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FLUVIAL, TIDAL AND STORM SEDIMENTATION IN THE CHILHOWEE GROUP (LOWER CAMBRIAN), NORTHEASTERN TENNESSEE

A Thesis

Presented for the

Master of Science

Degree

The University of Tennessee, Knoxville

Mary R. Cudzil

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ABSTRACT

The Lower Cambrian Chilhowee Group of northeastern Tennessee consists of the Unicoi, Hampton and Erwin Formations and is divided into four facies. Facies G occurs only within the lower 200 m of measured section (the Unicoi Formation) and consists of fine-grained to pebbly quartz wacke with rare thin beds of laminated siltstone. Subfacies Gh consists of low-angle to horizontally laminated, fine-grained sandstone with laminae and lenses of granules and pebbles. Subfacies Gh represents upper flow-regime, overbank deposition within a braided stream system that was proximal to a coastline. Subfacies Gmr consists of medium-scale, planar-tabular cross-stratified conglomerate in which megaripple bedforms are preserved. It is interpreted to represent deposition in interbar pools of braided channels, as flood stage waned and larger bedforms ceased to migrate. Subfacies Gp consists of large-scale, planar-tabular cross-stratified sandy conglomerate; the large sets represent migration of large transverse bars within a broad braided stream channel during high flood stage.

Facies S occurs throughout the Chilhowee Group, and is therefore interbedded with the three other facies. Subfacies Sfl consists of thinly interbedded, laminated siltstone and sandstone that may exhibit wavy or lenticular bedding. It represents deposition during slack water periods between ebb and flood tides. Both Subfacies Sls and Sms consist of medium- to very coarse-grained, subarkosic to arkosic arenite. Subfacies Sls is characterized by large-scale planar-tabular and trough

cross-stratification,whereas Sms is characterized by medium-scale cross-stratification. Subfacies Sfl generally drapes the cross-stratified beds of both Subfacies Sms and Sls.

Subfacies Sls is interpreted to have been deposited within the deepest areas of a subtidal channel and Sms represents deposition in shallower water on shoals separating channels. Subfacies Sls and Sms are arranged in a thinning-upward sequence that most likely resulted from the longshore migration of channels and shoals.

Facies Shcs occurs only in the Erwin Formation and consists of horizontally laminated to hummocky, fine-grained arkosic to subarkosic arenite interbedded with equal amounts of bioturbated mudstone. It represents deposition between storm and fairweather wave base.

Facies QA is characterized by an absence of fine-grained units and lithologically consists of a supermature, medium- to coarse-grained quartz arenite. Subfacies QA-Unicoi exhibits large-scale planar-tabular cross-stratification and abundant low-angle cross-stratification; symmetrical ripples are rare. Subfacies QA-Unicoi is interpreted to have been deposited within either a ridge-and-runnel system or a system of nearshore bars.

Subfacies QA-Erwin exhibits large-scale, planar-tabular cross-stratification and forms the top of two 40 m thick coarsening-upward sequences: Facies Shcs → Facies S → Subfacies QA-Erwin. Subfacies QA-Erwin probably represents deposition on sand ridges that formed on a sand-starved shelf as transgression caused the detachment and reworking of shoreface channel-shoal couplets.

Paleocurrent data for the Chilhowee Group are unimodal but widely dispersed from 0°-180°, and exhibit a minor mode to the west. The data are interpreted to reflect the influence of longshore, tidal and storm currents. The ichnofossil assemblage changes upsection from one characterized only by Paleophycus to a Skolithos ichnofacies and finally to a Cruziana ichnofacies. The facies sequence, biogenic and paleocurrent data reflect the interaction through time of (1) nonmarine and marine processes; and (2) transgression coupled with shoreline progradation. The Chilhowee Group represents an overall deepening from terrestrial deposition to a marine shoreface that experienced both longshore and tidal currents, and finally to a storm shelf environment that periodically shoaled upward.

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I. INTRODUCTION

Purpose

The Chilhowee Group is a 641-2288 m thick sequence of interbedded conglomerate, quartzite, siltstone and shale that crops out in a northeast-southwest trending belt along the northwestern margin of the Blue Ridge Province from Georgia to Pennsylvania (Mack, 1980; Schwab, 1972). Previous stratigraphic and petrologic studies of the Chilhowee Group have led to tentative interpretations of provenance and depositional environments (Whisonant, 1974; Schwab, 1972). The purpose of this study is to develop a depositional model for the Lower Cambrian Chilhowee Group as exposed along the Doe River of northeastern Tennessee based upon interpretations of the lithologies, paleocurrents and physical and biogenic structures.

This study is significant for three reasons. First, the Chilhowee Group is devoid of body fossils, and marine trace fossils are only present in the upper part of the group. Therefore, in the general absence of paleontologic data, the distinction between rocks deposited in a shallow-marine versus a nonmarine setting must be based upon physical sedimentary structures.

Nonmarine terrigenous sediment and sedimentary rocks generally are characterized by their immaturity or submaturity. However, a fluvial system draining a sedimentary sourceland (or a sourceland experiencing intense weathering) may deposit texturally and compositionally mature sediments. Shallow-marine sediment is typically more mature and

exhibits better sorting and rounding as well as an absence of labile grains. Because lithology alone may not be definitive indicator of marine or nonmarine environments, a suite of diagnostic physical structures particular to each setting is used here to interpret these unfossiliferous Cambrian rocks.

Second, this study will contribute towards a better understanding of regional patterns of Cambrian terrigenous deposition. The Chilhowee Group of east Tennessee (Whisonant, 1974), Virginia (Schwab, 1972) and Pennsylvania (Goodwin and Anderson, 1974) was deposited along the craton-margin. Variations in thickness along strike indicate that the margin was probably subsiding at different rates.

This study is a first attempt at recognizing the facies that characterize the Chilhowee Group by examining a single, well-exposed outcrop. This study should serve as a facies reference section for comparison with other localities and thus help to define sedimentation patterns at the craton-margin. From a broader persepective, this study will yield information on craton-margin sedimentation processes, which differed from processes operating in the Upper Cambrian epeiric seas (i.e., craton-interior) (Byers and Dott, 1981 and 1982).

Finally, because individual exposures of the Chilhowee Group are typically separated from one another by thrust faults, correlation between exposures, based on lithology alone, is difficult. However, a firm understanding of the facies and facies sequences present will aid in understanding how rocks of individual thrust slices relate to each other.

Previous Work and Regional Stratigraphy

The Chilhowee Group was named by Safford (1856, fide. King and Ferguson, 1960) from exposures on Chilhowee Mountain in central-east Tennessee. Formations within the group were named by Keith (1895) on the basis of interbedding of sandstones with shaley units. The Chilhowee Group has subsequently been studied at various localities along its strike belt (Fig. 1 and Table 1). The exposure at Walland, Tennessee, was studied by Swingle (1949) who examined the heavy mineral suites of the sandstone cropping out in the Little River Gorge. Neuman and Nelson (1965) mapped and described the type Chilhowee exposed on Chilhowee Mountain. To the southwest, the Chilhowee on Starr Mountain was mapped by Phillips (1952), and the Chilhowee exposed on Bean Mountain was mapped and described by Rackley (1951). Still farther southwest, Mack (1980) described the petrology and stratigraphy of Chilhowee equivalents exposed in Georgia and Alabama. There, however, the Chilhowee Groupoccurs as klippen and no complete sections are exposed.

To the northeast of Chilhowee Mountain, at the Doe River Gorge and Bald Mountains of Greene, Cocke and Carter Counties, Bearce (1966) mapped the structure and stratigraphy of the Chilhowee Group and the Precambrian Ocoee Group. Whisonant (1974) studied the petrography and paleocurrents of the Chilhowee Group in eastern Tennessee. Farther northeast, Schwab (1970, 1971, 1972) studied the petrography and sedimentology of the Chilhowee Group in Virginia and Maryland (Fig. 1). In Pennsylvania, Goodwin and Anderson (1974) described the facies present

