Sedimentology – Elsevier Publishing Company, Amsterdam – Printed in The Netherlands

FORMATION OF LARGE-SCALE CROSS-BEDDING IN A CARBONATE UNIT

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(Received July 12, 1967)

SUMMARY

The type Lindsey Bridge Member of the Moorefield Formation of northeastern Oklahoma consists of 24 ft. of massively cross-bedded limestone. Cross-bed shape, lithologic variation, grain size and sorting, distribution of insolubles, and distribution of fossils and fossil burrows can be explained with reference to a hydrodynamic model developed in recent flume studies.

Three facies can be distinguished in this unit: (*1*) foresets, thick-bedded, well-sorted, fine to medium crinoidal grainstones, dipping at angles up to 18°; (*2*) toesets, which are thin-bedded, poorly sorted, skeletal packstones notably more fossiliferous than the foresets, with which they are laterally gradational; toesets dip at approximately 5°–8°; (*3*) bottomsets, composed of argillaceous, fine-grained (mainly silt-size), skeletal limestones. Foresets overlie previously deposited bottomsets; this geometry is typical of regressive sedimentation.

The exposure is adjacent to a pre-Moorefield topographic high. As currents crossing this high entered a basin on the downcurrent side, flow separation occurred. Bed material load was deposited mainly on the foreset slope, suspension material mainly in toeset and bottomset areas. The poor sorting of the toesets is in part due to reverse circulation, formed by the flow separation, which transported bottomset sediment back toward the foreset. JOPLING (1965b) has shown that this depositional geometry produces tangential cross-beds similar to those seen in this outcrop. Differential settling velocity, substrate stability, and abundance of organic detritus influenced other sedimentologic properties of the deposit.

INTRODUCTION

The type section of the Lindsey Bridge Member of the Moorefield Formation (Upper Mississippian) of northeastern Oklahoma occurs in a bluff along the Grand River (Fig.1 and 2). At this locality, the member consists of a carbonate unit, 20–24 ft. thick, which is in part massively cross-bedded. Excellent exposures allow tracing of a single 14-ft. cross-bed set for several hundred feet along the outcrop and thus provide an excellent opportunity to study petrologic and sedimentologic changes associated with cross-bed formation in a carbonate deposit.

With the exception of studies by, e.g., IMBRIE and BUCHANAN (1965) and KLEIN (1965), detailed analysis of cross-bedding in carbonate rocks has been neglected. This paper represents an attempt to extend (*1*) our understanding of cross-bedding in carbonates, and (*2*) our knowledge of the formation of cross-beds particularly those of large scale. Analysis of this deposit in light of hydrodynamic principles derived from recent flume studies should be particularly interesting in view of the large difference in scale and the contrast in sedimentary materials (quartz sands in the flume, carbonate sands and silts in the outcrop). In addition, consideration of minor associated sedimentary features (distribution of fossils, variation and distribution of organism burrows, and so forth) within this interpretational framework may increase our understanding of the environmental significance of some of these phenomena. The purposes of this study, then, are: (*1*) to examine and describe lithologic variation and the occurrence and distribution of minor sedimentary features associated with large-scale cross-bedding in a carbonate unit, and (*2*) to interpret these features in light of hydrodynamic principles derived from recent flume studies.

Until recently, most environmental interpretations of cross-stratification were based on correlation of the type of cross-bed with the depositional environment in which it is generally found in Recent deposits. Association of cross-bed type with

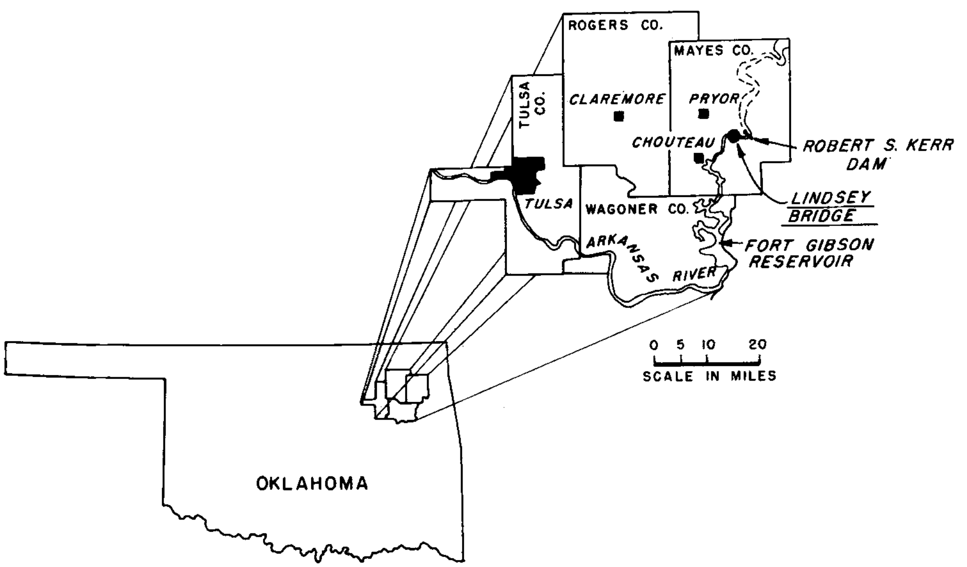

Fig.1. Locality map. The outcrop under discussion is located at Lindsey Bridge, the type locality for the Lindsey Bridge Member of the Moorefield Formation.

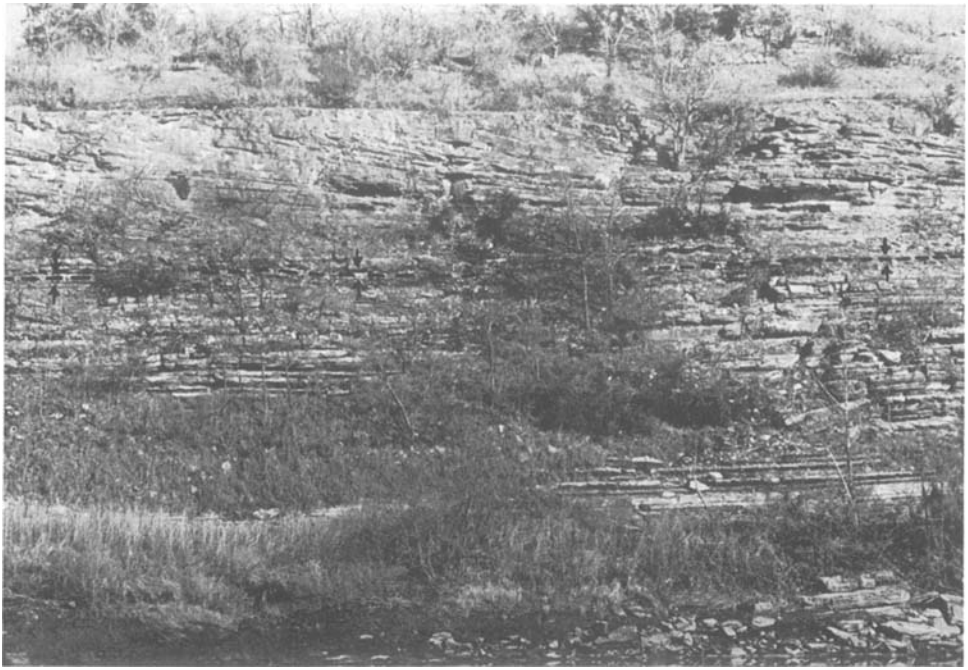

Fig.2. Photograph of the outcrop. The top of the Lindsey Bridge Member is the sharp break immediately below the vegetation at the top of the Bluff. The approximate lower boundary of the Lindsey Bridge Member is shown by the dashed line. The cross-bed set is approximately 14 ft. thick and is underlain by 10 ft. of thin-bedded limestones of the Lindsey Bridge Member. The remainder of the outcrop is Bayou Manard Member of the Moorefield Formation. Cross-bed dips seen in this photograph are apparent rather than real, as is the relief on the basal contact.

other sedimentary structures has also been used in this regard. McKEE (1964) has recently summarized much of the literature concerning this approach. Fewer investigators have attempted to analyze cross-strata in terms of the process of formation because of the general lack of understanding, until recently, of the relationship between sediment transport mechanics and bedding formation. Recent flume studies on sediment transport by SIMONS and RICHARDSON (1961) SIMONS et al. (1961), and others have led to significant advances in understanding the factors influencing configuration and movement of bed forms such as ripples, dunes, and sand-waves[1]. JOPLING (1961, 1963, 1964, 1965a, b) has reported a series of flume experiments in which he has applied the principles of sediment transport mechanics to the formation of bedding and lamination. He has also examined the formation and modification of small laboratory

[1] The important geologic aspects of much of this work have been summarized recently in MIDDLETON (1965).

deltas with particular emphasis on (*1*) factors influencing the shape of cross-strata, especially for the transition from tabular to tangential (low-angle) cross-bedding (1963, 1965a); (*2*) sorting processes which influence the development of bedding lamination (1964) and (*3*) grain-size distribution in laboratory cross-beds (1965b). ALLEN (1965) has reported results of laboratory experiments on the formation of cross-bedding which agree closely with those of JOPLING (1965b). The interested reader is referred to the papers cited above for detailed information on sediment transport mechanics.

