical exposures are given in Table 2.

Facies A and B are present in Chino Valley as far east as Hickey Mountain (Fig. 1). Both facies are characterized by reactivation surfaces (Figs. 3a, 3b), compound cross-stratification (Fig. 4), wide dispersion of foreset azimuths, and fairly common occurrence of herringbone pattern. These similarities suggest that the two facies are of similar origin. The scale of cross-stratification in facies B is significantly smaller and the grain size is generally finer than facies A. Facies B and, to a lesser extent, facies A become coarser grained in an easterly direction, and they are difficult to distinguish by grain-size criteria at Sullivan Buttes and Granite Creek (Fig. 1). However, the slope-forming character (Table 2) and thinner bedding of facies B persist and distinguish it from facies A. The contact between the two facies is gradational, and each facies contains elements typical of the other.

The bedding characteristics common to facies A and B occur on modern intertidal sand bars (Boersma, 1969; Klein, 1970a), and similar types of bedding have been reported from ancient rocks thought to be largely of intertidal origin (Klein, 1970b; Swett and others, 1971; de Raaf and Boersma, 1971; Banks, 1973a). The abundance of trace fossils and the high dispersion of cross-stratification azimuths (Table 1, group I; Table 2) associated with both facies indicate marine deposition. In a previously published paper (Hereford, 1974), I suggested that the Tapeats in the Chino Valley area was largely a subtidal deposit; however, the abundant compound cross-stratification and polymodal distribution of foreset azimuths (Fig. 6) probably formed in an intertidal rather than subtidal setting during late-stage tidal runoff and emergence (Klein, 1971). Furthermore, *Corophioides* and the other trace fossils common in both facies are, as previously indicated, generally restricted to intertidal environments.

Facies C, locally present as far east as Cherry, stands apart from the other facies because of its relatively fine grain size (medium- to coarse-grained sandstone compared with very coarse grained sandstone to granule conglomerate) and uniform sorting. Where the facies is well exposed and traceable for short distances, such as the sections at St. Mathews and Hickey Mountains, northwest of Cherry, and west of Big Black Mesa, it thins against topographic highs on the Precambrian surface. Apparently, deposition of the facies is related to the local relief on the depositional surface. The sedimentary structures in facies C (Table 2) are common in modern beach environments (Thompson, 1937; McKee, 1957; van Straaten, 1959; Reineck, 1967; Hayes and others, 1969); thus this facies probably formed on beaches that developed around local irregularities on the bedrock surface. Strata identical to this facies in sedimentary structures and occurrence were mapped by McKee (1945, p. 130) in the eastern Grand Canyon National Park.

Facies D, present from eastern Chino

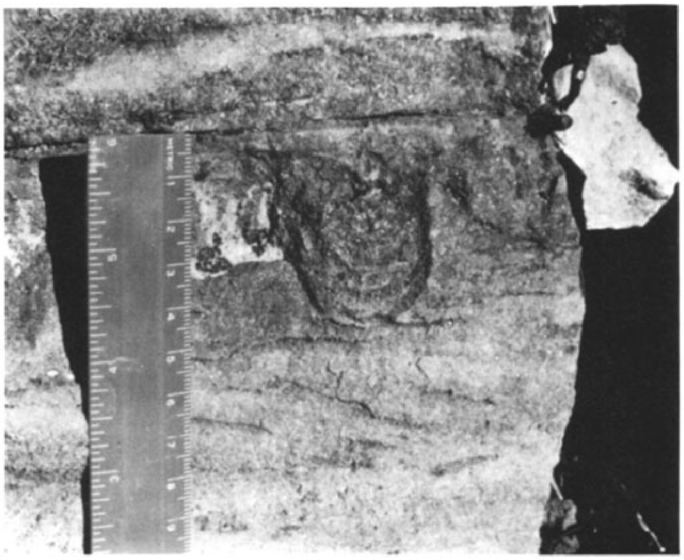

Figure 8A. Vertical section through the U-shaped tube of the trace fossil *Corophioides* (scale in centimetres) in facies D near Hickey Mountain.

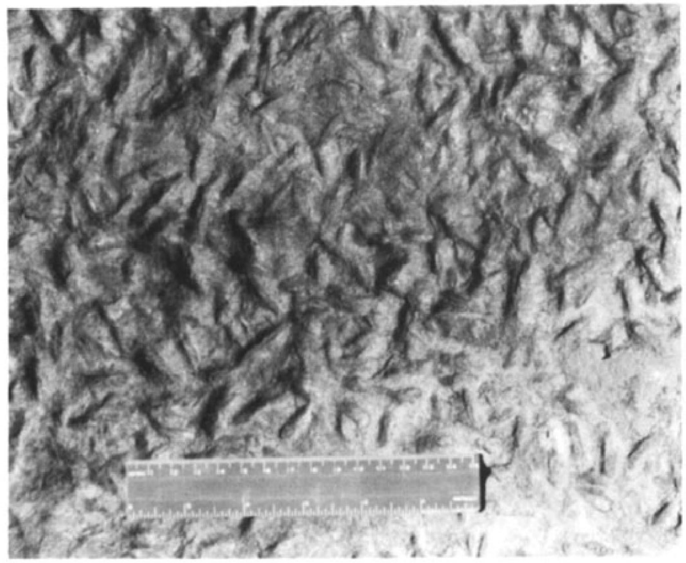

Figure 8B. Bedding plane with *Corophioides* eroded to the bottom end of the U-Shaped tube (scale in centimetres) in facies A near South Butte. This is the most common occurrence of *Corophioides*.

Valley as far southeast as the Limestone Hills and Table Mountain areas (Fig. 1), is a coarse-grained sandstone to granule conglomerate characterized by small-scale trough cross-stratification filling erosional scours and numerous thin lenticular sandy shale and coarse siltstone partings. Cosets of trough cross-strata and associated shale and siltstone partings occur in beds that range in thickness from 50 to 150 cm. The sandstone beds are typically underlain by fragments of detrital shale and siltstone that also occur at numerous places within the beds. Thin layers of granule conglomerate line the base of many of the erosional scours. The tops of the sandstone beds are densely covered with *Corophioides* burrows and are overlain by as much as 1 m of sandy shale or coarse siltstone which is continuous across the outcrop. These fine-grained deposits are interbedded, particularly in the lower parts, with thin coarse-grained sandstone that is highly burrowed. In favorably oriented exposures, reversals in foreset azimuths are common. Crudely developed fining-upward cycles, ranging in thickness from 30 to 60 cm, are locally present. These begin with a massive granule conglomerate that grades into cross-stratified very coarse-grained sandstone; this grades abruptly into sandy shale or siltstone. Outcrops of this facies are characterized by alternating beds of light-colored sandstone and pale red shale or siltstone.