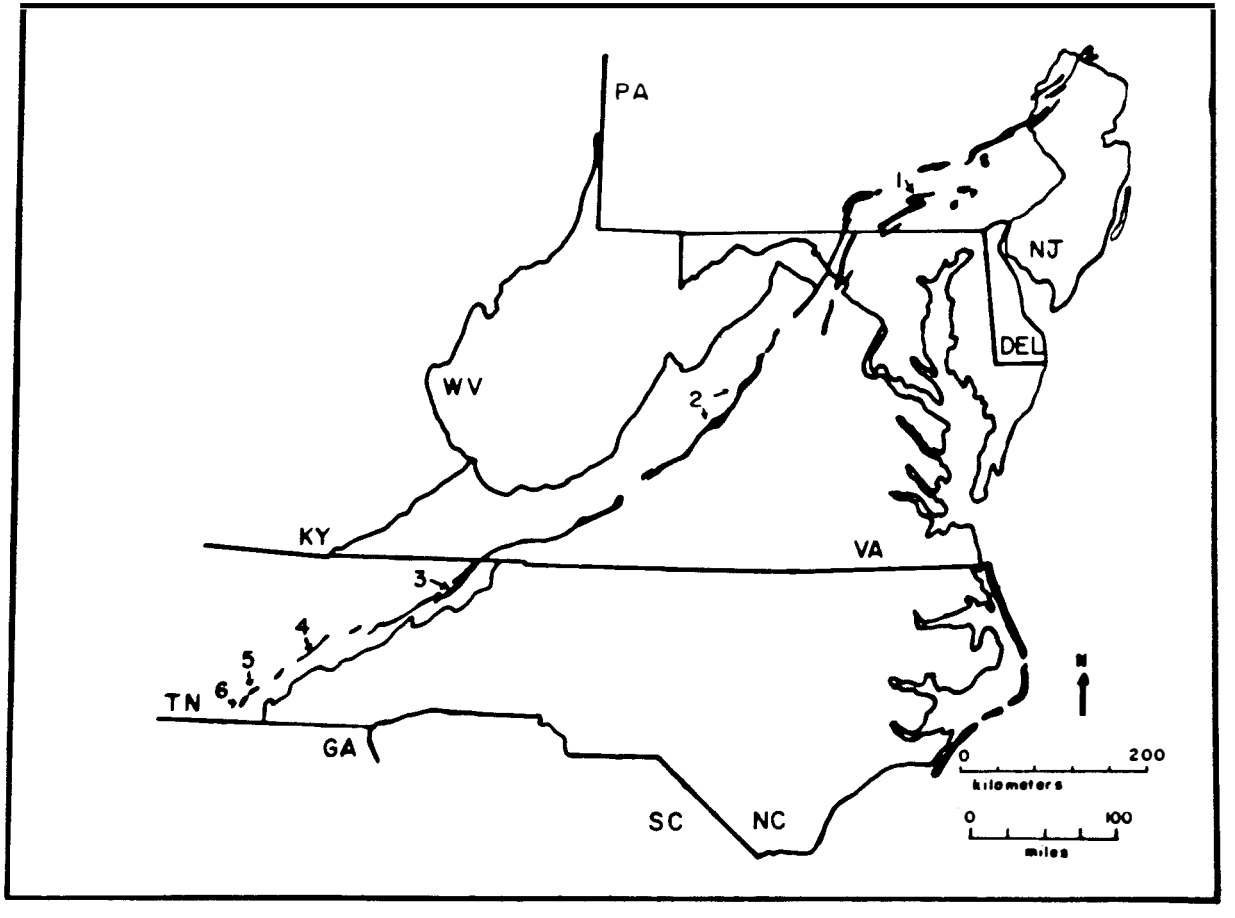


Figure 1. Distribution of Chilhowee Group outcrops in Tennessee, Virginia, Maryland and Pennsylvania. Arrows point to localities that have been studied: (1) Goodwin and Anderson (1974); (2) Schwab (1970, 1971); (3) Doe River Gorge (King and Ferguson, 1960; Bearce, 1966; this study); (4) Chilhowee Mountain and at Walland (Swingle, 1949; Neuman and Nelson, 1965); (5) Bean Mountain (Rackley, 1951); (6) Starr Mountain (Phillips, 1952); (modified from Schwab, 1972).

TABLE 1. Chilhowee equivalents in Virginia, Tennessee, Georgia and Alabama (King and Ferguson, 1960; Mack, 1980).

AGE		VIRGINIA	NORTHEAST TENNESSEE	EAST TENNESSEE	NORTH GEORGIA AND ALABAMA
EARLY CAMBRIAN	CHILHOWEE GROUP	Tomstown Dolomite	Shady Dolomite	Shady Dolomite	Shady Dolomite
		Antietam Quartzite	Erwin Formation (Helenmode Member at top)	Hesse Quartzite Murray Shale Nebo Quartzite	Weisner Formation Wilson Ridge Formation
		Harpers Formation	Hampton Formation (Cardens Bluff Shale Member at base)	Nichols Shale	Nichols Shale
		Weverton Quartzite Loudoun Formation	Unicoi Formation	Cochran Formation	Cochran Formation
PRECAMBRIAN		Catoctin Greenstone or Swift Run Formation or Injection complex	Mt. Rodgers Volcanic Group or Basement rocks	Ocoee Group	Base of section always faulted out.

in the Lower Cambrian Chickies Quartzite, which occurs within the same strike belt as the Chilhowee Group (Schwab, 1972). Although it may appear that much work has been done on the Chilhowee Group, it has either consisted of stratigraphic mapping or petrographic analyses. The paleoenvironmental interpretations arising from these previous studies have been imprecise and inadequate.

The regional stratigraphic terminology for the Chilhowee Group is summarized in Table 1, and although minor variations in the stratigraphic nomenclature exist along strike, there is at least a general sequence of interbedded quartzite and siltstone that is recognizable. The Chilhowee Group consists of terrigenous clastics that either: (1) unconformably overlie Precambrian basement rocks, or (2) conformably overlie the Mount Rodgers Volcanic Group (northeastern Tennessee) or the Ocoee Group (eastern Tennessee). Its upper contact with the Shady Dolomite (Tomstown Dolomite in Virginia) is conformable across the strike belt. All the formational and member units of the Chilhowee Group thicken and thin laterally along strike and the Group, as a whole, is the thickest (2288 m) in northeastern Tennessee and southwestern Virginia. It thins to 641-885 m in northern Virginia and eastern Tennessee (Schwab, 1972).

In northeastern Tennessee (the location of this study), the Group is divided into three formations; the basal Unicoi Formation, the Hampton Shale, and the uppermost Erwin Formation. The Helenmode Member of the Erwin Formation represents the transition from Lower Cambrian

terrigenous clastic sedimentation to Lower Paleozoic carbonate sedimentation that begins with the Shady Dolomite.

The Unicoi Formation is divisible into informal upper and lower units. The lower unit contains lenticular bodies of amygdaloidal basalt. Along Rt. 421 near Mountain City, Tennessee, some of the basalt flows are interbedded with arkosic conglomerate and sandstone. The upper unit, in contrast, contains no basalt and is not so predominantly coarse-grained. Quartz arenite also occurs in the thick sequences within the upper unit.

The Hampton Formation was described by King and Ferguson (1960) and is composed of interbedded clay shale, siltstone, arkosic sandstone and quartz arenite. In the Iron and Holston Mountain areas, the base of the Hampton is marked by 30.5-61.0 m of interbedded shale and siltstone. This unit is distinctive enough in this area to warrant designation as the Cardens Bluff Shale Member; however, it pinches out to the northeast and is not recognizable elsewhere.

The Erwin Formation consists of interbedded white quartzite, ferruginous quartzite, siltstone and shale. The Erwin differs from the Hampton only by degree; the Erwin contains more quartzite beds than the Hampton and fewer fine-grained units. The base of the Erwin is therefore arbitrarily placed beneath widely traceable quartzite beds that contain Skolithos.

Methods

The exposures of the Chilhowee Group that are the subject of this study are located in northeastern Tennessee along U.S. Route 19E as it

passes through Gap Creek and the Iron Mountains along the Doe River (Fig. 2). The Chilhowee here is exposed in a broadly folded and nearly complete section, except that the lower unit of the Unicoi Formation has been faulted out along the Iron Mountain Fault. The section under study is part of the southeast limb of a northeast-southwest trending syncline within the Shady Valley thrust sheet. Bedding therefore dips consistently to the northwest at about 45-50° and strikes approximately N60°E. Of the 996 m of section present, greater than half is covered by vegetation or lies in the riverbed. Only 362 meters of section is exposed as roadcuts and these are the rocks under study. Exposed and covered intervals are marked on the geologic map and the generalized stratigraphic column (Fig. 2 and Fig. 3).

This study involved measuring the exposures at a centimeter scale and describing lithologies, physical and biogenic sedimentary structures, and paleocurrent features. Thirty thin-sections were made of various lithologies to supplement hand sample observations. Petrography was not the intended focus of this study; therefore no point-counting was done. Modal percentages were visually estimated and lithologies were named according to Folk (1968) and Dott (1964).

There are four important reasons for looking at the thin-sections at least in a qualitative manner.

1. The detrital mineralogy and composition of the rock fragments hold information on the source terrain for the Chilhowee Group.

2. In some lithologies it is important to ascertain the relative proportions of depositional matrix, pseudomatrix resulting from