The results of these flume studies have been applied by several investigators to the hydrodynamic interpretation of field occurrences of cross-strata. HARMS et al. (1963), and HARMS and FAHNESTOCK (1965) have used these flume-derived data in the interpretation of Recent river deposits, while ALLEN (1963) and VISHER (1965a, b) have utilized them in regard to ancient fluvial deposits. ALLEN and NARAYAN (1964) have described large-scale cross-beds from the Cretaceous of Great Britain and have interpreted their origin, on the basis of flume experiments, in terms of migration of large-scale ripples influenced by tidal conditions. JOPLING (1966) has recently reported a study in which he has reconstructed the hydraulic parameters of an ancient flow regime by the application of the principles of sediment transport mechanics to the sedimentologic characteristics of the resultant deposit. This study is of importance not only as the first attempt to reconstruct a paleoflow regime quantitatively, but also as a detailed compilation and review of many facets of sediment transport mechanics.

Primary sedimentary structures in carbonate deposits have generally aroused less interest than those found in terrigenous clastics. Recent papers by, e.g., IMBRIE and BUCHANAN (1965) and by KLEIN (1965) are of particular interest in light of the lack of detailed investigation of cross-bedding in carbonate sediments. IMBRIE and BUCHANAN (1965) have applied the results of flume studies and theoretical consideration of hydrodynamics to the interpretation of several types of cross-stratification found in Recent carbonate deposits in the Bahamas and Florida. KLEIN (1965) made use of the hydraulic interpretation of primary structures in the environmental interpretation of the Great Oolite Series in southern England. A list of papers referring to sedimentary structures in carbonates may be found in his paper.

DESCRIPTION OF THE UNIT

The general nature of the Lindsey Bridge Member at the outcrop under discussion is shown in Fig.2 and diagrammatically in Fig.3. At the western end of the outcrop, a single set of tangential (JOPLING, 1963), low-angle cross-beds, 14 ft. in maximum thickness, overlies approximately 10 ft. of thinly-bedded, flat-lying, argillaceous limestones. To the east, the cross-beds grade into thinly-bedded, flatlying, argillaceous limestones identical to those at the base of the unit. The point of transition from dipping to flat-lying beds moves progressively higher in the unit toward the east so that the member comprises approximately 21 ft. of thinly-bedded, horizontal

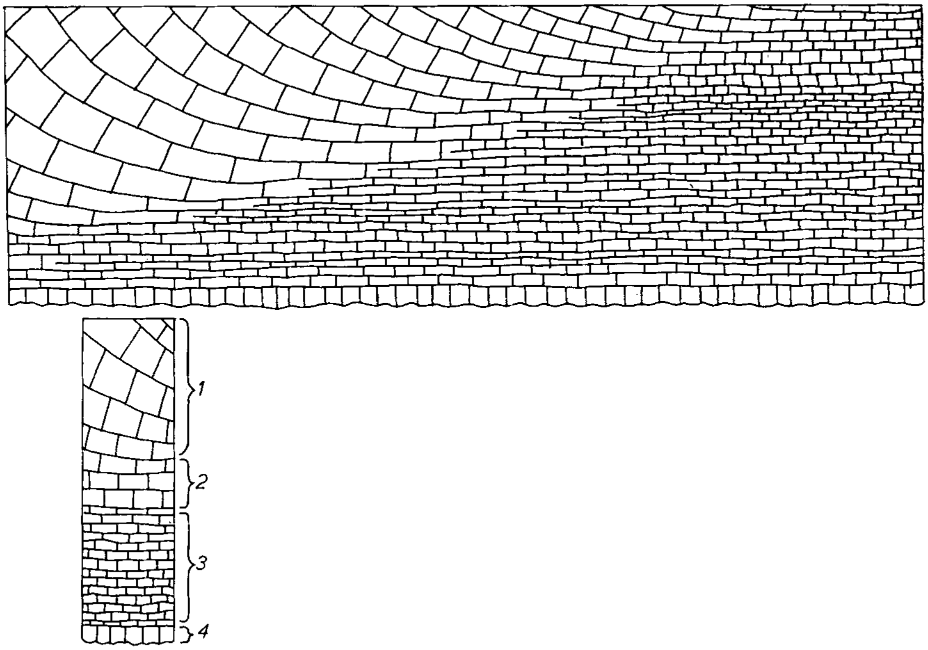

Fig.3. Diagrammatic representation of the Lindsey Bridge Member at the outcrop shown in Fig.2· Note lateral facies change and vertical variation. Dips are exaggerated. Note also how the thick-bedded unit and the transition zone move higher in the section along the outcrop, indicating regressive sedimentation. Legend: *1* = thick-bedded (beds 0.5–4.0 ft. thick) individual beds thicken toward top. Dip increases from bottom to top; *2* = transition zone. Beds 0.3–0.7 ft. thick, horizontal or slightly dipping; *3* = thin-bedded (beds 0.05–0.4 ft. thick) argillaceous, essentially horizontal, extensively burrowed; *4* = coarse-grained, fossiliferous, locally cross-bedded.

limestones at the eastern end of the outcrop. This type of depositional geometry, in which beds nearer the "source" prograde over the "basinal" beds, is typical of regressive sedimentation.

Both the upper and lower contacts of the Lindsey Bridge Member at this locality are sharply defined disconformity surfaces. The lower boundary is an irregular and highly burrowed discontinuity surface which is traceable along the length of the outcrop. The stratigraphic significance of this regionally developed discontinuity will be discussed elsewhere (CIRIACKS et al., in preparation). The upper boundary of the unit is also disconformable as shown by truncation and removal of the upper parts of the cross-beds. This upper truncation surface limits the interpretation of the deposit in that the original thickness of the cross-bed set cannot be measured nor the total length of the foreset beds sampled. The lower reaches of the foresets have been preserved, however, and contribute much information to the interpretation of the unit.

The cross-beds under discussion have well-defined and lithologically distinctive foreset, toeset, and bottomset segments. The lithologic characteristics and spatial relationships of these segments or sets account for the lateral facies change and vertical variation seen on the outcrop. Due to bedding irregularities and weathering phenomena, no single cross-bed can be traced from the upper truncation surface to the distal portion of the bottomset beds. "Zones" of cross-beds are traceable, however, and this study is based on detailed observation, measurement, and sampling of two such zones. These zones are essentially packets of cross-beds which are bounded on top and bottom by well-defined bedding planes. The persistence of a particular zone is highly variable. In one case, the zone thins into a single bottomset bed which is traceable for approximately 150 ft. along the outcrop; this zone is the main subject of description here and will be referred to as zone *1*. In contrast, zone *2* grades rapidly into thinly-bedded bottomset units and loses its individuality within 110 ft. (horizontally) of the upper truncation. This variation in persistence is a reflection of depositional factors which will be discussed below. Data from zone *2* support the conclusions based on detailed analysis of zone *1*. Further discussion of zone *2* is omitted from this paper to avoid duplication.

One limitation of this study is that the face of the exposure is not parallel to the direction of maximum dip of the cross-beds. The exposure trend is N73°E; direction of maximum dip of the cross-beds varies from N35°E near the top of the exposed foresets to N10°E at the toeset–bottomset transition (Fig.4). Unfortunately, evidence of the three-dimensional shape of the deposit is lacking, though the dip directions indicate either deposition in a scoop-shaped trough (basin) or as a delta-shaped lobe of sediment. In either case, sediment transport was from the southwest.

Due to divergence between trend of the outcrop and direction of maximum cross-bed dip, lateral distances along the outcrop are quite different from distances in a maximum dip section. This difference varies along the outcrop and is maximum at the toeset–bottomset transition (Fig.4). Thus, the degree of lateral persistence of a

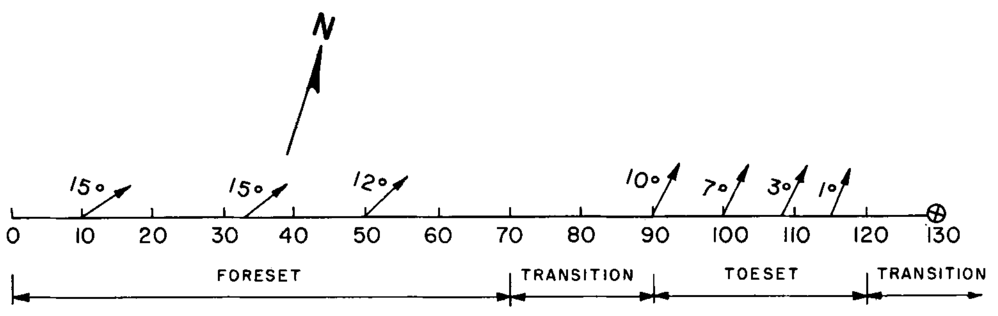

Fig.4. Diagrammatic plan view showing variation in direction and degree of maximum dip eastward along the outcrop. Outcrop trend is N 73°E. Scale along outcrop trend shows distance in feet from the point of intersection of zone *1* with the upper truncation surface. Arrows show direction of maximum dip, numbers give degree of dip. Note the variation in direction of maximum dip along the oucrop, and the abrupt decrease in degree of dip between 90 and 115 ft., in the toeset zone. Beds become horizontal at 130 ft.