Coarse detrital fragments and granule conglomerate beneath sets of cross strata probably were lag concentrates deposited within a channel. The stacked arrangement of erosional scours and lag concentrates along with thin drapes of shale or siltstone implies that deposition within the channels was frequently interrupted. These features and the presence of *Corophioides* at the top of the channel sequence indicate a tidal channel. Fining-upward cycles, well known from meandering rivers (Allen, 1965), also occur in meandering tidal streams (van Straaten, 1954; Klein, 1967), suggesting that these tidal channels were, at least locally, meandering. Shale and coarse-grained siltstone overlying the channel sequences were probably deposited from suspension during periods of slack water. However, the intercalated sandstone suggests that deposition outside the channels was not entirely from suspension. Detrital shale and siltstone fragments within the channel deposits were probably derived by undercutting and erosion of the fine-grained sediment by the meandering tidal channels. Thus, facies D was deposited primarily within tidal channels with subordinate preservation of fine-grained tidal flat deposits. Johnson (1975, p. 51–65) described tidal-channel deposits from the younger Precambrian of Norway that are remarkably similar to facies D. In addition, I have examined the bedding sequences described by Stewart (1970, p. 45–47, Figs. 19–21; also p. 13) as typical of the Middle member of the Wood Canyon Formation, which is the stratigraphic equivalent of the Tapeats in eastern California, and they are nearly identical to facies D. Preliminary studies in the eastern Grand Canyon National Park show that the basal 20 m of the Tapeats on both sides of Bright Angel Creek is identical to facies D described in central Arizona.

Facies E is present in the Black Hills as far south as Squaw Peak. The facies is typically a very coarse-grained sandstone to small pebble conglomerate. Discontinuous layers of granule and small pebble conglomerate, generally less than 20 cm thick, commonly underlie sets of planar cross-strata, the dominant type of cross-stratification. Bedding sequences similar to those in facies D occur locally, but the two facies are readily distinguished because facies E is generally much coarser-grained than facies D. Sets of cross-strata frequently fill wide, essentially planar erosional scours which accounts for the common occurrence of lenticular bedding. In favorably oriented exposures, the apparent angle of foreset inclination decreases along the outcrop. Smith (1972, p. 627) described a similar decrease for transverse bars in a modern braided river. The conglomerates, where not overlain by a cross-stratified set, may grade upward into sandy, micaceous siltstone or shale. Similar sequences, although of larger scale, are common in cut-off channels of modern braided streams (Doeglas, 1962, p. 173). Abandoned channels are also common on modern tidal flats (van Straaten, 1954, p. 7) and could produce such sequences (Klein and Sanders, 1964, p. 22).

Of very limited occurrence in facies E are type A ripple drift cross-lamination and sinusoidal ripple laminae (Jopling and Walker, 1968). These structures are generally restricted to fluvial or turbidite sequences, but Chanda and Bhattacharyya (1974) showed that ripple drift cross-lamination and presumably sinusoidal ripple laminae can form under tidal conditions. Many of the criteria, however, that distinguish ripple drift cross-lamination formed in a tidal setting from similar structures produced in fluvial or turbidite sequences (Chanda and Bhattacharyya, 1974,

TABLE 2. DISTINGUISHING FEATURES OF THE SIX FACIES OF THE TAPEATS SANDSTONE IN CENTRAL ARIZONA

Facies	Bedding characteristics	Bedding thickness (cm) $\bar{x}$	s	n	Dispersion of cross-stratification azimuths: Average vector magnitude (L%)	Range of values (%)	Topographic expression	Trace fossils	Interpretation	Lithology	Typical exposures
A	Compound cross-stratification, reactivation surfaces, herringbone cross-stratification	45.0	45.3	164	36.7	27–48	Ledge	Abundant	Low tidal flat sandbar	Coarse sandstone to granule conglomerate, moderately to well sorted quartzarenite to subarkose, color variable from white to pale red	South Butte area; secs. 22, 26, 36 T. 20 N., R. 4 W. Picacho Butte 15′ quad. (1947)
B	Similar to A except for abundant trough cross-stratification	22.0	22.8	82	52.9	43–79	Slope	Abundant	Midtidal flat sandbar	Coarse to very coarse sandstone, moderately to well sorted, color and composition similar to facies A	South Butte area
C	Low (2°–10°) to medium (10°–20°) cross-stratification, trough cross-stratification, tangential foresets, no alternation of coarse-fine foresets, continuous parallel stratification	20.0	13.5	12	Insufficient data		Slope	Rare to absent	Beach with ridges	Medium to coarse sandstone, well sorted, pale red	South slope of Hickey Mtn., 1.1 km north of BM 5771, Mingus Mtn. 15′ quad. (1944)

p. 1120–1121) are absent in facies E. Despite this, a fluvial origin is not suggested for the entire facies, because it contains *Corophioides* and is characterized by wide dispersion of foreset azimuths (Table 2). Ripple drift cross laminae and sinusoidal ripple laminae form during ripple migration only when a very large amount of sand is available for deposition (McKee, 1966). In facies E, beds containing these structures probably formed when the shoreward margins of the tidal flats were inundated by sediment-laden floodwaters.