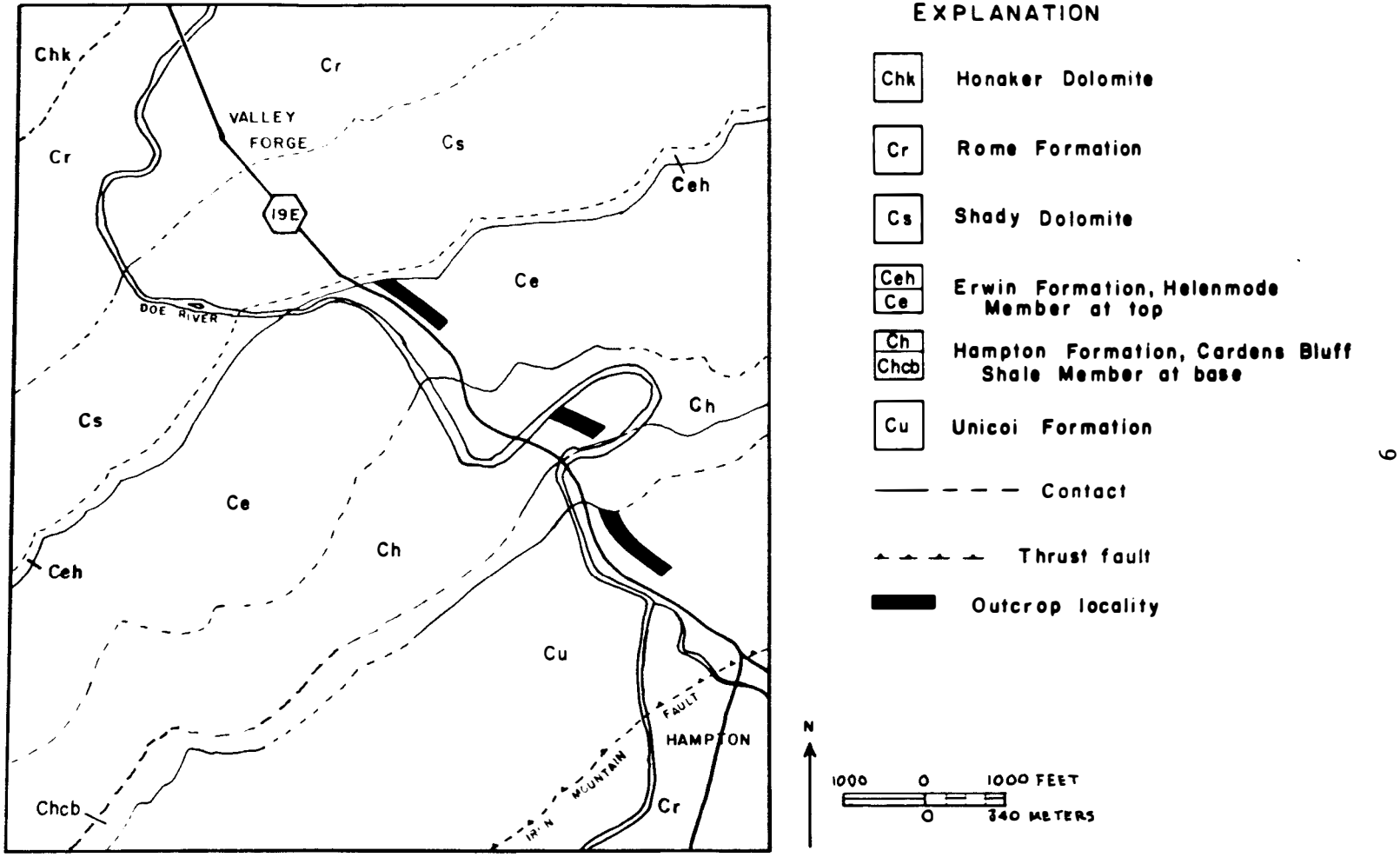


Figure 2. Location of the study area and local geology (modified from King and Ferguson, 1960).

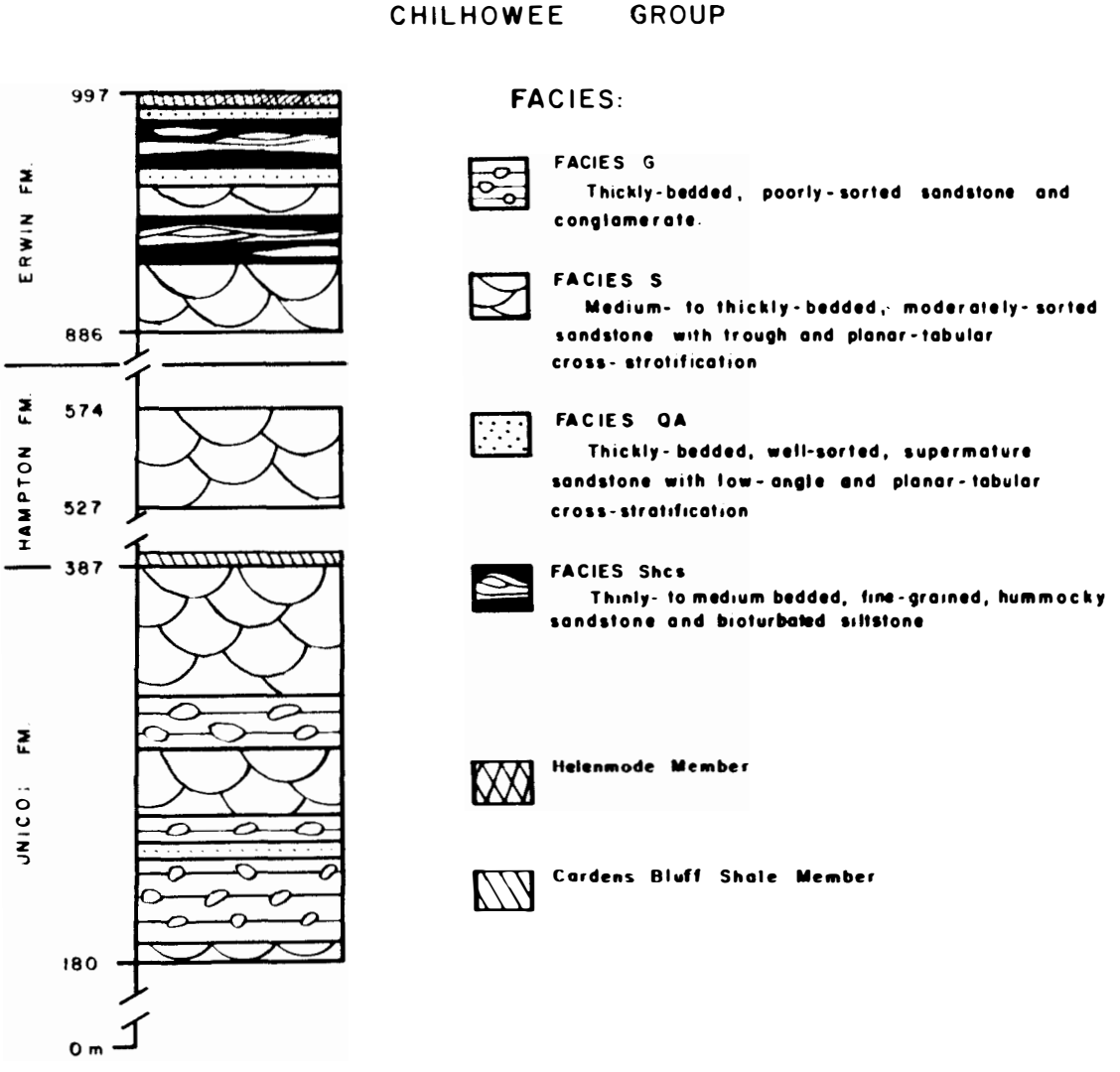


Figure 3. Generalized Stratigraphic column showing covered intervals and distribution of facies.

diagenetic alteration of sedimentary rock fragments, and types of cement (diagenetic clay, silica, calcite).

3. Petrography is useful for identifying the presence of second-cycle grains.

4. Because pressure solution and silica cementation are so pervasive in the arenite lithologies, initial grain boundaries are all but obscured in hand sample; therefore, degree of rounding, grain size and sorting are best seen in thin-section.

Paleocurrent measurements were taken from 221 directional structures including foresets of planar-tabular and trough cross-beds, low-angle cross-beds, symmetrical ripple crests, parting lineations, and tool marks. All measurements were restored to their original horizontal position by stereonet rotation (Ragan, 1973). Restored data were grouped according to facies and formation; rose diagrams of each facies as well as a composite rose diagram of the Chilhowee Group were drawn.

Statistical parameters such as vector mean, dispersion, and angular deviation were calculated using the vector summation method described by Curray (1956) as adapted for the computer in a program developed by Maher and Dott (1979). An F-statistic was calculated for each group by comparing the group's variance to the variance of a uniform circular distribution in order to test whether the data were significantly oriented.

II. SEDIMENTARY FACIES

Introduction

The exposures of the Chilhowee Group along the Doe River are probably the best in northeastern Tennessee. However, the great vertical extent and poor lateral exposure and accessibility of these rocks require that facies analyses be based principally upon Walther's Law, which states the facies now superimposed must at one time have been laterally adjacent (Reading, 1978). Because only one locality was studied, the author was not able to determine the lateral extent of and relationships between facies. Therefore, the main focus of this study was the pattern of vertical relationships between facies and vertical variations within a single facies. Vertical variations within facies and subfacies are discussed in this section along with the facies descriptions. Vertical facies relationships are also described in this section, but their environmental significance is discussed in the following section containing interpretations.

The Chilhowee Group can be divided into four broad lithofacies:

1. G: cross-stratified, arkosic lithic wacke;
2. S: cross-stratified subarkosic arenite;
3. Shcs: hummocky cross-stratified subarkosic arenite;
4. QA: supermature cross-stratified quartz arenite.

Facies G and S are each further divided into three subfacies based upon the sequence of sedimentary structures present. The facies and subfacies are illustrated in Figures 4, 5 and 6.