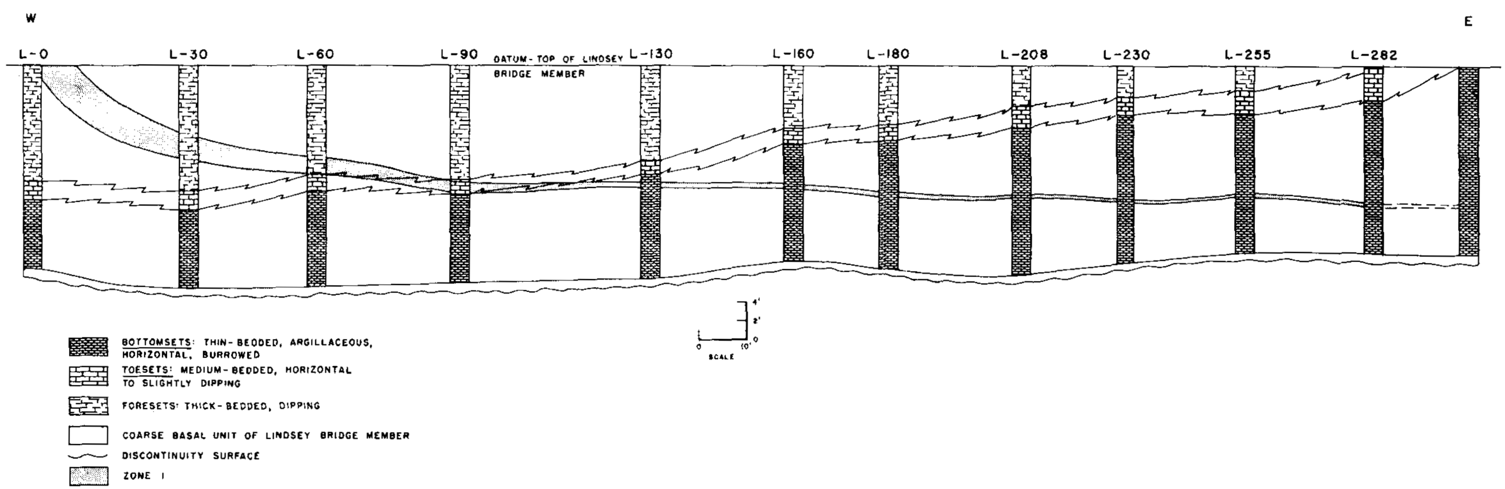

Fig.5. Cross-section, to scale, showing relationship of the cross-bed "zone" studied in detail (zone *I*) to foreset, toeset and bottomset distribution. Also shows thinning of zone *I* along the outcrop. Dips exaggerated due to vertical scale exaggeration. Upper and lower boundaries of the transition zone are based on field determinations. Bedding irregularity makes the picking of boundaries difficult and they are therefore only approximate. In the area where the transition zone crosses zone *I*, however, the transition zone boundaries picked in the field coincide well with boundaries based on detailed observation of hand specimens and thin sections.

particular zone, or distances measured between two points on the outcrop, hold little relevance other than for reference purposes. This divergence does account, in part, for the lateral persistence of bottomset beds for 150 ft. along the outcrop. The actual distance from the toeset–bottomset transition to the point at which bottomset beds become unrecognizable may be only $^1/_3$ or $^1/_5$ of the distance measured on the outcrop.

At its up-dip edge, zone *1* is 4.3 ft. thick and dips at approximately 18°. From this point, the zone thins down-dip, becomes progressively more thinly-bedded, and gradually approaches horizontality (Fig.5). At a lateral distance of 70 ft. from the truncation edge, the zone has thinned to 1.8 ft. and consists of five or six fairly distinct beds which dip at 7–8°. This is approximately the uppermost limit of the area of toeset deposition. The zone thins rapidly through the toeset area and at 120 ft. from the truncation intersection consists of a single horizontal bed 0.35 ft. thick. From here, the zone remains uniform in thickness and bedding until, at 280 ft., it loses its individuality and can no longer be traced. Transitions from foreset to toeset and toeset to bottomset beds are gradational over distances of 25–30 ft.

Though laterally gradational, foreset, toeset and bottomset segments are lithologically distinctive. Each has characteristics which are unique to its position in the deposit and which either aid in the interpretation of the depositional regime or can be explained by this interpretation.

Foreset beds

Foreset beds vary in thickness from a maximum of over 2 ft. at the upper exposed part of the foreset, to a minimum of a few inches at the transition to toeset beds (Fig.6). Major bedding planes are defined by thin layers of quartz silt or argillaceous material a few millimeters thick which weather back preferentially. Internally (i.e., between major bedding planes) the foresets are composed of individual laminae up to approximately 2 cm in thickness. These layers, which parallel the dip of the major bedding planes, are defined by thin laminae (1–2 mm) of quartz silt or argillaceous material, or by orientation of elongate skeletal fragments, and are individually well sorted.

Foresets are composed of well-sorted fine- to medium-grained[1] skeletal limestones (grainstones) (Fig.7). Rounded and abraded crinoid parts are the dominant grain component, but rounded, highly spherical bryozoan fragments and rounded, elongate fragments of brachiopod valves are present (Fig.8). Foreset sediments are cemented mainly by rim cement around the crinoid fragments, though some granular cement is present surrounding the brachiopod and bryozoan debris. Table I supplies data on constituent composition derived from point counts of thin sections. Though foreset beds are for the most part well sorted, some poorly sorted beds do occur. These generally

[1] All references to grain size are based on the WENTWORTH (1922) scale.

Fig.6. Outcrop photograph showing character of foreset and toeset beds. Foreset beds are the more massive beds in the upper part of the outcrop, toesets are the beds of medium thickness immediately below the foresets. Thin-bedded bottomsets can be seen below the toesets. The lines mark the approximate limits of the sets.

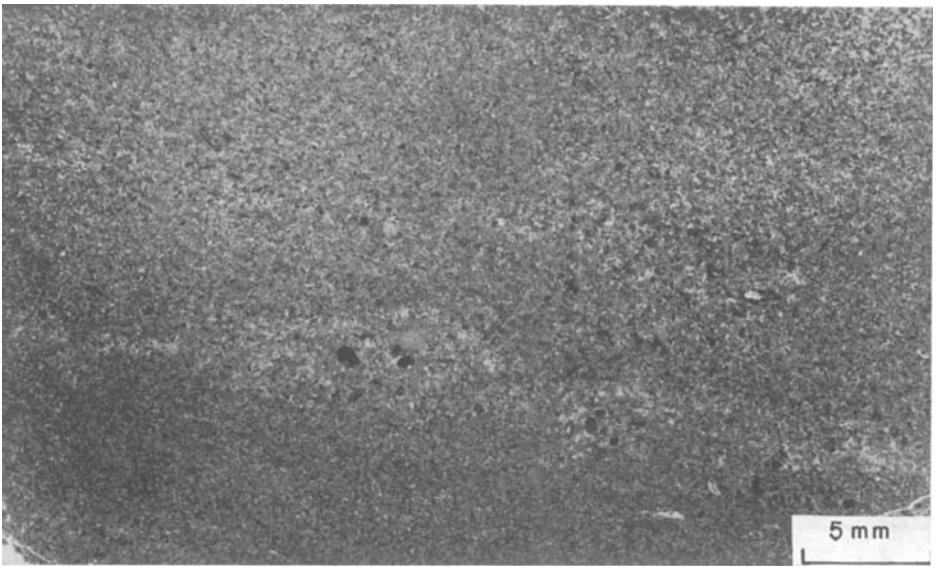

Fig.7. Foreset lithology—fine- to medium-grained well-sorted crinoid limestone. Positive print from thin-section.

contain admixtures both coarser and finer than the dominant fine to medium grain sizes. Whole fossils are virtually absent from the foresets, though large fragments of both brachiopod valves and fenestrate or ramose bryozoans are encountered at widely scattered points through the beds.

Maximum dips in the exposed foresets occur at the upper truncated edge and are approximately 18–20°; beds become progressively more flat-lying in a down-dip direction along the outcrop. As the foreset beds are traced down-dip (along the outcrop), several types of changes occur. The most obvious is the thinning of the zone as described above and shown in Fig.5. In addition, changes in grain size and sorting, and the amount of quartz silt or argillaceous admixture are important. Laminations of quartz silt and clay become more numerous down-dip. The thicker laminations contribute to the formation and definition of bedding planes, thus accounting for the general down-dip increase in the number of beds. Minor laminations occur approximately every 1–2 cm or less and cause the beds to weather into thinner layers. These layers are not consistently developed, however, and add to bedding irregularity in the lower part of the foreset.