Very coarse-grained deposits, lenticular conglomerates, absence of thick beds of shale and siltstone, abundant planar cross-stratification, and lenticular bedding, all characteristic of facies E, are also typical of modern and ancient braided stream deposits (Doeglas, 1962; Ore, 1964; Allen, 1965; Smith, 1970; Williams, 1971; Costello and Walker, 1972; Banks, 1973b). Braided stream deposits seem to be a reasonable depositional analog of facies E, but this interpretation does not consider the occurrence of trace fossils in this facies. Braided stream patterns commonly occur on modern tidal flats where the sediments are predominantly sand sized (van Straaten, 1954; Klein, 1963; Swift and McMullen, 1968). Apparently, the sedimentary structures produced on tidal flats drained by braided streams have not been thoroughly studied, but Klein (1967, p. 210) reports that lenticular cross-stratification is common. The sedimentary structures developed by meandering tidal creeks and those developed in meandering rivers are similar in many ways (van Straaten, 1954; Klein, 1965; Land and Hoyt, 1966; Tucker, 1973). This similarity implies that the internal structures of tidal flats drained by braided channels may be similar to those produced by braided rivers. Facies E probably represents the deposits of a tidal flat drained by braided channels that at times was inundated by sediment derived from a nearby terrestrial environment. Consequently this facies represents a complex mixture of marine with nonmarine strata.

Facies F is present north of the East Verde River, in the Pine Mountain and Payson areas, and perhaps at Sycamore Canyon (Fig. 1), where no trace fossils were found in the basal part of the section. It superficially resembles facies E but differs significantly in detail in that cut-and-fill structure is common, trace fossils are absent, grain size is coarser, and the facies lies within the area characterized by low variance of cross-stratification azimuths (Fig. 5; Table 1, group III). Facies F is typically a very coarse grained sandstone to small pebble conglomerate of subarkosic to arkosic composition. Shale and siltstone form thin lenticular beds. Facies E is dominated by

TABLE 2. *(Continued)*

Facies	Bedding characteristics	Bedding thickness (cm) $\bar{x}$	s	n	Dispersion of cross-stratification azimuths: Average vector magnitude (L%)	Range of values (%)	Topographic expression	Trace fossils	Interpretation	Lithology	Typical exposures
D	Alternating sandstone and shale, local fining-upward cycles, numerous thin shales, trough cross-stratification, occasional ripple forms	11.6	6.5	18	38.5	..	Slope	Abundant	High tidal flat drained by meandering channels	Coarse sandstone to granule conglomerate, sorting variable, conglomerate at base of sandstone; interbedded with micaceous shale, abundant silt and sand near base; sandstone 10–75 cm, shale 10–100 cm thick	Northeast slope of Woodchute Mtn. in quarry near spring 0.6 km south of BM 5677, Clarkdale 15′ quad. (1944)
E	Discontinuous bedding, numerous lenticular conglomerates, abundant thin shales	24.0	18.4	95	47.8	44–52	Weak ledge	Rare to common	High tidal flat sand bars drained by braided channels	Coarse sandstone to small-pebble conglomerate, moderately sorted, subarkose	Northeast of Jerome near abandoned railway 0.8 km southeast of Hopewell Tunnel Clarkdale 15′ quad. (1944)
F	Abundant cut-and-fill structure, discontinuous bedding, numerous conglomerates, thin lenticular shales	35.0	23.3	102	72.8	59–84	Ledge	Absent	Prevegetation bedload streams	Granule to small-pebble conglomerate, poorly to moderately sorted, subarkose to arkose	Head of Limestone Canyon, SE¼NW¼ sec. 16 T. 11 N., R. 9 E., Pine 15′ quad. (1952)

large-scale cut-and-fill structure that truncates adjacent beds. The cuts, commonly as deep as 1 m, have steep walls and are as much as 11 m wide. They are filled with cross-strata that conform to the walls of the cut. The bases of the cuts are typically lined with conglomerate having rounded clasts up to 5 cm in diameter. Shale and coarse siltstone in discontinuous beds, generally less than 50 cm thick, commonly occur at the top of the cross-strata filling the cuts. Figure 9 illustrates the stratification typical of facies F. A striking feature of this facies in thick vertical sections is the stacked arrangement of conglomerate and recessed differentially weathered shale and siltstone beds.

The basal contact of this facies with Precambrian rocks, unlike the other facies, is locally gradational through 1 or 2 m, and it is commonly difficult, despite contrasting rock types, to accurately place the contact. Titley (1962, p. 124) also described a gradational contact between the Tapeats Sandstone and the underlying Precambrian granite. The gradational interval apparently represents partially reworked regolith.

The sedimentary structures and stratification sequences characteristic of facies F are similar to those previously described from modern braided streams. However, prevegetation fluvial deposits such as facies F are thought to differ in many subtle but important ways from their modern or postvegetation counterparts (Schumm, 1968). For instance, bed-load channels with wide braided streams, subject to large variation in discharge, probably were common in prevegetation fluvial systems. These streams spread coarse, poorly sorted sediment as sheets, forming vast alluvial plains and preserving erosion surfaces and regolith (Schumm, 1968, Wanless, 1973).

A seventh facies (conglomerate facies) restricted to a small area south and southeast of Jerome consists of massive small pebble to pebble conglomerate filling relatively narrow channels cut on the Precambrian surface. The channels have a width to depth ratio of 10:1 or more and are filled with rounded to subrounded clasts similar in composition to the underlying Precambrian rocks. In places, this facies grades upward into facies C. From the orientation and fill of the channels, it seems likely that they drained a landmass of unknown extent that lay in the vicinity of Hickey Mountain (Fig. 1).

Stratigraphic Relations of the Six Facies and a Depositional Model

The stratigraphic relations of the six facies are shown on Figure 10. Clearly displayed is the beveling of the Cambrian section by the Chino Valley Formation.

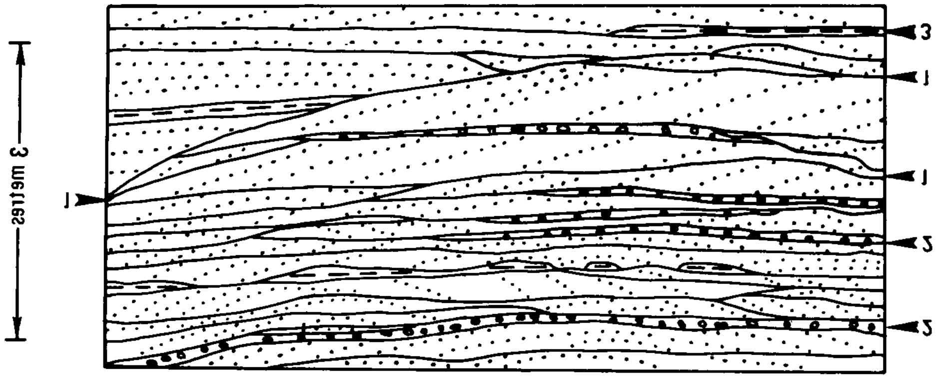

Figure 9. Vertical exposure of facies F north of East Verde River showing typical cut-and-fill structure. Direction of flow is toward observer. 1 = cross-stratified sandstone and conglomerate filling channels cut into underlying beds. 2 = small pebble conglomerate lining base of channel (lag deposition); 3 = coarse siltstone and sandy shale filling erosional scour (abandoned channel).