KEY TO LITHOLOGIC SYMBOLS

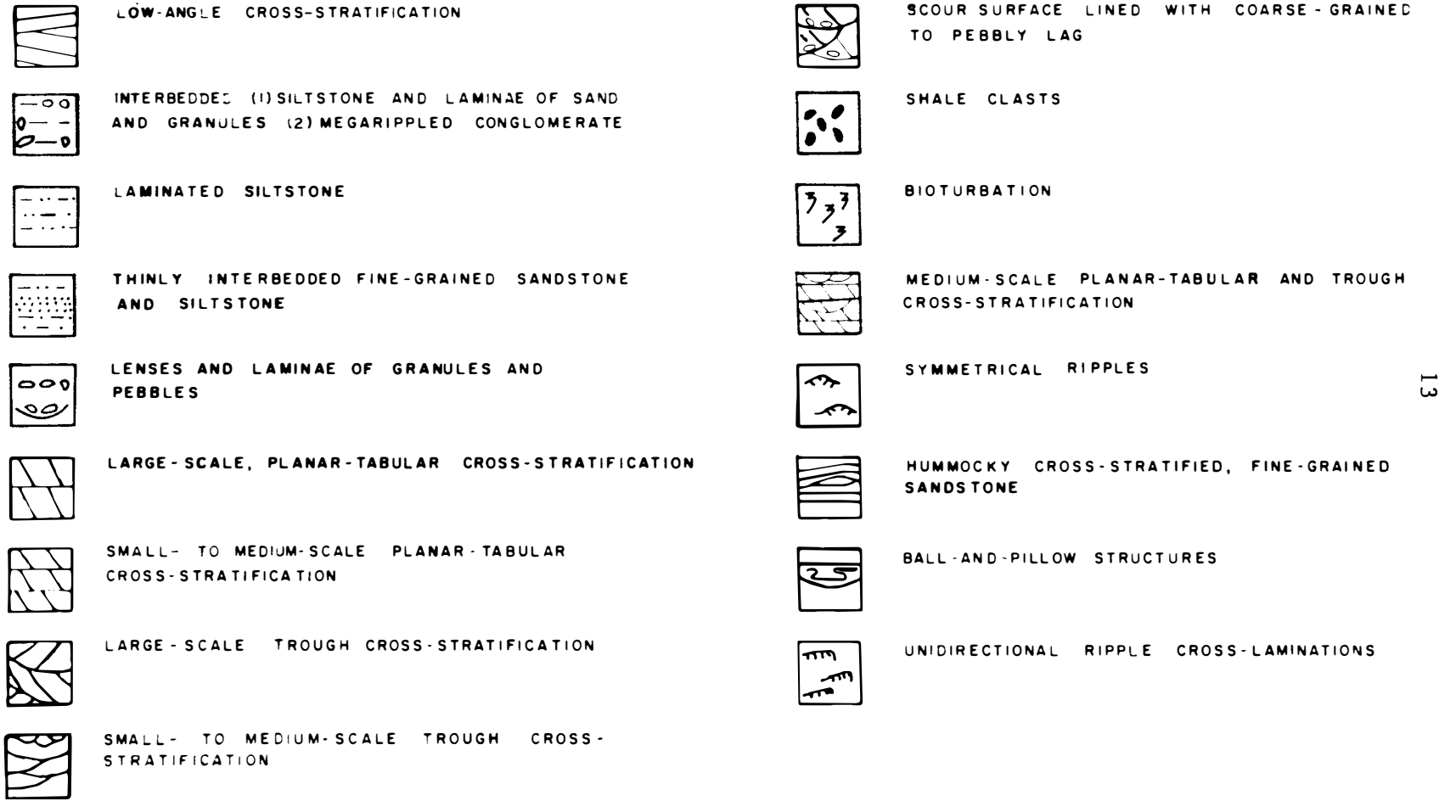


Figure 4. Key to lithologic symbols used in Figs. 3, 4, and 5 (pp. 10, 14, 15).

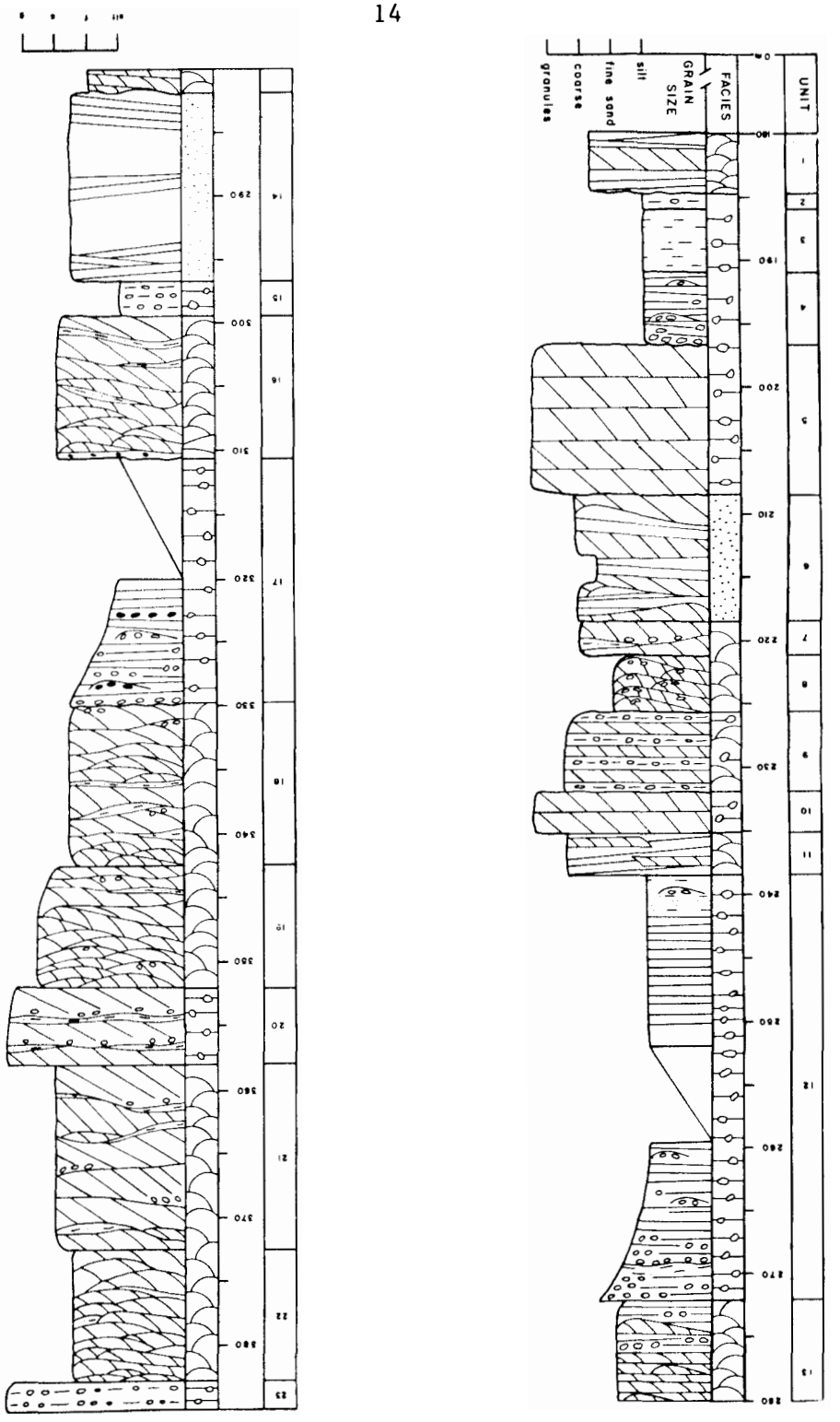


Figure 5. Detailed stratigraphic log of the Unicoi Formation. Unit numbers refer to the units in Appendix A, where lithologic descriptions are recorded. The "FACIES" Column next to the log shows to which facies each unit is assigned. Symbols for each facies are given in Fig. 4.

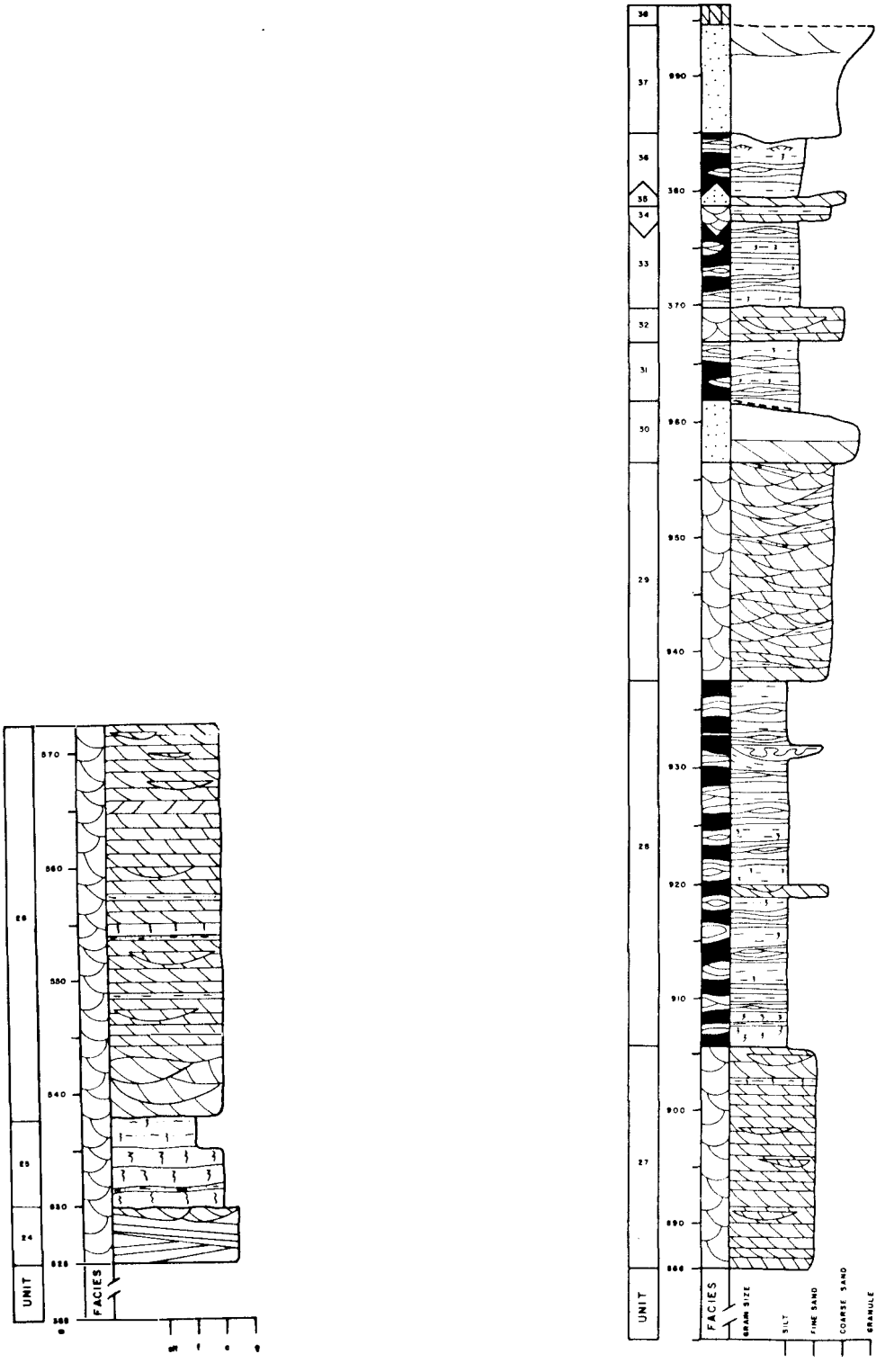


Figure 6. Detailed stratigraphic log of (A) the Hampton Formation and (B) the Erwin Formation.