Qualitatively, the median grain size decreases slightly down the foreset due to increased admixture of silt-size quartz, silt-size and clay-size carbonate, and clay minerals. In addition, however, the abundance of coarse-grained particles (brachiopod shells and fragments, bryozoan fragments and large crinoid plates) increases as the toesets are approached. This addition of both finer and coarser material has its most marked effect on sorting which decreases progressively toward the toesets. Adding

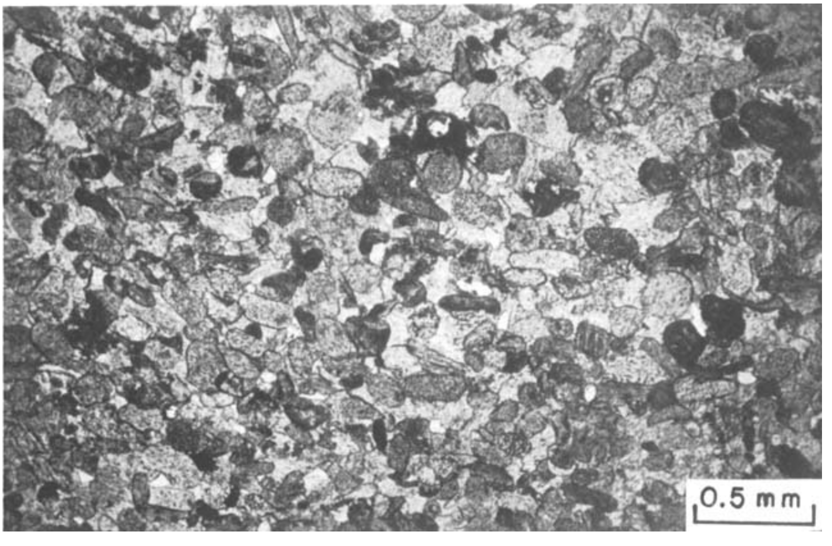

Fig.8. Thin-section photomicrograph of foreset lithology. Fine- to medium-grained crinoid limestone. Darker grains are mainly fragments of bryozoans and brachiopods. Note high degree of sorting and rounded nature of most grains.

TABLE I

CONSTITUENT COMPOSITION OF FORESET, TOESET AND BOTTOMSET LITHOLOGIES[1]

Set	*Echinoderm*	*Bryozoan*	*Brachiopod*	*Molluscs*	*Unknown skeletal*	*Ooids*	*Silt and clay*	*Granular cement*	*Rim cement*	*Unknown*
Foreset (average of 3 thin-sections)	45	11	5	1	7	1	6	4	17	2
Toeset (average of 3 thin-sections)	32	12	9	2	12	0	23	3	5	1
Bottomset (average of 3 thin-sections)	7	2	1	0	33	0	56	0	0	0

[1] Based on counts of 400 or more points on each of three thin-sections from each "set". The most important change to note here is the increase in silt-size and finer material from the foresets into the bottom sets. This material is mainly finer than coarse silt. The increase in unknown skeletal debris reflects the general decrease in grain size toward the bottomsets and the correlative difficulty of identifying small skeletal grains as to origin. The major part of the unknown skeletal category, however, is composed of bryozoan and brachiopod debris. Crinoidal material is relatively clear, single crystal, calcite whereas bryozoan and brachiopod debris is composed of very fine-grained fibrous calcite. The majority of the unknown skeletal grains show this fibrous microstructure but cannot be attributed to either bryozoans or brachiopods, Qualitatively, then, bryozoan and brachiopod debris increases from the foresets into the bottomsets, a fact which reflects the concentration of brachiopod and bryozoan debris in the finer sizes due probably to the control of the size of the breakdown product by skeletal microstructure.

to the general sorting decrease is the gradual loss of the discrete, individually well-sorted layers characteristic of the upper part of the exposed foreset. This loss is mainly controlled by an increasing abundance of organism burrows which disrupt and homogenize the layers.

The increase in burrowing is one of the characteristics which marks the transition from foreset to toeset beds. Evidence of toeset deposition appears at a distance of 60 ft. from the truncation edge. Sediment characteristics at this point are dominantly those of foreset deposition yet features typical of the toeset begin to appear. The main transition to toeset deposits starts at 70 ft. and is complete at 90 ft.

Toeset beds

Toeset characteristics begin to appear at about 60 ft. and are absent past 145 ft., though the main area of toeset deposition lies between 80 and 120 ft. As zone *1* thins through the toeset area, individual beds either pinch out totally or thin to highly argillaceous laminae a few millimeters thick. One bed which varies from 0.3 to 0.5 ft. in thickness persists into the bottomsets and forms the only traceable bottomset bed. Layers on either side of this bed in the toeset area are recognized in the bottomsets as thin, laminated, very argillaceous layers at top and bottom of the main bottomset bed.

The toeset deposits are characterized by thin beds composed of irregular, alternating layers of finer- and coarser-grained sediment (Fig.9). The coarser layers are generally fine- to medium-grained or medium- to coarse-grained skeletal limestones. Brachiopod valves and articulated shells, and large fragments of fenestrate and ramose bryozoans, 1 cm or more in length, are common in these layers. Crinoid plates up to 4 mm in diameter are also common. Average grain composition differs somewhat from the foreset beds in that crinoid grains are less abundant relative to brachiopod and bryozoan grains (Table I). These coarser-grained toeset beds are less sorted than similar sediments in the foresets and contain abundant very fine-grained skeletal debris in addition to the coarse fragments.

The composition of the fine-grained layers is more difficult to determine due to

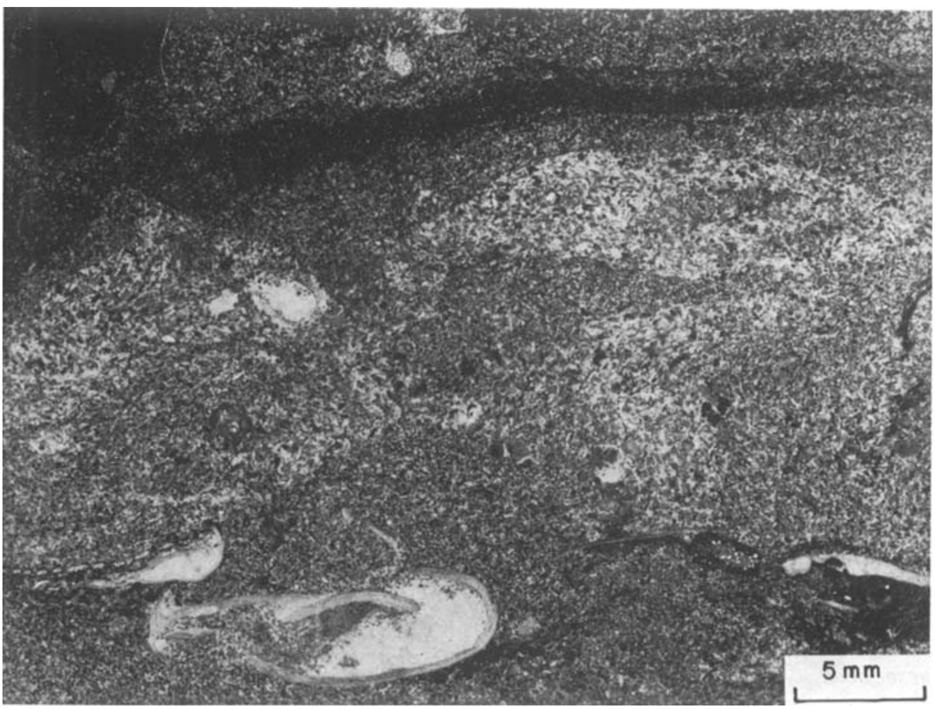

Fig.9. Toeset lithology. Fine- to coarse-grained argillaceous skeletal limestone. Note particularly the concentration of fossil material at the base of the coarser central layer, the poor sorting, the argillaceous layers, and the "swirly" pattern of grain orientation. Positive print from thin section.

Fig.10. Toeset lithology. Abundant fine-grained skeletal material is masked by dark argillaceous sediment. Note the poor sorting and the coarse skeletal material including a brachiopod shell filled with drusy calcite mosaic, a ramose bryozoan, and large crinoid grains. Light colored, angular, silt-size grains are detrital quartz.

the small particle size and the abundance of dark argillaceous material which tends to mask the character of other grains. In general, these finer-grained layers are composed of silt-size and very fine sand-size carbonate grains, angular silt-size quartz grains, and dark brown argillaceous material. The carbonate is dominantly of skeletal origin; skeletal microstructure can often be seen even where the skeletal type is unidentifiable. The quartz silt is identical to that present in the foreset beds. A thin section of a very poorly sorted toeset bed is shown in Fig.10. The relative increase in abundance of finer grain sizes is shown in Table II.

Both relative proportion of fine-grained sediment (silt-size carbonate, silt-size quartz, and argillaceous material) and the degree of mixing increase down the crossbed zone through the toeset area. Sorting thus becomes increasingly poorer toward the bottomset beds. Individual coarse, moderately well-sorted layers, common at the upper end of the toeset zone, become increasingly rare toward the bottomset area. One or two of these coarser layers persist nearly through the toeset area. One can be traced to 130 ft. where it becomes finer grained, more poorly sorted and slightly disrupted by burrowing. The lower limit of the toeset area in the broadest sense may be drawn where these coarser layers disappear. At 130 ft., the zone consists of a middle coarser layer 0.3 ft. thick which is underlain and overlain by thinly-laminated, highly argillaceous, layers each 0.1 ft. in thickness. The middle coarser layer grades laterally into poorly sorted argillaceous lime silt which has been extensively churned and homogenized. Though the composition of this middle layer tends to approach that of the

TABLE II

APPROXIMATE GRAIN SIZE VARIATION IN FORESETS, TOESETS AND BOTTOMSETS[1]

	Volume %		
	Fine sand and coarser	*Coarse silt to fine sand*	*Finer than coarse silt*
Foreset	84	8	8
Toeset	53	22	26
Bottomset	4	39	57

[1] Based on estimation of grain size in thin section, no correction factors applied. This table gives a general picture of the grain size variation along zone *1*. Figures based on measurement of 400 or more grains in each of three thin-sections for each set.

laminated layers above and below, it does maintain a lithologic distinctness to the limit of its traceability.