McNair (1951) first noted the easterly truncation by Devonian strata of the Cambrian section across northwestern Arizona; however, he concluded that the section was completely removed east of Sullivan Buttes. No distinctive or traceable strata had been recognized in the Tapeats Sandstone, and the extent, if any, to which the formation was eroded in central Arizona was not previously known. Erosion appears to have progressively truncated, in a southeasterly direction, the facies sequence A-B-A in Chino Valley (Fig. 10). At the Verde River section, only the sequence A-B remains. These relations indicate that there is a major stratigraphic break between the Tapeats Sandstone and Chino Valley Formation which represents a much longer time interval than shown by the apparently gradational contact (Hereford, 1975) between the formations. Therefore, I suggest that the Cambrian age of the Chino Valley Formation now seems unlikely and that the formation is Devonian because it, as well as the overlying Martin Formation, truncates the Cambrian section. Reasons for not considering the Chino Valley Formation to be Ordovician or Silurian are given in Hereford (1975).

Facies F in the Payson and Pine Mountain areas (Fig. 1) has been considered the Cambrian Tapeats Sandstone by some and a Devonian sandstone underlying the lower dolomitic sandstone of the Martin Formation (Devonian) by others (according to Shride, 1967, p. 76). Teichert (1965, p. 14–27), in the most recent study of the stratigraphy of the basal sandstone (Tapeats Sandstone) and overlying dolomitic sandstone in these areas, concluded that both units are Devonian and equivalent to the Beckers Butte Member (defined by Teichert) of the Martin Formation. However, the Tapeats Sandstone can be traced from Jerome into the Pine Mountain area where it is lithologically identical to the basal sandstone in the Payson area. There are no exposures between Pine Mountain and the East Verde River, but the lithologic similarities between the sandstone in each area justifies correlation. Teichert (1965, p. 25–27) recognized that the basal sandstone in the Black Hills and the Payson area are lithologically similar; nevertheless, he thought the two sandstones were of different ages because he found *Corophioides* in the Tapeats Sandstone. North of Pine Mountain at the Houston Creek section and 5 km to the east near the Verde River (Figs. 1, 10), facies D, containing numerous *Corophioides,* lies within a sequence that otherwise is identical to facies F in the Payson area. This establishes beyond reasonable doubt the stratigraphic correlation of the Tapeats Sandstone across central Arizona. In the Payson area, the term "Beckers Butte" should be applied only to the slope-forming dolomitic sandstone and local conglomerate (described by Teichert, 1965) underlying typical Martin Formation and overlying facies F of the Tapeats Sandstone. Strata herein assigned to the Beckers Butte differs lithologically from the Chino Valley Formation; additional work in the Limestone and southern Black Hills will be necessary to determine the stratigraphic relations between the units.

The unconformity along the top of the Tapeats is marked by as much as 1 to 6 m of relief. In these depressions or channels, particularly in Chino Valley, the Tapeats is extensively reworked and incorporated into the overlying Chino Valley Formation. Outside the channels, the contact is sharp and only slightly erosional. Consequently,

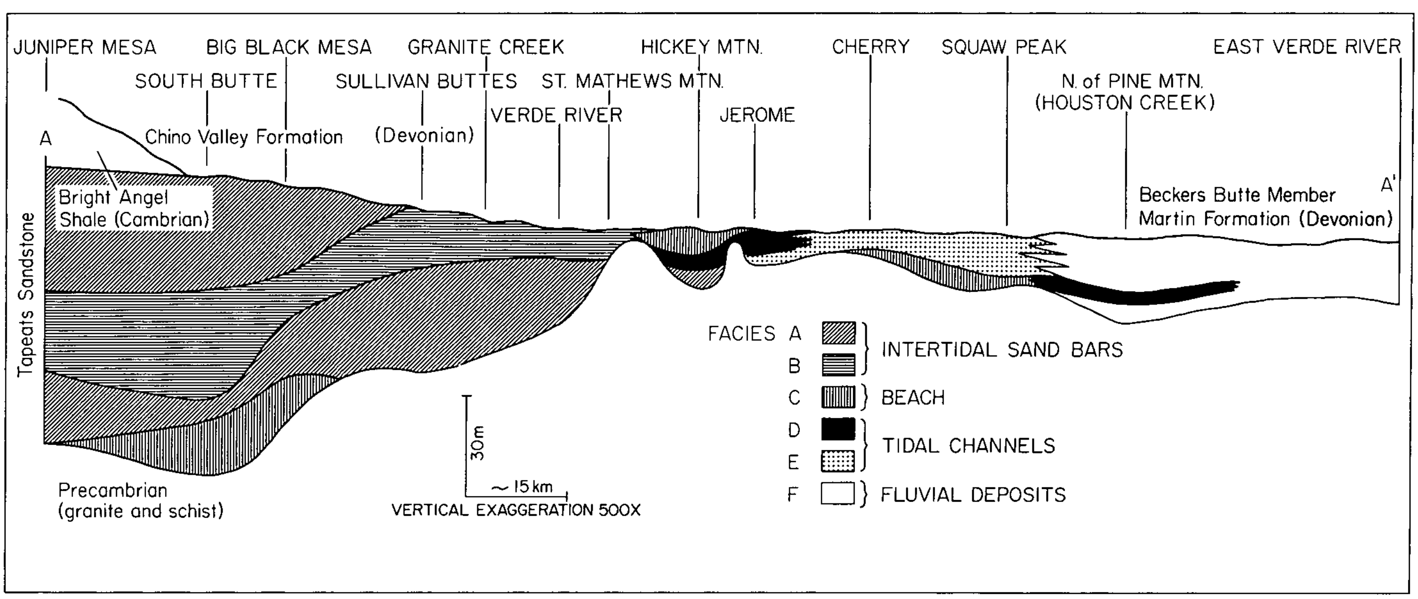

Figure 10. Correlation of the six facies along the section A-A′ (Fig. 1).