Facies G: Cross-Stratified, Arkosic Lithic Wacke

General Comments

Facies G occurs only in the Unicoi Formation and generally within the lower part of the sequence where it intertongues with Facies S and Facies QA (Fig. 5). The characteristic common to the three subfacies (Gp, Gh, Gmr) is their immature mineralogy. Their differences lie in grain size and sorting, bedding thickness and stratification sequence.

For ease of discussion, each facies has been assigned an upper-case letter abbreviation. "G" stands for gravel and reflects the abundant granule and pebble-sized grains which are most abundant in this facies. The three subfacies are designated by one or two lower case letters that are abbreviations for the most common sedimentary structure present ("p"-planar-tabular cross-stratification; "h"-horizontal laminations; "mr"-megaripple bedding.

Both hand-sample observation and thin-section examination indicate the compositional and textural immaturity of Facies G. A summary of visual estimates of relative proportions of framework grains from thin-section is recorded in Fig. 7. The samples were classified according to Folk (1968) and Dott (1964). Framework grains consist of: monocrystalline quartz with undulatory extinction, polycrystalline quartz (<5 crystallets), microcrystalline quartz, plagioclase, microcline, foliated polycrystalline quartz, metamorphic rock fragments and sedimentary rock fragments. Compositionally, the plagioclase is An_{10-30}, oligoclase, as determined by the Michel-Levy method.

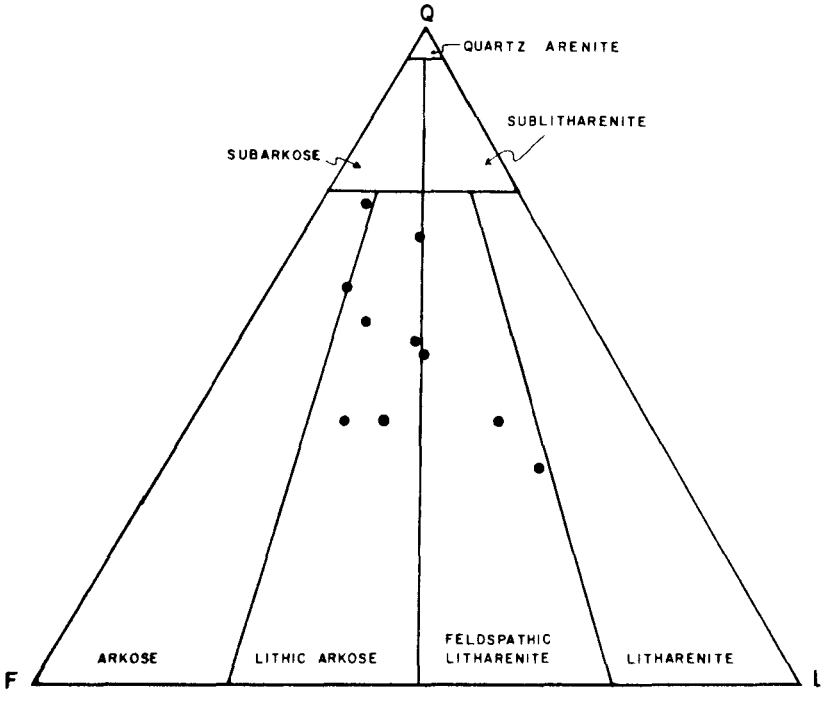


Figure 7. QFL compositional diagram for Facies G based upon visual estimates.

Framework grains range from fine sand to pebbles; granules and pebbles are composed of either polycrystalline quartz, shale clasts or feldspars. Minor constituents comprise 10-15% of the whole rock and include zircon, tourmaline, magnetite, ilmenite, pyroxene, amphibole, biotite and muscovite and are segregated within the finer-grained fraction of each sample.

The matrix consists of optically unidentifiable detrital and authigenic clays (illite-sericite and smectite); pore-filling fibrous and radial-fibrous clays were rarely observed. Mention should be made at this point of the difficulty in identifying depositional matrix. This facies is characterized by abundant lithic fragments, particularly pelitic rock fragments, and these fragments deformed plastically by compaction and/or tectonic deformation. Consequently the fragments now form a pseudomatrix (Dickinson, 1970) due to plastic flowage into pore spaces in response to stress. Fortunately, remnant detrital outlines are commonly visible. Facies G has also been extensively chloritized so that the initial depositional fabric has been obliterated. Feldspars, rock fragments and primary depositional matrix are commonly heavily altered to chlorite and the difference between the three is not always clear. Extensive chloritization is peculiar to Facies G. Fig. 8 shows a typical example of poorly sorted, immature Facies G.

Quartz overgrowths occur, although silica rarely fills entire pores. Pressure solution is probably the most important factor controlling the well-indurated nature of this facies. Interpenetration of grains is common where there is grain-to-grain contact.

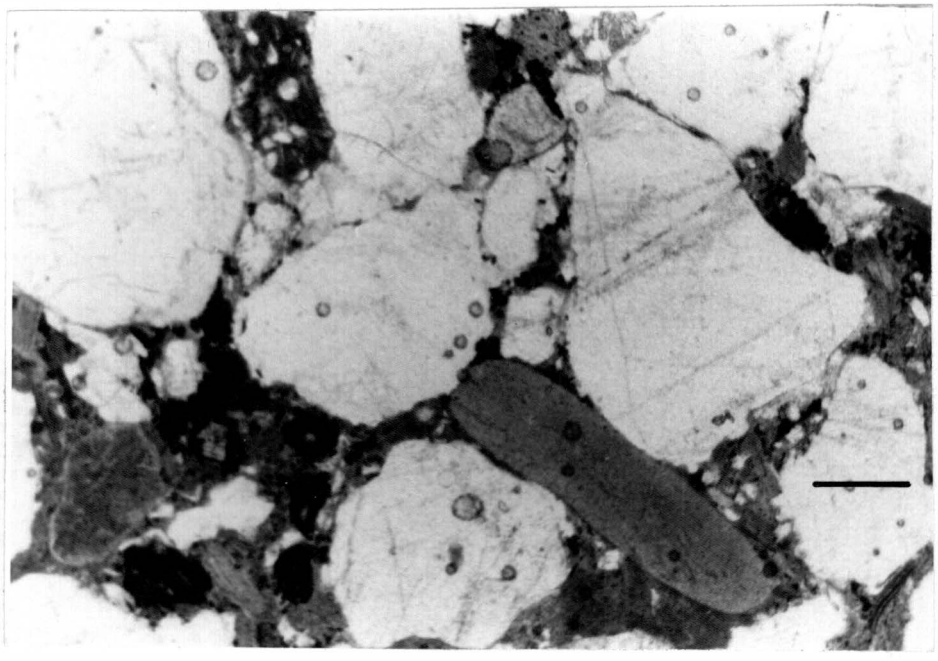


Figure 8. Photomicrograph of a typical Facies C lithology illustrating the poor-sorting and lithic fragments. Scale bar is 1 mm. Sample 11-18-1 U, Unit 2.

Texturally, framework grains are subangular to subrounded and grain size ranges from silt to pebbles. The medium-grained to pebbly fraction tends to show better rounding than do the fine-grained constituents. Facies G tends to be poorly sorted except for Subfacies Gh, which is entirely fine-grained.

Sedimentologically, the three subfacies are dissimilar except for a conspicuous lack of subaerial exposure indicators. The subfacies are:

1. Gp: large-scale, planar-tabular cross-stratified, pebbly lithic arkosic wacke;
2. Gmr: megaripple cross-stratified, pebbly feldspathic lithic wacke;
3. Gh: low-angle to horizontally laminated, fine-grained, feldspathic lithic wacke.

The following is a discussion of the megascopic features of each subfacies.

Subfacies Gp

Gp is characterized by 0.5-3.0 m thick sets of planar-tabular cross-strata (Fig. 9). The base of each set is erosional and horizontal, with no relief on the erosion surface. The 30 cm bottomset contains crude thin laminations of silt, granules and pebbles (Fig. 10). The 4-15 cm thick foresets display tangential bottomsets and are typically graded from pebbles to coarse sand. The foresets are commonly truncated by a megaripple train exhibiting cross-laminations (Subfacies Gmr).



Figure 9. Field photograph of Subfacies Gp, large-scale planar-tabular cross-sets. Unit 5.

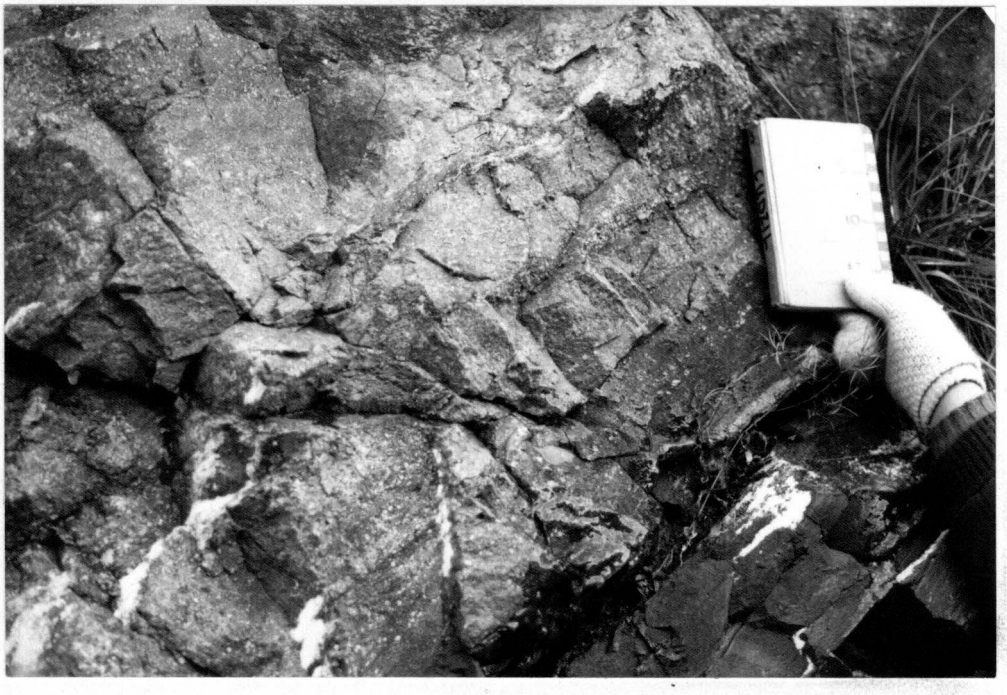


Figure 10. Close-up view of the bottom-set of the cross-set shown in Fig. 9. Unit 5.