Bedding is, in general, more irregular in the toeset than in the foreset. Thinning of some layers to argillaceous laminae a few millimeters thick, and disappearance of other layers leads to a general irregularity of bedding. The increased effectiveness of burrowing organisms in this area causes mixing of discrete coarse and fine layers and general churning and homogenization of beds. Bedding planes or laminae are partially destroyed. Tracing individual layers or subzones (small packets of layers) through the toeset area is thus difficult.

Organic burrowing of two distinct types becomes evident in the toeset beds. One of these burrow types is represented by tubular, vertically or subvertically oriented burrows with sharply-defined walls; the other is shown only by the homogenization of sediment and by the swirly pattern of crudely aligned elongate fragments (Fig.11).

The sharply-defined burrows are 1–2 cm wide and up to several centimeters long. In most cases, these burrows originate in an argillaceous, laminated, lime-silt layer overlying a coarser, moderately well-sorted layer. The burrows are filled with argillaceous sediment, never with the coarser material. Where the burrow cuts across a coarser layer, no mixing at the edge is evident.

The second burrow type is evidenced mainly by the swirling pattern of oriented elongate skeletal fragments and small pieces of dark organic (?) matter. Rarely, the discrete shape and orientation of a burrow is shown by the swirling pattern. Pellets up to 4 mm in diameter and 6 mm in length, composed dominantly of silt-size and clay-size material, are frequently associated with burrows of this type. These pellets are always darker colored than the matrix in which they lie, which may reflect a concentration of organic material within the pellet. The shape, size, association with burrows, and the apparent organic content indicate a fecal origin. These pellets are never found closely associated with the sharply-defined burrows.

The swirly burrows occur in all three depositional areas; they are rare in the foresets, common in the toesets, and abundant in the bottomsets. The sharply-defined burrows, however, are absent in the foreset beds, abundant in the toesets and common in the bottomsets.

One of the striking differences between toesets and bottomsets is the relative abundance of fossils in each. Bryozoans and brachiopods are common in the toesets as concentrations in the coarser layers. Where the coarser layers disappear, however, so do concentrations of fossils. Articulated brachiopod shells filled with sparry calcite are scattered through the bottomsets but concentrations of fossils are essentially absent.

The transition from toeset to bottomset beds is thus marked by the gradual and finally complete disappearance of coarse layers and fossil concentrations, and by the decline in abundance of sharply defined burrows. These features begin to disappear at approximately 120–130 ft; the transition is complete between 145 and 160 ft.

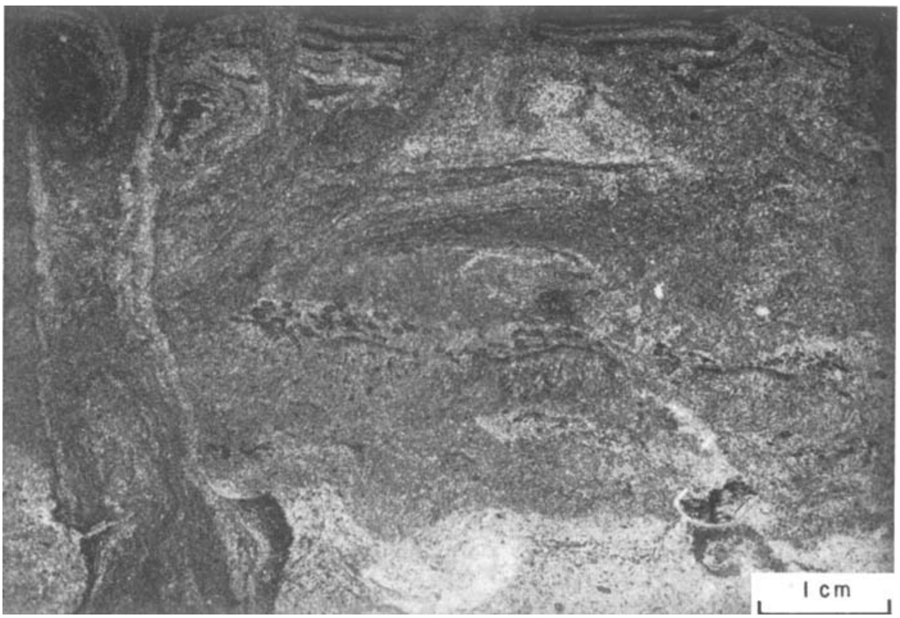

Fig.11. Positive peel print showing the two types of burrows present in the deposit. The vertical, sharp-walled burrow on the left is interpreted as evidence of the presence of suspension feeding organisms in the original sediment. The general "swirly" pattern of the sediment, and the disruption of laminae, are evidence of deposit-feeding organisms. These organisms, probably soft bodied, browsed through the sediment causing disruption of bedding, and general churning and homogenization of the sediment. Note that the sharp-walled burrow originates in a highly argillaceous laminated layer and is filled mainly with sediment similar to that at the top of the burrow. Note also the concentration of dark fecal pellets in the center of the photograph.

Bottomset beds

At the toeset–bottomset transition, zone *1* thins to a single bed of relatively constant thickness (0.3–0.5 ft.) which can be traced from the lower end of the toeset area to approximately 280 ft. from the truncated edge of zone *1*. Most cross-bed zones cannot be traced through the bottomset area due to thinning of beds to untraceable minor laminations, and partially due to loss of distinctiveness of the original major bedding planes.

Tracing of zones through the bottomset area is relatively unimportant, however, as few changes occur. Gradual thinning of bottomset beds due to attenuation of the distal portions is reflected in the gradual convergence of a cross-bed zone with the underlying discontinuity surface. For instance, between 140 and 255 ft., the section between the lower discontinuity and the bottom of zone *1* thins from 8.8 ft. to 5.7 ft. Changes other than this gradual thinning are slight.

Bottomset beds are mainly argillaceous skeletal limestones in which the carbonate fraction is mainly silt-size. The abundance of argillaceous material varies from layer to layer. More resistant calcareous beds up to 0.5 ft. in thickness alternate with varying numbers of thinner, recessive, highly argillaceous beds (Fig.12, 13). The latter beds generally are less than 0.1 ft. in thickness. The main components of the bottomset beds are silt-size carbonate, silt-size quartz, and argillaceous material (Fig.14). The carbonate is dominantly of skeletal origin; though the organism type most often cannot be identified due to the small size, most carbonate grains show some evidence of organic microstructure. Of those few grains which are identifiable, echinoderm frag-

Fig.12. View of bottomset beds. Note alternation of thicker, more resistant beds with recessive units.

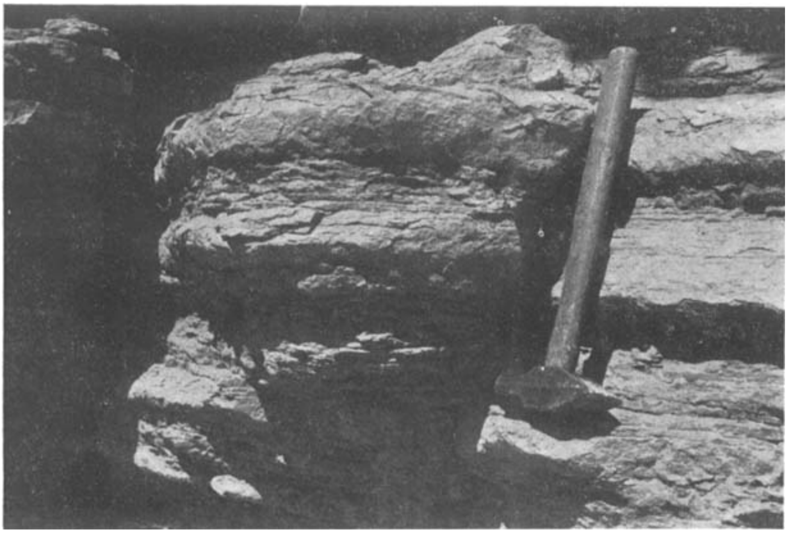

Fig.13. Close up of area of Fig.12 to show the thinly-bedded nature of the argillaceous, recessive intervals. The hammer is 14 inches long. The more resistant beds are coarser grained and more calcareous.