the contact between the Tapeats Sandstone and Chino Valley Formation in many places appears gradational, although close inspection invariably reveals cross-cutting relations between the formations. The apparent conformity between the Tapeats Sandstone and the Chino Valley Formation was the evidence I used for assigning the Chino Valley Formation to the Cambrian(?) (Hereford, 1975). Recognition of the purely erosional origin of the contact between the formations and of the regional truncation of the Tapeats by the Chino Valley Formation (Fig. 10) is compelling evidence for assigning the latter to the Devonian.

The outcrops at Houston Creek and near the Verde River east of Houston Creek (Figs. 1, 10) have been studied extensively because the intertonguing relations between facies D and F are well exposed. The contact between the facies is sharp but not erosional. In the eastern Limestone Hills and Table Mountain, the two facies alternate in a complex manner through most of the section, which exceeds 40 m in thickness. North of Table Mountain and the Limestone Hills, however, the section consists entirely of a comparable thickness of facies F. Apparently, the area generally south of the bend in the East Verde River was near the interface between entirely fluvial and entirely marine deposition.

Despite the fact that only a partial record of Cambrian deposition is preserved in central Arizona, enough remains to construct a reasonable depositional model. The model proposed here assumes that in the western half of central Arizona the Tapeats Sandstone was deposited primarily on sandy intertidal flats where the rise and fall of the tides molded the coarse sediment into many different bedforms. Some of these forms retain the imprint of a periodic transport system. Deposition on the tidal flats, by analogy with modern counterparts, is thought to have been governed by the diminishing energy of tidal currents flowing shoreward across the gently sloping tidal flats. A second component of this model assumes that prevegetation bed-load streams, originating in east-central Arizona, transported coarse, poorly sorted sediment westward to the principal site of deposition. Figure 11 illustrates the depositional model.

In modern intertidal environments, the diminishing energy of tidal currents produces a shoreward distribution of sediments generally recognizable by variations in grain size, bedding thickness, sedimentary structures, and lithology. Such a zonation apparently exists in the Tapeats Sandstone. The coarse-grained, thickly bedded deposits of facies A formed on the low tidal flat; the thinner-bedded, finer-grained facies B formed on mid-tidal flat; and facies D and E formed on the higher parts of the tidal flat (Fig. 11). Although they are coarse grained, facies D and E are enriched in fine-grained sediment. A distribution of facies similar to that proposed for the Tapeats Sandstone occurs on the intertidal flats of the Minas Basin, Bay of Fundy. Here Swift and McMullen (1968, p. 179) mapped megarippled intertidal sand bars (corresponding to facies A and B) bounded by higher tidal flats consisting of sand bars drained by braided channels (perhaps similar to facies E).

Facies F should terminate at the shoreline against either facies D or facies E. The stratigraphic (Fig. 10) and spatial relations in the Pine Mountain and Squaw Peak areas suggest that facies E, rather than facies D, generally terminated against facies F. The shoreline was probably unstable because it was not restrained by vegetation and the prevegetation bed-load channels could undergo large variations in discharge (Schumm, 1968). As a result, the shoreline was free to shift well seaward, far onto the tidal flats, producing a complex interfingering of marine and nonmarine strata which as yet is recognizable only in facies E. Finally, facies C developed where the slope on the depositional surface increased over that on the adjoining tidal flats. Under these conditions, shoaling waves produced swash zones on which a suite of sedimentary structures similar to modern beaches developed.

The long-term position of the shoreline determines the vertical distribution of facies. A prograding shoreline should ideally produce the sequence A-B-D or E. In northwestern Chino Valley, the sequence A-B-A broadly suggests progradation followed by submergence. Farther east, this sequence was truncated by erosion, but the preserved record suggests the expected lateral relation between facies B and D (Fig. 10, Hickey Mountain area).

The amplitude of tides during the late Precambrian and early Paleozoic has received considerable attention (in particular, see Olson, 1970; Merifield and Lamar, 1970). Klein (1971) reviewed several depositional models based on modern intertidal environments and proposed a model for recognition of ancient intertidal deposits and for determination of paleotidal range. Complete sequences that define paleotidal range (Klein, 1971) were not found in the Tapeats Sandstone. However, judging from the thickest sets of cross-stratification, tidal ranges in excess of 3 m must have been common during deposition of the Tapeats. It is likely, however, that tidal ranges were two to three times larger than this figure. Klein and Whaley (1972) reported that under intertidal conditions movement begins in water depths of 5 to 10 m for sand and bedforms roughly comparable in size and height to the grain size and bedding thickness of the Tapeats Sandstone. Applying his model to equivalent Cambrian strata in the nearby southeastern Great Basin, Klein (1972) measured minimum paleotidal ranges between 1 and 8 m, which suggests that ranges between 5 and 10 m are reasonable for the Tapeats Sandstone. These figures (5 to 10 m) may seem excessive, but the relief on the Precambrian surface was