Subfacies Gp occurs in units 5, 10 and 20 (Fig. 5, p. 14); it is best developed in Unit 5 in which four stacked cross-sets are separated by thin beds of Subfacies Gmr. The matrix content decreases slightly in each successive set without a change in the framework mineralogy.

Paleocurrent data indicate dominant flow to the northeast and southeast with three azimuths directed to the southwest and northwest (Fig. 11). Only fifteen measurements were made for all of Facies G due to its limited three-dimensional exposure.

Subfacies Gmr

Lithologically, Gmr is identical to Gp but exhibits medium-scale, megaripple cross-stratification in pebbly wacke interbedded with laminated siltstone. The cross-stratified conglomerate beds have flat, erosional bases and wavy tops which apparently represent the original megaripple bedform (Fig. 12). The spacing of the ripples is about 1 m and amplitude is about 5 cm. Internally, foreset and stoss laminae of pebbles, granules and coarse sand are visible. Megaripples are typically overlain with laminated siltstone, which drapes the bedform. Within the siltstone are discontinuous laminae, one grain thick, which consist of very-coarse sand and granules. The interbedded sequences of siltstone and conglomerate range in thickness from 25-100 cm. Because there are no bedding-plane exposures of Gmr, the surficial, three-dimensional geometry of the megaripple is unknown.

The vertical occurrences of Gmr (Units 2, 9, 15, 23) are indicated on Fig. 5. Closely associated with Gp and Gh, Subfacies Gmr truncates

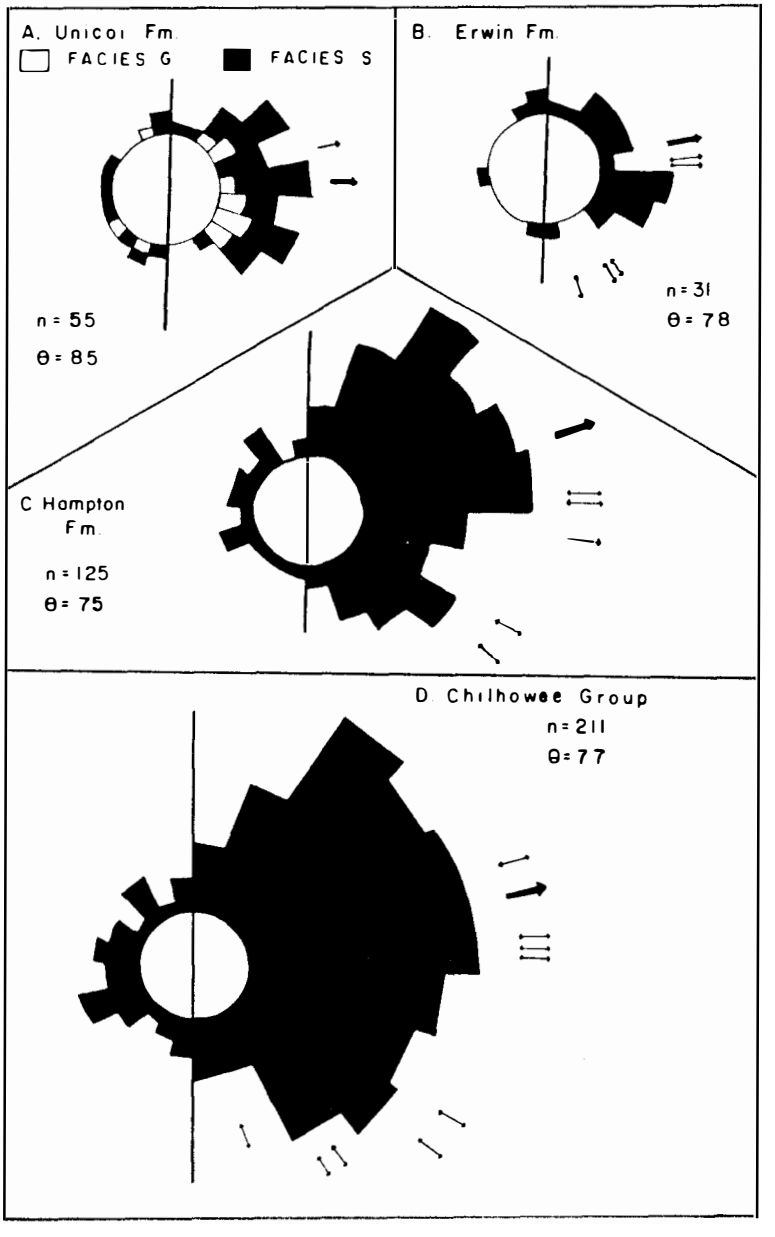


Figure 11. Summary of paleocurrent data. F-statistic for each rose is significant. Double-pointed arrows indicate bidirectional structures and single-pointed arrows indicate unidirectional ripples.



Figure 12. Subfacies Gmr, cross-stratified megaripple with siltstone drape.

the foresets of Gp and caps each sequence of Gh. Subfacies Gmr is commonly interbedded on a scale of 15-30 cm with Subfacies Sms.

Paleocurrents of Gmr are grouped with the rest of Facies G and are summarized in Fig. 11.

Subfacies Gh

Gh consists predominantly of very fine- to fine-grained, feldspathic lithic wacke. However, siltstone lenses, thin siltstone beds, granule/pebble-filled lenses (Fig. 13), granule/pebble laminae and scattered gravel grains are common. The dominant sedimentary structures are horizontal to very low-angle, wavy to even, parallel laminations (Fig. 14). In thin section, each layer is a few grains thick; heavy and opaque minerals as well as micas also define the laminae (Fig. 15). Low-angle erosional surfaces may separate 10-20 cm sets of evenly laminated, fine-grained sandstone. Laminations within the fine sand are indiscernable except for the laminae of gravel.

The typical stratification sequence is a basal gravel lense (lag) that is commonly cross-stratified and is overlain by laminated sandstone. Small-scale trough cross-stratification may cap the sequence. Commonly, there is no ordering of internal stratification features. Parting lineation was not observed; its absence may be a function of rare bedding-plane exposures or a function of the ubiquitous bedding plane slip which would have destroyed such delicate surface markings.

Subfacies Gh occurs three times in the Unicoi Formation in thicknesses of 10 m (Units 3 and 4), 33 m (Unit 12) and 20 m (Unit 17), (Fig. 5). At each stratigraphic level, Gh gradually coarsens upward as

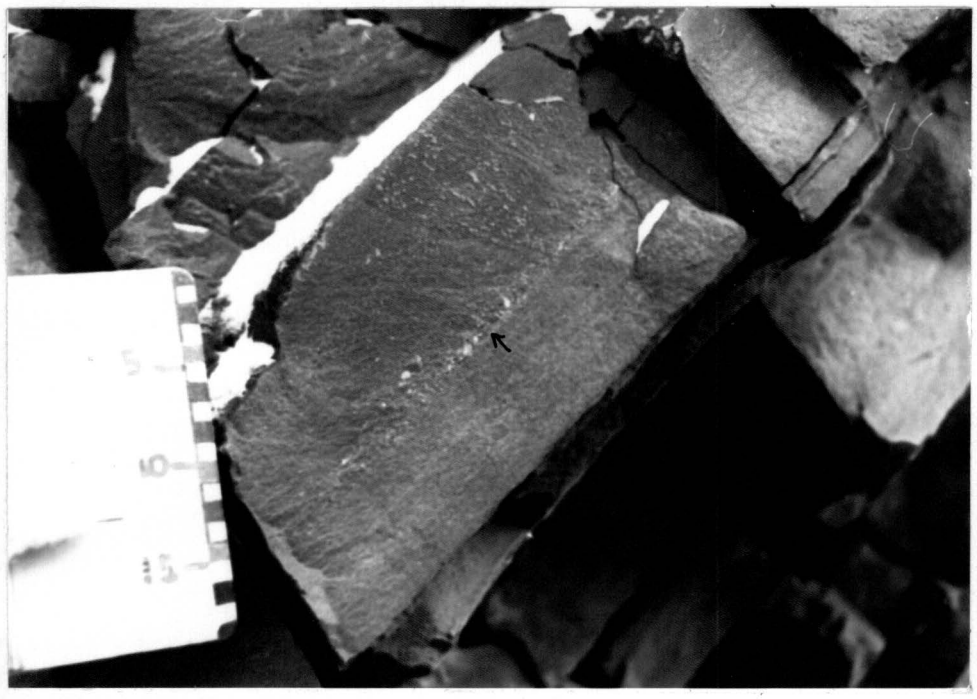


Figure 13. Characteristic horizontal laminations and granule laminae of subfacies Gh. Unit 4.



Figure 14. Unwinnowed granule lag of Subfacies Gh. Unit 4.