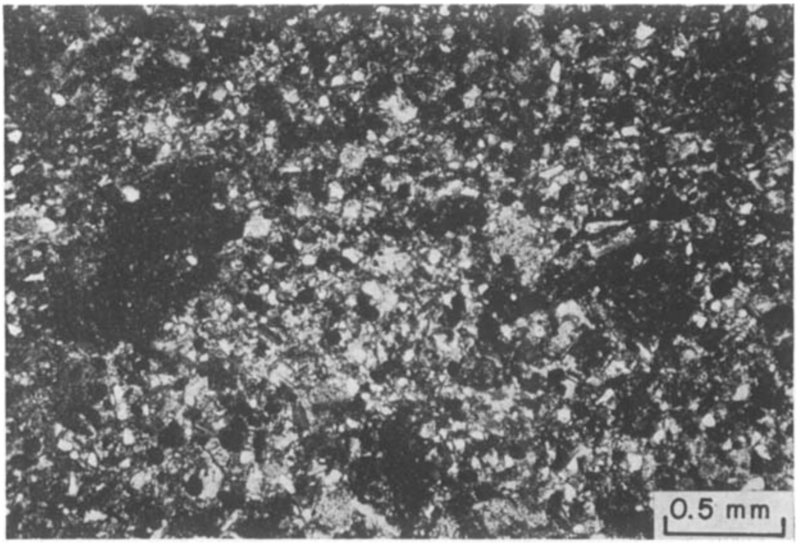

Fig.14. Bottomset lithology. Note fine grain size (mainly silt-size and very fine sand-size material), relative abundance of dark argillaceous sediment, abundance of quartz silt, and the presence of fecal pellets (large, dark grains at right and left center). Many of the apparently opaque grains in this sediment show skeletal microstructure when viewed through the microscope and are probably bryozoan and brachiopod debris. Light gray material is mainly crinoidal debris.

ments are the most abundant; bryozoan and brachiopod grains are common (Table I). Much of the silt- and clay-size material listed in Table I is also carbonate but is of unknown origin. Most of this material, however, probably owes its origin to the abrasion of skeletal carbonate.

Bottomset beds are sorted, containing little material coarser than very fine sand (Table II). Occasional medium to coarse sand-size skeletal fragments are present but do not contribute a significant percentage to the sediment. Whole brachiopod shells are found scattered through the bottomset beds. These generally lie parallel to bedding and are apparently in place or nearly so. Bryozoans are rare but do occur in the bottomsets; fenestrate bryozoan fronds are generally more complete than those found in the foresets or toesets indicating that they have been subject to less transport.

Bottomset sediments have been extensively churned by burrowing organisms. Homogenization of originally discrete layers is frequently shown by thin remnants of the original layered sediment in the midst of a highly churned bed (Fig.15). Sharply defined burrows are present in bottomset units but are significantly less frequent than in the toesets.

Fecal pellets similar to those described from toeset strata are present in nearly all samples from the bottomset units (Fig.15). In some areas, these pellets are randomly scattered through a bed; in other cases, when bottomset beds are split parallel to bedding, small mounds of these pellets are found along the bedding planes. These

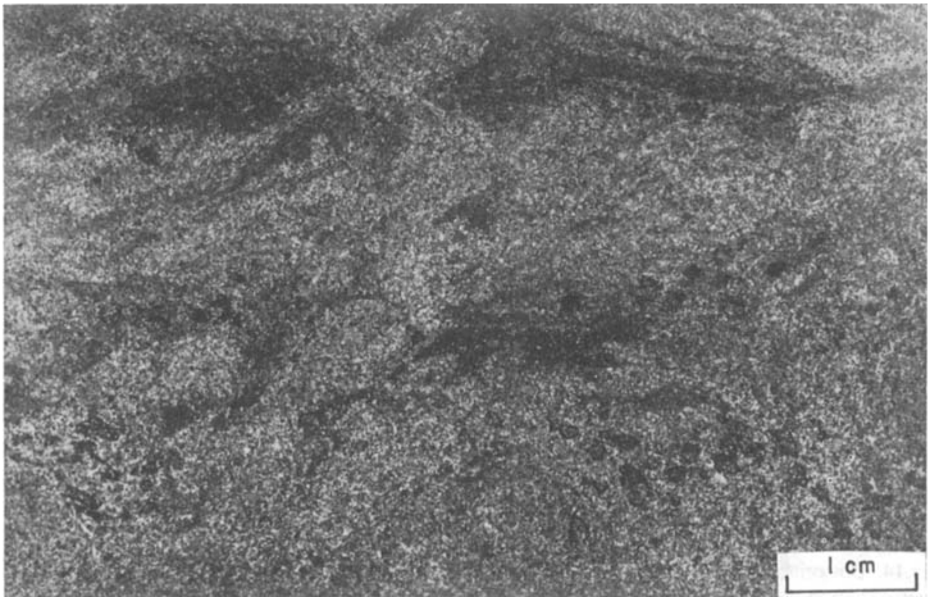

Fig.15. Positive peel print to show general churned nature of bottomset beds. Original laminated nature of argillaceous layers can be seen in upper right. Note also the abundance of fecal pellets in the bottom part of photograph, and the absence of well-defined burrows.

mounds are generally 2–3 cm in diameter and have relief of a few millimeters. Pellets up to 2 or 3 mm in length are crudely oriented in a radial pattern around the center of the mound. Similar mounds of fecal pellets can be observed at the sediment surface in areas of modern-day carbonate sedimentation.

One further interesting feature of these pellets is that they are always composed of dark-brown (organic-rich?) silt-size and clay-size material with only scattered grains of silt-size and fine sand-size quartz and skeletal carbonate (Fig.14). The composition is the same whether the pellets are found in coarser layers of the toesets or argillaceous layers of toesets or bottomsets. The organism which extruded these pellets was evidently a very discriminating one which chose to ingest only the finer-grained, more organic-rich, sediment.

INTERPRETATION

Field relationships of the deposit described above suggest that it was probably formed by aggradation to a profile of equilibrium controlled by a base level. Deposition of the Moorefield Formation was related to a pre-Moorefield erosion surface on which the relief commonly exceeds 100 ft. (SWINCHATT and CIRIACKS, in preparation). The exposure discussed here is immediately adjacent to a pre-Moorefield topographic high. In a roadcut at the western end of the outcrop, the upper part of the lower member of the Moorefield (Bayou Manard Member) is seen to lap onto a topographic high. This topographic high presumably was once a submarine ridge which may have acted as a partial barrier to circulation. Sediments transported across this barrier were deposited in a basin on the down-current side of the topographic high, forming the cross-bedded unit described above (Fig.16). As aggradation continued, the basin gradually filled, resulting in the regressive geometry seen in the present outcrop. Some time after deposition of the Lindsey Bridge Member, and prior to deposition of the

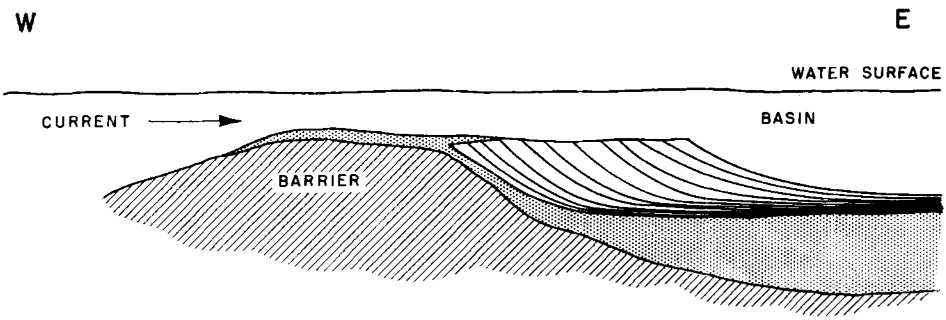

Fig.16. Diagrammatic representation of the inferred depositional geometry of the Lindsey Bridge Member. Sediment transported across the barrier was deposited on the down-current side forming the set of large-scale cross-beds seen on the present outcrop.

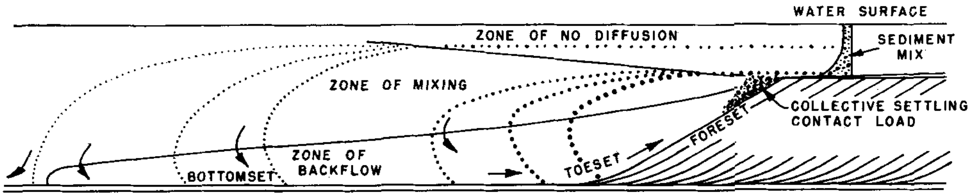

Fig.17. Diagrammatic representation of hydraulic geometry discussed here, modified from JOPLING (1965a and other papers). Idealized particle trajectories show that distance of transport downstream from the cross-bed lip depends on grain size (differential settling velocity) and height of transport above channel floor. Arrows indicate direction of water movement.

overlying Ordnance Plant Member, the upper part of the Lindsey Bridge Member was truncated by erosion.

The discussion presented below is an attempt to interpret the depositional mechanics of this deposit on the basis of principles derived from recent flume studies (ALLEN, 1965; JOPLING, 1963, 1964, 1965b; SIMONS and RICHARDSON, 1961; SIMONS et al., 1961). An investigation of this type should prove instructive in terms of the applicability of these principles to a natural occurrence which differs significantly in scale and component materials from the flume models. In addition, if the conditions of deposition of the outcrop example can be explained in terms of established principles, then increased understanding of associated minor sedimentary phenomena may result.