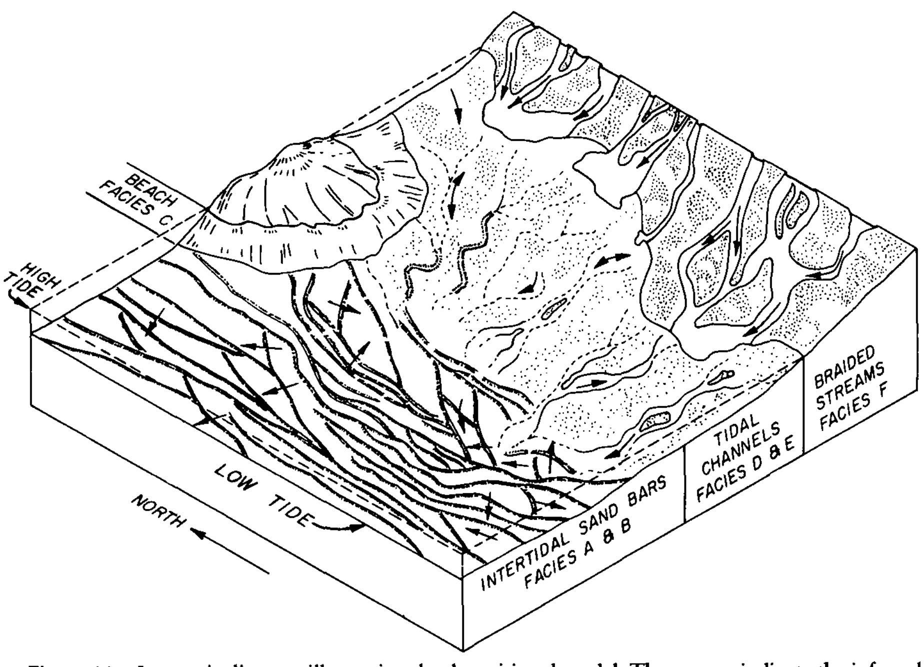

Figure 11. Isometric diagram illustrating the depositional model. The arrows indicate the inferred direction of sediment transport. Horizontal scale is unknown, but the west to east edge could reasonably be 10 to 50 km in length.

probably sufficient to increase tidal amplitudes.

CONCLUSIONS

Distinctive types of cross-stratification, relatively wide dispersion of cross-stratification azimuths, and abundance of trace fossils suggest that in the western part of the study area the Tapeats Sandstone was deposited primarily on sandy intertidal areas. In contrast, abundant scour-and-fill structure, low dispersion of cross-stratification, and absence of trace fossils suggest that fluvial deposition prevailed in the eastern part of the area. Dispersion of cross-stratification, often used to distinguish marine from fluvial strata, decreases in an easterly direction. In the western part of the area, the range of dispersion lies within the limits typical of ancient marine rocks; in the eastern part, the dispersion is within the limits of ancient fluvial rocks. In addition, frequency distributions of foreset azimuths are commonly polymodal in the west, whereas in the east they are unimodal.

In the area of intertidal deposition, the relief on the Precambrian surface locally influenced deposition and may have increased tidal amplitudes. In the former case, an increase in the slope of the surface allowed shoaling waves, rather than the ebb and flood of tidal currents, to produce structures similar to those on modern beaches (facies C). Throughout most of this area, however, the sedimentary structures and distribution of sediments appear to have been controlled primarily by the ebb and flood of tides. Tidal ranges were probably high but are within reasonable limits. The waning energy of currents spreading shoreward across the tidal flats produced a distinctive zonation of sediments. The low tidal flats consisted of coarse-grained, thickly cross-stratified sand bars (facies A and B), and on the high tidal flat, coarse-grained sand bars drained by braided channels (facies E) or meandering streams (facies D) developed. A roughly comparable zonation of sediments has been described from modern intertidal flats.

The high tidal flat deposits resemble modern fluvial deposits but contain marine trace fossils and have a relatively high dispersion of cross-stratification azimuths. Shoreline migration and the similarities in stratification sequences of meandering and braided tidal channels with their fluvial counterparts produced these similarities. The shoreline, which was not confined by vegetation, freely migrated well onto the tidal flats, thereby producing a complex interfingering of nonmarine with marine strata as yet distinguishable only in facies E. As a result, the high tidal flat deposits (facies D and E) and even those of the lower parts of the tidal flat (facies A and B) probably contain some nonmarine strata.

Throughout the eastern part of the area, coarse, poorly sorted sediment was transported westward and deposited by a fluvial system that resembled modern braided streams. This was a prevegetation fluvial system in which wide bed-load streams with braided channels are expected to predominate (facies F). The coarse sediment was probably spread as sheets over the deeply weathered Precambrian surface, covering and only slightly reworking regolith.

Erosion beginning not later than Late Cambrian beveled the Tapeats Sandstone in an easterly direction. The partially preserved section in west-central Arizona appears to record a prograding shoreline because the facies deposited on midtidal flats overlie those of the lower tidal flat. In the westernmost portion, where a complete section is preserved, low tidal flat deposits lie on those of the midtidal flat, suggesting that progradation was followed by submergence.

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REFERENCES CITED

Ager, D. V., and Wallace, Peigi, 1970, The distribution and significance of trace fossils in the uppermost Jurassic rocks of the Boulonnais, northern France, *in* Crimes, T. P., and Harper, J. C., eds., Trace fossils: Geol. Jour. Spec. Issue no. 3, p. 1–18.

Allen, J.R.L., 1965, A review of the origin and characteristics of Recent alluvial sediments: Sedimentology, v. 5, p. 89–191.

——1967, Notes on some fundamentals of paleocurrent analysis, with reference to preservation potential and sources of variance: Sedimentology, v. 9, p. 75–88.

Alpert, S. P., 1974, Trace fossils of Precambrian-Cambrian succession, White-Inyo Mountains, California [Ph.D. thesis]: Los Angeles, Univ. California, 162 p.

Banks, N. L., 1973a, Tide-dominated offshore sedimentation, Lower Cambrian, north Norway: Sedimentology, v. 20, p. 213–228.

——1973b, Falling-stage features of Precambrian braided stream: Criteria for sub-aerial exposure: Sed. Geology, v. 10, p. 147–154.

Banks, N. L., and Collinson, J. D., 1973, Discussion of "Some sedimentological aspects of planar cross-stratification in a sandy braided river": Jour. Sed. Petrology, v. 43, p. 265–267.

Bather, F. A., 1925, U-shaped markings near Blea Wyke: Yorkshire Geol. Soc. Proc., v. 20, p. 185–199.

Bluck, B. J., 1974, Structure and directional properties of some valley sandur deposits in southern Iceland: Sedimentology, v. 21, p. 533–554.

Boersma, J. R., 1969, Internal structure of some tidal mega-ripples on a shoal in the Westerschelde estuary, the Netherlands report of a preliminary investigation: Geologie en Mijnbouw, v. 48, p. 409–414.