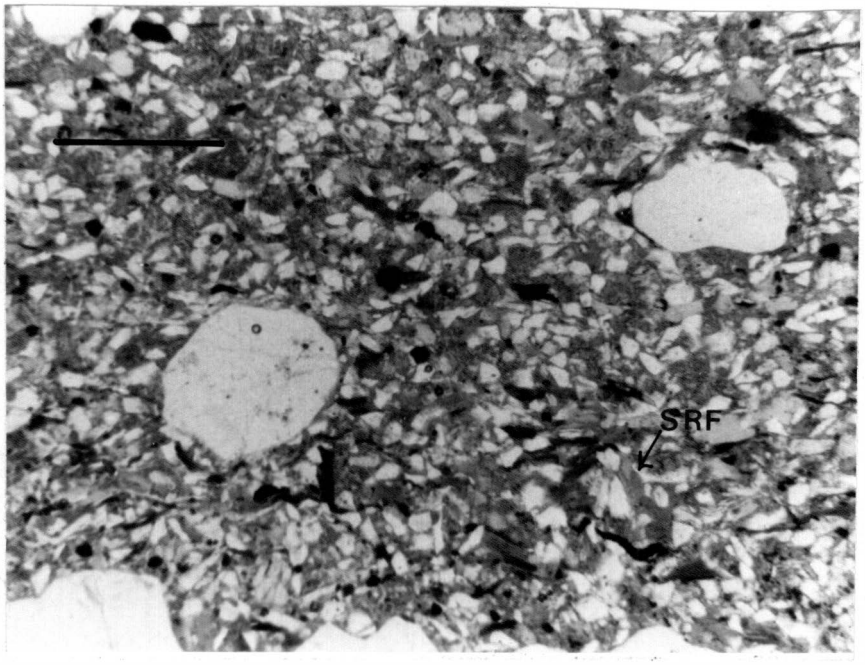


Figure 15. Photomicrograph of Gh showing fine laminae, abundant micas and abundant sedimentary rock fragments. Scale bar is 1 mm. Sample 11-18-1 U, Unit 2.

the frequency of granule/pebble laminae increases and the frequency of siltstone laminae decreases. However, Gh is slightly different at each stratigraphic level. These differences are illustrated in Fig. 16 and each occurrence is discussed in turn.

Units 3 and 4 represent a 10 m coarsening-upward sequence. Within the thinly interbedded siltstone and sandstone at the base of this sequence are two beds containing the trace fossils Paleophycus and Taphrhelminthopsis(?) (Crimes et al., 1977, p. 127). These traces do not occur elsewhere in Subfacies Gh. Paleophycus is a narrow (5 mm diameter), short (5 cm long), lined, smooth-walled burrow. It occurs as hyporelief on the base of a fine-grained laminated wacke bed and is simply branching to nonbranching. The filling of the burrow is identical to the lithology above it (Fig. 17). Pemberton and Frey (1982) discussed the distinguishing features of Planolites-Paleophycus. Distinction between these two horizontal traces in the past has been based upon the presence or absence of branches. Planolites was defined as rarely branched and Paleophycus as being as either branched or unbranched (Alpert, 1975). Pemberton and Frey described Planolites as being an unlined burrow that is filled with sediment having a texture and fabric unlike the host rock. The behavior indicated by this trace is active backfilling of sediment (i.e., fecal material) in a burrow constructed by a mobile deposit-feeder.

Paleophycus, on the other hand, is a lined burrow filled with sediment that is identical to that of the surrounding matrix. Paleophycus forms from passive sedimentation within an open dwelling burrow

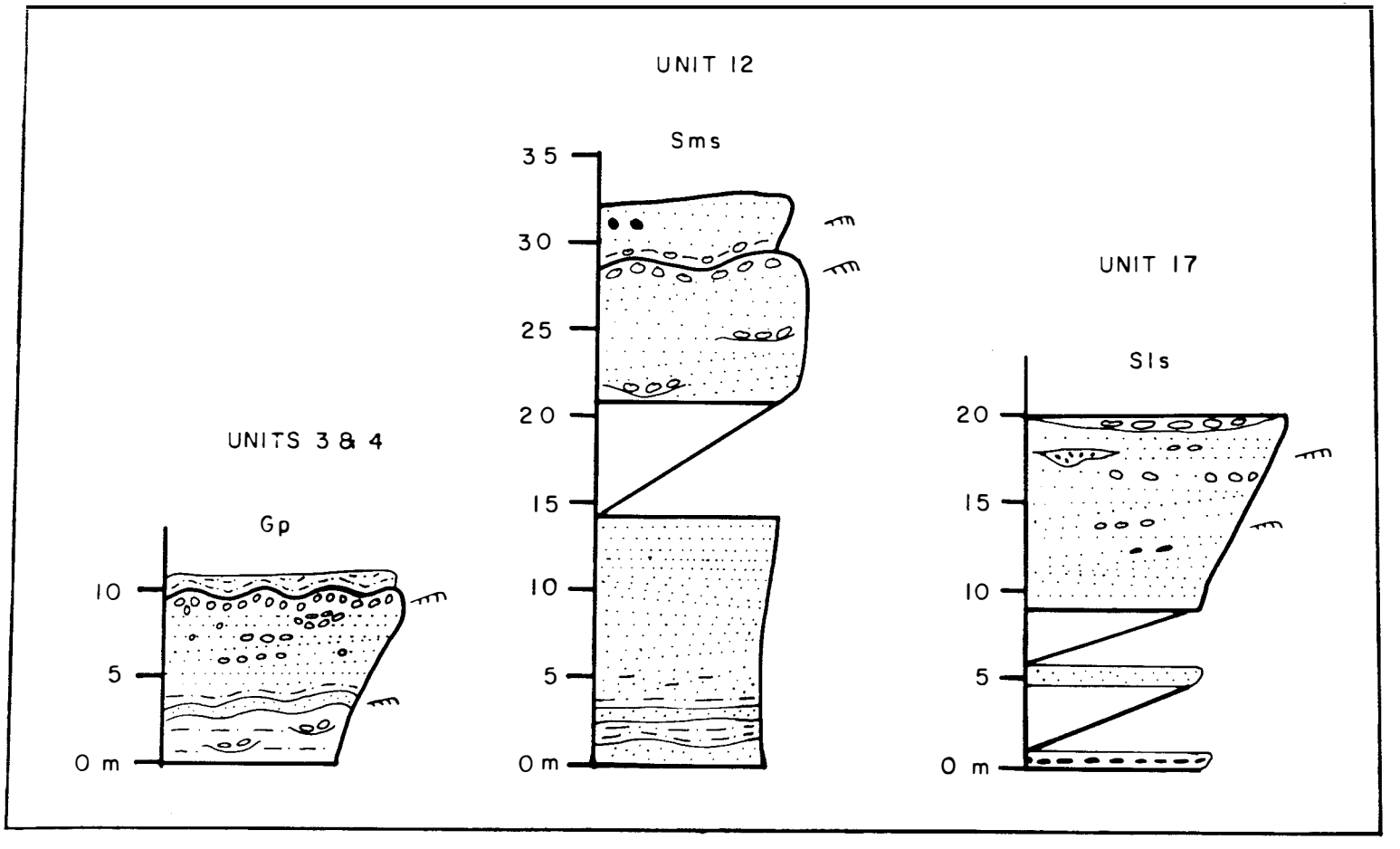
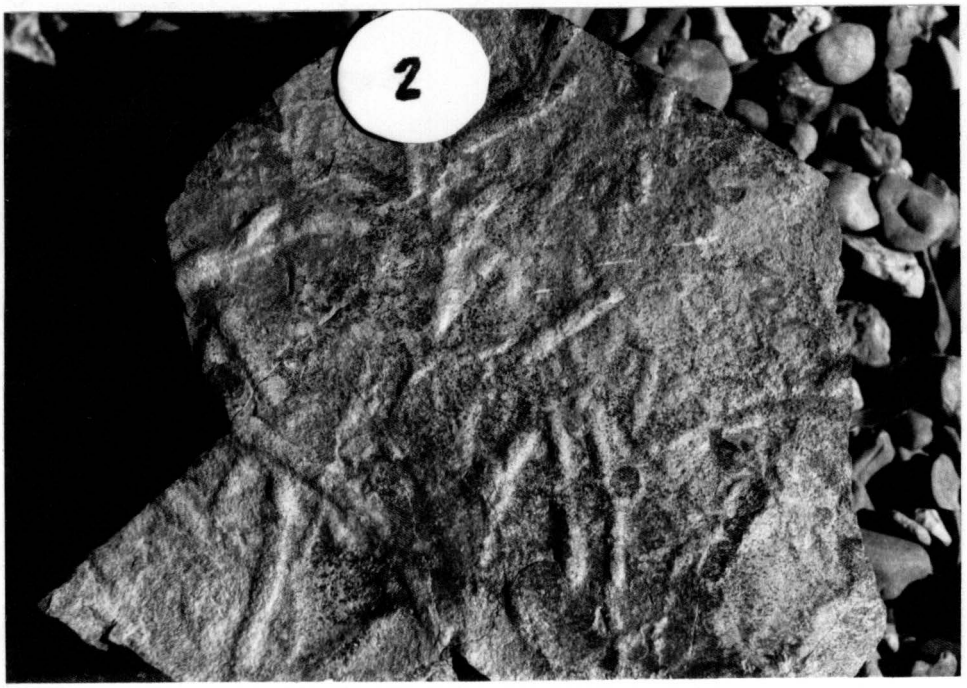
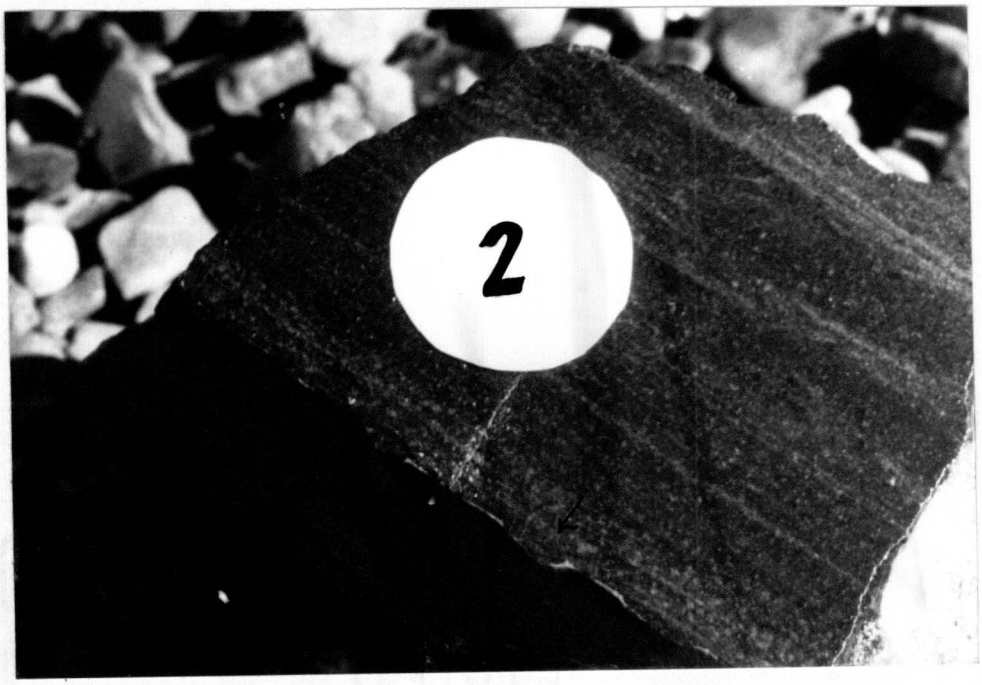


Figure 16. Vertical variability of Subfacies Gh and a comparison between its three stratigraphic occurrences.