Characteristics of cross-beds (e.g., shape, lamination, and distribution of grain sizes) are in part controlled by the expansion of a current as it passes over a sharp discontinuity in the channel floor. The main features of the hydraulic geometry associated with this flow transition, described by JOPLING (1963, 1964), are shown diagrammatically in Fig.17. The main factor of interest to this discussion is the flow separation at the lip of the deposit and the maintenance of a discrete current for some distance downstream. Ultimately, mixing at the edge of the current causes diffusion and deceleration. In some cases, a reverse circulation or backflow of fluid into the toe area of the foreset and up the foreset slope may occur. The backflow may result in the formation of regressive ripples directed toward the foreset as described by JOPLING (1961) and ALLEN (1965) in experimental studies, and ALLEN and NARAYAN (1964) in a field occurrence. The interested reader is referred to JOPLING (1963, 1964) for a detailed explanation of this hydraulic geometry and its affect on bedding structures.

Cross-bed shape

The limestone deposit discussed here consists of large-scale, low-angle cross-beds with a tangential basal contact. On the outcrop, the basal contact appears to be smoothly concave-up, but detailed examination shows that this relationship is only apparent. A change in direction of maximum dip and a rather abrupt decrease in the dip angle coincident with the major foreset–toeset lithologic change indicate an abrupt change in slope between the sets, thus precluding a smoothly concave transition.

The tangential basal contact and the well-developed toeset and bottomset deposits indicate effective suspension transport. This is supported by the general distribution of grain sizes in the deposit (Table II). Bottomsets are composed almost entirely of silt-size and clay-size sediment and toesets have an extensive silt-size and clay-size component. In contrast, foresets are composed of fine to medium sand-size sediment. The almost total by-passing of the foresets by the fine-grained component indicates separation of sediment load into bed load and suspended load components.

JOPLING (1965b) has concluded, from experimental studies, that the development of a tangential contact between foreset and bottomset units is primarily dependent upon the amount of material transported in suspension. With an increase of suspension transport, a greater proportion of material is carried beyond the foreset slope and deposited in the toe sector forming a tangential contact. Further increase in suspension transport results in a decrease of foreset slope inclination and the ultimate establishment of a concave-up profile. The shape of the cross-bed set under discussion here thus can be attributed to well-developed suspension transport in a current which entered a still basin. Coarser material was deposited on the foreset slope, while finer material was carried farther into the basin to be deposited as toeset and bottomset units forming a tangential contact. An additional factor in the formation of a tangential contact is the depth ratio (the ratio of the depth of stream to depth of basin). In the present case, a moderate depth ratio must have existed in conjunction with suspension transport. A large depth ratio (basin shallow relative to current) would have produced a pronounced tangential contact (concave-up), while a small depth ratio (basin deep relative to current) would have resulted in an angular basal contact.

Relationship of grain size and sorting to hydraulic geometry

A general increase of fine-grained sediment from the foreset into the bottomset is shown by variation in the percent of total insoluble residue (Fig.18). The insolubles are composed of silt-size quartz, clay minerals (mainly illite) and fine-grained organic detritus. Increase in the abundance of insolubles can be taken to reflect an increase in the total abundance of finer grain sizes, i.e., including silt-size (and clay-size?) carbonate. Fig.18 shows a relative increase of insolubles through the foreset (up to about 70 ft), a general uniformity through the toeset area (70 ft.–ca. 140 ft.), and a further increase combined with greater variability through the bottomsets.

The increase of fines in the toeset and bottomset is related to flow separation and suspension transport. Details of the variation shown in Fig.18 can be attributed to three distinct processes: (*1*) settling out of progressively finer sizes away from the lip of the foreset; (*2*) collective settling of the contact load, and (*3*) variation of flow velocity and competence of the transporting current.

The first of these is essentially self-explanatory. Increased mixing and decrease

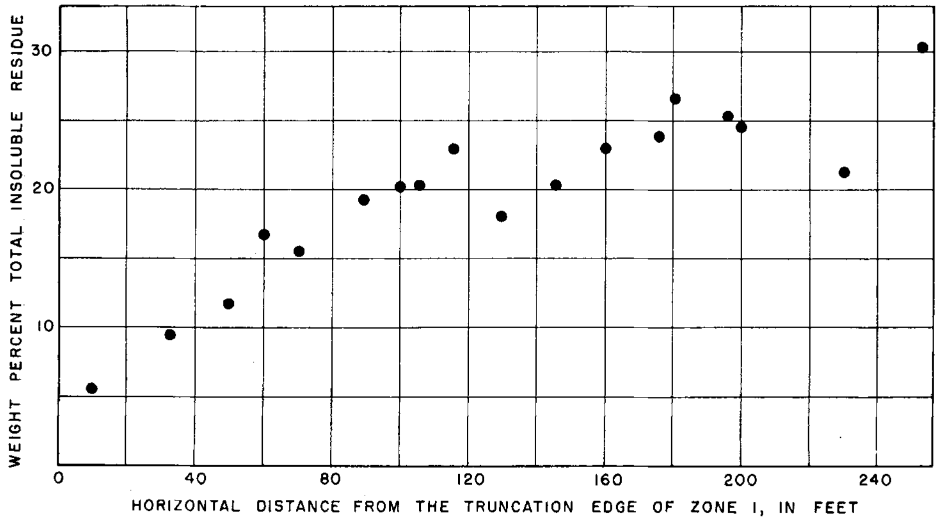

Fig.18. Variation of percent total insoluble residue along cross-bed zone *1*. Note increase through the foreset, uniformity in toeset, and increase plus greater variability in the bottomset beds. Rate of increase through bottomsets minimized due to relationship between outcrop trend and direction of maximum dip, i.e., distances along the outcrop are considerably greater than they would be in direction of maximum dip (depositional dip).

in flow velocity and competence downstream from the foreset lip result in settling of progressively finer fractions from suspension. Distance of transport into the toeset and bottomset area depends on differential settling velocity and the height of transport above the channel floor (JOPLING, 1965a; Fig.17).

The contact load includes both bed load and sediment carried in suspension. Collective settling of this "layer" over the lip of the foreset results in the deposition of most of the coarser material on the upper foreset slope. However, some of the sediment, including part of the suspended load (in this case silt-sizes), may be shifted downslope in "sediment-laden vortices" (JOPLING, 1965b). This material is deposited mainly in the toe sector, thereby increasing the concentration of fine sediment in the toeset and lower foreset.

The presence in the foresets of thin laminae composed of quartz and carbonate silt, and the irregular alternation of slightly coarser and finer layers in the foresets, indicate variation of flow velocity and competence of the transporting current. "Normal" or average flow conditions resulted in deposition of silt-size and finer material in the toeset and bottomset areas. The decrease in abundance of quartz silt laminae up the foresets is a reflection of the relative frequency and degree of deviation from normal flow conditions. Thus, only very infrequently were conditions such that quartz silt laminae were deposited on the upper reaches of the foreset slope. Less extreme deviation from normal flow conditions resulted in deposition of quartz silt laminae only a short distance up the foresets.

Distribution of quartz silt along the foreset may also be a reflection of normal pulses in sediment transport. JOPLING (1964) notes that even under steady-state conditions (e.g., uniform flow velocity) small variations of velocity are a natural counterpart of turbulent flow. In addition, the rapidly changing pattern of macroturbulence in the flow transition at the tip of the cross-bed adds to the variability of sediment transport across the tip at any moment of time. Jopling attributed the fine lamination seen in bottomset deposits to the effects of velocity pulsation and macroturbulence on the sediment mix. Laminations in the bottomsets of this field deposit may be caused in a similar way, as may the quartz silt laminations of the foresets. An alternative explanation would involve a pulsatory supply of quartz silt to the depositional area. The increase of quartz silt laminae toward the toeset, however, favors either variation in flow competence or normal pulses of turbulent flow.

The uniformity of percent insoluble residue exhibited by the toeset beds (Fig.18) may reflect the homogenizing affect of reverse circulation at the toe of the foreset. Reworking and mixing of sediment by the eddy system associated with the backflow in the toeset region may result in the uniform distribution of fine and coarse material through this area. The greater degree of variability of percent insolubles characteristic of the bottomset beds reflects the general lack of reworking by bottom currents. Sediment which settles out in the bottomset region is not subject to redistribution by bottom currents and thus reflects the vagaries of deposition of material from suspension. The increase of insolubles through the bottomset zone may indicate that the finer-grained sediment is dominantly silicate and organic debris rather than carbonate. Thus, the increase would reflect the gradual disappearance of coarser (silt-size) carbonate detritus rather than an actual increase in the abundance of insolubles. This relationship is also suggested by the decrease of bed thickness through the bottomsets.

Periodic increase in flow velocity and competence are demonstrated by the persistence into the bottomset units of resistant beds, 0.3–0.5 ft. thick, which generally contain a central zone of very fine sand-size skeletal debris (Fig.12, 13). This central layer may persist some tens of feet into the bottomset area. Beds of this type are generally bounded above and below by more argillaceous, thinly-bedded, laminated, calcisiltites typical of "normal" bottomsets.

Variation in sorting is also directly related to the hydraulic geometry of deposition. Both foreset and bottomset beds are moderately well-sorted, though the median size differs. Sorting of these units is controlled mainly by differential settling velocity. When the current expands over the foreset, coarser particles are dropped on the upper foreset slope while the finer grain sizes are carried into the basin. Finest grain sizes settle out in the bottomset area. Foresets are sorted because the fines are bypassed, whereas bottomsets are sorted because of the narrow size range of material available for deposition.