Chanda, S. K., and Bhattacharyya, Ajit, 1974, Ripple-drift cross-lamination in tidal deposits: Examples from the Precambrian Bhander Formation of Maihar, Satna District, Madhya Pradesh, India: Geol. Soc. America Bull., v. 85, p. 1117–1122.

Collinson, J. D., 1968, Deltaic sedimentation units in the upper Carboniferous of northern England: Sedimentology, v. 10, p. 233–254.

——1970, Bedforms of the Tana River, Norway: Geog. Annaler, ser. A, v. 52, p. 31–56.

Costello, W. R., and Walker, R. G., 1972, Pleistocene sedimentology, Credit River, southern Ontario: A new component of the braided river model: Jour. Sed. Petrology, v. 42, p. 389–400.

Curray, J. R., 1956, The analysis of two-dimensional orientation data: Jour. Geology, v. 64, p. 117–131.

de Raaf, J.F.M., and Boersma, J. R., 1971, Tidal deposits and their sedimentary structures: Geologie en Mijnbouw, v. 50, p. 479–504.

Doeglas, D. J., 1962, The structure of sedimentary deposits of braided rivers: Sedimentology, v. 1, p. 167–190.

Häntzschel, Walter, 1962, Trace fossils and problematica, *in* Treatise on invertebrate paleontology, Pt. W, Miscellanea: New York, Geol. Soc. America (and Univ. Kansas Press), p. W177–W245.

Harms, J. C., and others, 1975, Depositional environments as interpreted from primary sedimentary structures and stratification sequences: Tulsa, Okla., Soc. Econ. Paleontologists and Mineralogists Short Course 2, 161 p.

Hayes, M. O., Anan, F. S., and Bozeman, R. N., 1969, Sediment dispersal trends in the littoral zone, a problem in paleogeographic reconstruction, *in* Coastal environments of northeastern Massachusetts and New Hampshire, Univ. Mass. Coastal Research Group, Field Trip Guidebook: Amherst, Univ. Massachusetts, Geology Dept., p. 290–315.

Hecker, R. T., 1970, Palaeoichnological research in the Palaeontological Institute of the Academy of Sciences of the USSR, *in* Crimes, T. P., and Harper, J. C., eds., Trace fossils: Geol. Jour. Spec. Issue no. 3, p. 215–226.

Heinberg, C., 1970, Some Jurassic trace fossils from Jameson Land (East Greenland), *in* Crimes, T. P., and Harper, J. C., eds., Trace fossils: Geol. Jour. Spec. Issue no. 3, p. 227–234.

Hereford, Richard, 1974, Depositional environments and paleotides as inferred from sedimentary structures in the Cambrian Tapeats Sandstone, west central Arizona [abs.]: Geol. Soc. America Abs. with Programs (Rocky Mountain Sec.), v. 6, p. 448.

——1975, Chino Valley Formation (Cambrian?) in northwestern Arizona: Geol. Soc. America Bull., v. 86, p. 677–682.

Hollander, Myles, and Wolfe, D. A., 1973, Nonparametric statistical methods: New York, John Wiley & Sons, Inc., 503 p.

Johnson, H. D., 1975, Tide- and wave-dominated inshore and shoreline sequences from the late Precambrian, Finnmark, North Norway: Sedimentology, v. 22, p. 45–74.

Jopling, A. V., and Walker, R. G., 1968, Morphology and origin of ripple-drift cross-lamination, with examples from the Pleistocene of Massachusetts: Jour. Sed. Petrology, v. 38, p. 971–984.

Klein, G. deV., 1963, Bay of Fundy intertidal zone sediments: Jour. Sed. Petrology, v. 33, p. 844–854.

——1965, Dynamic significance of primary structures in the Middle Jurassic Great Oolite Series, southern England, *in* Middleton, G. V., ed., Primary sedimentary structures and their hydrodynamic interpretation: Tulsa, Okla., Soc. Econ. Paleontologists and Mineralogists Spec. Pub. No. 12, p. 173–191.

——1967, Comparison of recent and ancient tidal flat and estuarine sediments, *in* Lauff, G. H., ed., Estuaries: Washington, D.C., Am. Assoc. Adv. Sci. Pub. no. 83, p. 207–218.

——1970a, Depositional and dispersal dynamics of intertidal sand bars: Jour. Sed. Petrology, v. 40, p. 1095–1127.

——1970b, Tidal origin of a Precambrian quartzite: The lower fine-grained quartzite (Middle Dalradian) of Islay, Scotland: Jour. Sed. Petrology, v. 40, p. 973–985.

——1971, A sedimentary model for determining paleotidal range: Geol. Soc. America Bull., v. 82, p. 2585–2592.

——1972, Determination of paleotidal range in clastic sedimentary rocks, *in* Stratigraphy and sedimentology, Section 6, Internat. Geol. Cong., no. 24, Proc., p. 397–405.

Klein, G. deV., and Sanders, J. E., 1964, Comparison of sediments from Bay of Fundy and Dutch Wadden Sea tidal flats: Jour. Sed. Petrology, v. 34, p. 18–24.

Klein, G. deV., and Whaley, M. L., 1972, Hydraulic parameters controlling bedform migration of an intertidal sand body: Geol. Soc. America Bull., v. 83, p. 3465–3470.

Krieger, M. H., 1959, Cambrian age of some of the basal Paleozoic sandstone in central Arizona [abs.]: Geol. Soc. America Bull., v. 70, p. 1729.

——1965, Geology of the Prescott and Paulden quadrangles, Arizona: U.S. Geol. Survey Prof. Paper 467, 127 p.

Land, C. B., Jr., 1972, Stratigraphy of Fox Hills Sandstone and associated formations, Rock Springs uplift and Wamsutter arch area, Sweetwater County, Wyoming: A shoreline-estuary sandstone model for the late Cretaceous: Colorado School Mines Quart., v. 67, no. 2, p. 1–69.

Land, L. S., and Hoyt, J. H., 1966, Sedimentation in a meandering estuary: Sedimentology, v. 6, p. 191–207.

Lochman-Balk, Christina, 1971, The Cambrian of the craton of the United States, *in* Holland, C. H., ed., Cambrian of the New World: New York, John Wiley & Sons, Inc., p. 79–167.

MacKenzie, D. B., 1968, Studies for students, sedimentary features of Alameda Avenue road cut, Denver, Colorado: Mtn. Geologist, v. 5, p. 3–13.

Mardia, K. V., 1972, Statistics of directional data: New York, Academic Press, 357 p.

McDonald, B. C., and Banerjee, Indranil, 1971, Sediments and bedforms on a braided outwash plain: Canadian Jour. Earth Sci., v. 8, p. 1282–1301.

McKee, E. D., 1945, Stratigraphy and ecology of the Grand Canyon Cambrian, Pt. 1 *of* Cambrian history of the Grand Canyon region: Carnegie Inst. Washington Pub. 563, p. 3–168.

——1957, Primary structures in some Recent sediments (U.S. and Mexico): Am. Assoc. Petroleum Geologists, v. 41, p. 1704–1747.

——1966, Significance of climbing-ripple structure: U.S. Geol. Survey Prof. Paper 550-D, p. D94–D103.

McNair, A. H., 1951, Paleozoic stratigraphy of part of northwestern Arizona: Am. Assoc. Petroleum Geologists Bull., v. 35, p. 503–541.

Merifield, P. M., and Lamar, D. L., 1970, Paleotides and the geologic record, *in* Paleogeophysics: New York, Academic Press, p. 31–40.

Olson, W. S., 1970, Tidal amplitudes in geologic history: New York Acad. Sci. Trans., ser. 2, v. 32, p. 220–233.

Ore, H. T., 1964, Some criteria for recognition of braided stream deposits: Wyoming Univ. Contr. Geology, v. 3, p. 1–14.

Palmer, A. R., 1971, The Cambrian of the Great Basin and adjacent areas, western United States, *in* Holland, C. H., ed., Cambrian of the New World: New York, John Wiley & Sons, Inc., p. 1–78.

Potter, P. E., and Pettijohn, F. J., 1963, Paleocurrents and basin analysis: Berlin, Springer-Verlag, 296 p.

Reineck, H. E., 1967, Layered sediments of tidal flats, beaches, and shelf bottoms of the North Sea, *in* Lauff, G. H., ed., Estuaries: Washington, D.C., Am. Assoc. Adv. Sci. Pub. no. 83, p. 191–206.

Richter, Rudolf, 1926, Flachseebeobachtungen zur Paläontologie und geologie XII-XIV: Senckenbergiana Lethaea, v. 8, p. 200–224.

Schumm, S. A., 1968, Speculations concerning paleohydrologic controls of terrestrial sedimentation: Geol. Soc. America Bull., v. 79, p. 1573–1588.

Seeland, D. A., 1969, Marine current directions in upper Precambrian and Cambrian rocks of the southern United States, *in* Geology and natural history of the Grand Canyon region: Durango, Colo., Four Corners Geol. Soc. 5th Field Conf., p. 123–126.

Sellwood, B. W., 1970, The relation of trace fossils to small scale sedimentary cycles in the British Lias, *in* Crimes, T. C., and Harper, J. C., eds., Trace fossils: Geol. Jour. Spec. Issue no. 3, p. 489–504.

Shride, A. F., 1967, Younger Precambrian geology in southern Arizona: U.S. Geol. Survey Prof. Paper 566, 89 p.

Smith, John, 1893, Peculiar U-shaped tubes in sandstone near Crawfurdland Castle and in Gowkha Quarry, near Kilwinning: Geol. Soc. Glasgow Trans. v. 9, p. 289–292.

Smith, N. D., 1970, The braided stream depositional environment: Comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians: Geol. Soc. America Bull., v. 81, p. 2993–3013.

——1972, Some sedimentological aspects of planar cross-stratification in a sandy braided river: Jour. Sed. Petrology, v. 42, p. 624–634.

——1973, Some sedimentological aspects of planar cross-stratification in a sandy braided river: A reply to N. L. Banks and J. D. Collinson: Jour. Sed. Petrology, v. 43, p. 267–269.

Stewart, J. H., 1970, Upper Precambrian and Lower Cambrian strata in the southern Great Basin, California and Nevada: U.S. Geol. Survey Prof. Paper 620, 206 p.

Swett, Keene, Klein, G. deV., and Smit, D. E., 1971, A Cambrian tidal sand body: The Eriboll sandstone of northwest Scotland: An ancient-Recent analog: Jour. Geology, v. 79, p. 400–415.

Swift, D.J.P., and McMullen, R. M., 1968, Preliminary studies of intertidal sand bodies in the Minas Basin, Bay of Fundy, Nova Scotia: Canadian Jour. Earth Sci., v. 5, p. 175–183.

Teichert, Curt, 1965, Devonian rocks and paleogeography of central Arizona: U.S. Geol. Survey Prof. Paper 464, 181 p.

Thompson, W. O., 1937, Original structures of beaches, bars, and dunes: Geol. Soc. America Bull., v. 48, p. 723–751.

Titley, S. R., 1962, Geology along the Diamond Rim and adjoining areas, Gila and Navajo Counties, Arizona, *in* Guidebook of the Mogollon Rim region, east-central Arizona, New Mexico Geol. Soc., 13th Field Conf.: Socorro, New Mexico Bur. Mines and Mineral Resources, p. 123–128.

Tucker, M. E., 1973, The sedimentary environments of tropical African estuaries, Freetown Peninsula, Sierra Leone: Geologie en Mijnbouw, v. 52, p. 203–215.

van Straaten, L.M.J.U., 1954, Composition and structure of Recent marine sediments in the Netherlands: Leidse Geol. Meded., v. 19, p. 1–110.

——1959, Minor structures of some Recent littoral and neritic sediments: Geologie en Mijnbouw, n.s., v. 21, p. 197–216.

Wanless, H. R., Jr., 1973, Cambrian of the Grand Canyon: A reevaluation of the depositional environment [Ph.D. thesis]: Baltimore, Maryland, Johns Hopkins Univ., 128 p.

Williams, G. E., 1971, Flood deposits of the sand-bed ephemeral streams of central Australia: Sedimentology, v. 17, p. 1–40.

Williams, P. F., and Rust, B. R., 1969, The sedimentology of a braided river: Jour. Sed. Petrology, v. 39, p. 649–679.

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