In contrast, toesets are poorly sorted and contain material ranging from clay-size to whole fossils a centimeter or more in diameter. The coarsest crinoidal material is also found in this area. As noted previously, deposition of toesets is complex. The contact load and the stream bed may be partly broken up by eddy action at the edge of the foreset and some of the entrained sediment may be transported collectively down the foreset by turbulent eddies. Deposition taking place in the toe sector thus partially reflects the composition of the contact load. Because the contact load comprises both bed load and suspended load components, the toeset is poorly sorted. In addition, the phenomenon of reverse circulation leads to an increase in the size range of the toeset sediment. Transport of bottomset material into the toe sector by reverse circulation adds fine-grained sediment to the toeset, while scouring of the lower foreset by eddies associated with reverse circulation contributes coarser grained material. The influence of scouring of the lower foreset is accentuated by the tendency for the coarser particles of the bed load to accumulate near the base of the foreset (JOPLING, 1964). Furthermore, scouring of the entire foreset slope by complex cross currents may result in addition of coarser foreset sediment to the toeset deposit.

Concentration of coarse fossil materials (brachiopod shells and large fragments of both ramose and fenestrate bryozoans) in the toeset deposits is also attributed to the mechanism of formation of the toeset. Association of these fossil materials with coarser toeset layers, concentration in thin layers, and orientation, indicate transport rather than in situ deposition. MENARD and BOUCOT (1951) have shown that brachiopod shells filled with water are highly buoyant and may be transported more readily than fine and medium sand-size grains of the substrate[1]. Bryozoans might be buoyed up similarly by water filling all open space within the skeleton. These shells and large skeletal grains, transported as bed load, would settle collectively with the contact load on the upper part of the foreset. Transport of these relatively buoyant grains down the foreset slope and into the area of toeset deposition can probably be attributed to the "sediment-laden vortices" described by JOPLING (1965b). Concentration of fossils near the bottom of well-defined coarser-grained beds may indicate transport of shell accumulations from the topsets during periods of increased flow velocity and competence such as during periodic storms. Shells found scattered in the bottomsets are apparently in place, as shown by preservation and geopetal criteria.

Relationship of burrow distribution to the inferred depositional geometry

The two types of burrows described in the first part of this paper are interpreted as indicating the effects of two distinct groups of burrowers. These burrows are of particular interest in that their distribution reflects a control by the hydraulic regime.

Burrows with sharply defined walls (Fig.11) oriented vertically or subvertically are interpreted as having been formed by burrowing suspension feeders such as pele-

[1] Entrapment of air would obviously create an even greater buoyant effect. It is highly unlikely, however, that the brachiopod shells were air tight, particularly in light of the fact that they were ultimately filled with internal sediment and/or precipitated calcite.

cypods. These organisms need a well-established connection with the sediment–water interface in order to feed, thus necessitating sturdy-walled burrows. Connection with the sediment surface is also indicated by the nature of the burrow fill; no matter what type of sediment the burrow cuts across, the fill is always similar to the sediment at the top of the burrow. These sharply defined burrows are limited to toeset and bottomset beds; they are abundant in the toesets and less abundant in the bottomsets, particularly in the distal parts.

Burrowing indicated only by "swirly" orientation of elongate particles and small fragments of organic (?) material (Fig.11) is evidence of deposit feeders. These organisms browse through the sediment and extract organic nutrients, destroy stratification and produce their characteristic structures while homogenizing the deposit. This type of burrowing is present in all three "sets" but is most characteristic of bottomset beds and rare in the foresets.

Distribution of these burrow types can be explained as follows: suspension feeders could not survive in the foresets due to the instability of the more steeply dipping sediment. Frequent slip of sediment would continuously clog and damage the well-defined burrows thus disturbing the necessary connection to the sediment surface. Deposit feeders would be considerably less abundant in the foresets due simply to the lack of organic matter in the well sorted fine- to medium-grained sands. The lack of fixed burrows in the foresets is not due to destruction by post-burrowing effects such as compaction. Sediment laminae preserved in these beds indicate that the foresets contained only a sparse burrowing fauna composed entirely of deposit feeders.

Distribution of suspension feeders in toesets and bottomsets may have been a function of the relative amount of organic material delivered to a particular area. Continuous supply of organic nutrients to the toeset area is facilitated by eddies rich in contact-load material, by normal settling from suspension, and by transport of material into the toe sector by reverse circulation. This constant supply of organic detritus accounts for the highest concentration of suspension feeders in the toeset area. The bottomset is supplied with sediment by material settling from suspension and to a much lesser extent by collective settling of the contact load (JOPLING, 1965b). A large amount of organic material would be delivered to the bottomsets by these mechanisms. Burrowing suspension feeders, however, depend on currents to transport this organic matter past the mouth of the burrow. Bottomsets, supplied mainly by material settling from the expanding current (Fig.17), show little evidence of bottom traction and current movement, accounting for the lesser abundance of burrowing suspension feeders. In addition, or alternatively, decrease in bottom circulation may have been accompanied by a minor decrease in aeration or other chemical conditions which were toxic for the suspension feeders.

Evidence of deposit feeders is abundant in both toeset and bottomset beds. Organic matter, reacting to transport conditions in a manner similar to silt-size and clay-size mineral detritus, accumulated in toeset and bottomset areas. Fine sediment and, therefore, the associated organic detritus, is proportionally more abundant in the bottomsets[1]. A larger population of deposit feeders in the bottomsets, supported by the greater availability of organic matter, may have resulted in the more extensive reworking of these beds.

The typical occurrence of suspension-feeder burrows supports the thesis that the persistent coarser-grained layers which occur periodically in the bottomsets represent relatively rapid deposition during times of increased flow velocity and competence. All the well-defined burrows observed during this study originate in the finely-laminated, more argillaceous layers which overlie the coarser-grained persistent beds. The burrows generally cut through the laminated layers and either end within or cut through the coarser layers. In addition, these burrows are invariably filled with the type of sediment found in the thinly-laminated, more argillaceous layers. The contention here is that the coarser layers represent times of relatively rapid deposition whereas the argillaceous laminae represent times of relatively slow "normal" deposition. It was only during these intervening periods of slower sedimentation that the suspension feeders could establish themselves.

The fecal pellets described previously are associated with deposit-feeder burrowing in the toesets but become more abundant in the bottomset beds, particularly in the distal parts. Their increase through the bottomsets, then, reflects the general increase of the effects of deposit feeders. In addition, small mounds and concentrations of these pellets found in the distal bottomsets indicate the lack of current reworking.

SIMILARITY OF VERTICAL AND LATERAL SEQUENCE

The general regressive character of the unit provides a compact example of the similarity between a lateral facies change and vertical variation (Fig.3). Starting in the upper massively cross-bedded section, a vertical traverse shows the following: (*1*) a unit of thick, dipping beds of fine- to medium-grained, well-sorted, skeletal limestone (grainstone); (*2*) a unit of more thinly-bedded, poorly-sorted, argillaceous skeletal limestone (packstone); and (*3*) a lower unit of very thin-bedded, highly argillaceous, sorted, skeletal limestone (packstone). Moving laterally along the outcrop reveals a similar sequence of lithologic changes, though the transitions may be more gradual. Recognition of vertical and/or lateral changes of this general type on a larger scale are often indicative of regressive deposition, though the details may differ. In any depositional situation in which sediment is progressively filling a basin, changes of this type may be expected.

[1] Measurement of total organic carbon by Roger Ames indicates that organic matter is, in fact, more abundant in the bottomsets than elsewhere in the deposit.

CONCLUSIONS

Formation of large-scale cross-bedding in the carbonate deposit described above has been attributed to a combination of two factors: (*1*) segregation of transported sediment into distinct bed load and suspended load components, and (*2*) the hydraulic geometry associated with flow separation over a channel-floor discontinuity. The resultant deposition of coarse sediment on the foreset slope and selective transport of finer grain sizes farther into the depositional basin caused the development of tangential cross-beds and the distribution of grain sizes seen in the deposit. Minor factors, including pulsatory sediment transport associated with turbulent flow, and reverse circulation caused by flow separation, modified the basic pattern. Thus, sedimentary attributes such as cross-bed shape, lithologic differences between foresets, toesets, and bottomsets, variation in grain size and sorting along a cross-bed zone, and other features, are generally compatible with principles derived from recent flume studies. In addition, minor associated sedimentary features, such as the distribution of fossils, the occurrence and variable abundance of different types of organism burrows, and the distribution of fecal pellets can be related to the hydraulic and depositional characteristics of the cross-bed set in which they occur. Distribution of these features is directly affected by factors such as the relative stability of the substrate, the amount of organic material deposited, and the presence or absence of bottom currents, as well as the general processes of transport and sorting which controlled the over-all structure of the deposit. That the flume-derived principles can be applied in spite of a significant scale difference (approximately 50 : 1) and a distinct contrast in sedimentary materials is of significance to future field studies of the formation of cross-beds.

ACKNOWLEDGEMENTS

This study was completed while I was employed by Pan American Petroleum Company. I am grateful to R. L. Ames and K. W. Ciriacks of Pan American for assistance in the field, for discussions of many facets of the study, and for critical reading of the original manuscript. Permission to publish was given by Pan American.

I also wish to thank Alan Jopling of Harvard University and Vinton Gwinn of Louisiana State University whose constructive comments led to improvement of the manuscript.

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