A.



B.

Figure 17. Bedding plane view of Paleophycus (A) and cross-sectional view (B) that is found only once in Subfacies Ch. In B, the burrow is lined and its fill is identical to sediment above the burrow. Circle is 2 cm. Unit 3.

constructed by a predaceous or suspension-feeding animal such as a polychaete worm. The trace tenatively identified as Taphrhelminthopsis(?) consists of two smooth and parallel furrows separated by a medial ridge in negative and positive relief on the base of a laminated, fine-grained sandstone bed. This trace is about 2 cm wide, 20 cm long and exhibits an irregular, nonbranching and circling habit. Generally, most documented examples of this ichnogenus occur as positive bilobate features on soles of beds, although Ksiazkiewicz (1970, p. 292) mentioned forms occurring as positive features on the upper surface of beds. Bilobate traces of these types are normally inferred to have been produced by gastropods (Häntzschel, 1974, p. W113). The presence of these two traces can be taken as evidence for a marine influence on the interbedded sandstone and siltstone of the lower 4.5 m of Gh at this stratigraphic level (Units 3 and 4).

Small 10 cm sand volcanoes occur in a single bed of Unit 3; they are a result of dewatering of fine-grained sediment due to rapid deposition. This penecontemporaneous deformation is restricted to one bed and the sediment above and below it was not deformed; these structures are not a result of tectonic deformation. Dewatering structures were not observed in the other two units of Gh.

The second exposure of Gh (Unit 12) contains the greatest thickness of interbedded siltstone and sandstone (5.5 m), which only occur at the base of the sequence and the subfacies again gradually coarsens upward. Gh is unique at this level due to its 8 m thickness of horizontally laminated arkosic lithic wacke (Fig. 16). This rock type is what King

and Ferguson (1960) termed a "sedimentary greenstone," because it weathers with conchoidal fracture and has the superficial appearance of a metabasalt (Fig. 18). It is only faintly laminated and there are no granule/pebble or silt lenses. The upper 25 m exhibit a coarsening-upward trend. Unit 12 is overlain by trough cross-stratified subarkosic arenite of Subfacies Sms.

There is a break in the coarsening-upward trend of Unit 12, 29 m above the base of the unit, where a thin bed of Subfacies Gmr is abruptly overlain by a 1 m bed of laminated black siltstone with sand and gravel stringers. Above the siltstone, the coarsening trend continues and for the first time, shaley siltstone clasts are present.

The third occurrence of Gh (Unit 17) also gradually coarsens upward, and lenses and beds of silty-shale pebbles (rip-up clasts), about 2 cm in length, are common. Any original orientation of the clasts is overprinted by compaction and tectonic deformation. This unit gradually becomes cleaner (more arenitic) up-unit, whereas the matrix content remains constant in the other Gh units.

Paleocurrent measurements on low-angle cross-stratification and on the cross-stratification within conglomerate lenses indicate flow predominantly to the east-southeast. Fig. 11A (p. 23) summarizes the paleocurrent data for Facies G (n = 15) and shows paleocurrents directed to the east. Azimuths of three pebble/granule laminae of Gh, dipping at low-angles, indicate paleoflow to the northwest and southwest.



Figure 18. "Sedimentary greenstone" of King and Ferguson (1960) with conchoidal fracture. This is Subfacies Ch; fine, horizontal laminae can be seen in the center bottom of the photograph. Unit 12.

Facies S: Cross-Stratified Arkosic Arenite

General Comments

Hand-sample and thin-section study shows that Facies S consists of subarkosic to arkosic arenite (Fig. 19) with minor amounts of micaceous siltstone. Framework grains are, in order of decreasing abundance: monocrystalline quartz (70-98%), plagioclase (An_{10-30}), orthoclase, microcline, foliated polycrystalline quartz, and less than 1% zircon, tourmaline, magnetite, and ilmenite. Grain size ranges from fine sand to granules. Grains are rounded to well-rounded and moderately sorted. There is no depositional matrix present, although secondary chlorite-sericite-illite on rare occasions rims grains and fills pores. Glauconite is a locally abundant constituent of Facies S in the Erwin Formation. However, no glauconite was observed in the Unicoi or the Hampton Formations. Quartz overgrowths and interpenetration of grains resulting from pressure solution make the rocks of this facies well-indurated.

A total of 221 paleocurrent measurements was taken from the whole Chilhowee Group; 156 of these measurements were from Facies S (Fig. 11). Paleocurrents are directed to the northeast, east and southeast with only a minor mode to the west. Only rarely are the azimuths for superjacent sets oriented at 180° to each other. There is no relationship between cross-set thickness and respective paleocurrent direction (i.e., all the small-scale sets are not characterized by a single preferred direction). Oscillatory currents were directed east-west or southeast-northwest.

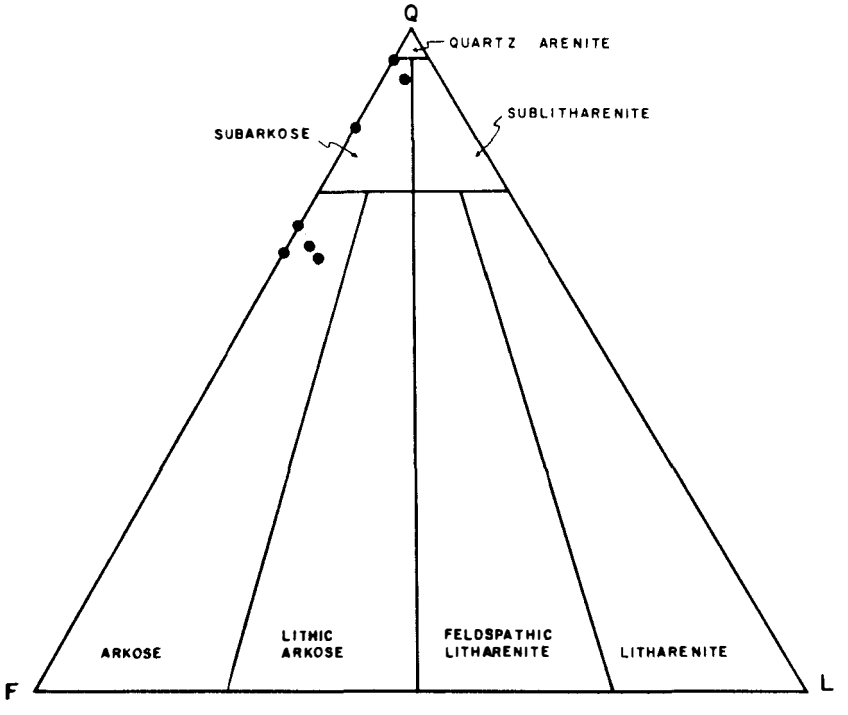


Figure 19. QFL diagram for samples of Facies S, based upon visual estimates.

Facies S occurs in beds from 10 cm up to 1.5 m thick. Three subfacies are distinguishable based upon the stratification sequences present. They are:

1. Sfl: laminated siltstone and fine-grained sandstone with wavy to lenticular bedding;

2. Sls: fine- to very coarse-grained subarkose with large-scale planar-tabular and trough cross-stratification;

3. Sms: fine- to coarse-grained subarkose with small- to medium-scale planar-tabular and trough cross-stratification.

Subfacies Sfl

Although this subfacies is lithologically dissimilar to Sms and Sls, its common occurrence with the coarse-grained subfacies warrants its placement within this Facies S. Subfacies Sfl consists of interlaminated to thinly interbedded, micaceous siltstone and fine- to coarse-grained sandstone; Sfl commonly occurs in thicknesses ranging from 1 cm to 1 m (Fig. 20). The laminae characteristically drape the relief on underlying cross-stratified sandstone of Sms or Sls.

The sandstone commonly occurs as single-grain thick stringers within the siltstone, however, when the sandstone is abundant enough to form thin beds, it is typically characterized by wavy to lenticular bedding and unidirectional ripple cross-lamination.

Trace fossils of this subfacies are absent in the Unicoi, and bioturbation gradually increases through the Hampton Formation and into the Erwin Formation. In these two formations, views perpendicular to



Figure 20. Interlaminated, bioturbated siltstone and sandstone of Subfacies Sf1. Note the cross-section of horizontal burrows. Unit 25.

bedding show that sandstone layers and lenses have been biogenically disrupted (Fig. 20); only rarely do primary ripple cross-laminations remain. Elliptical to circular horizontal burrow cross-sections are commonly visible (Fig. 20). On a bedding surface, these burrows are smooth-walled, simply branching and are about 1-10 cm long. There is no evidence for complex deposit-feeding behavior; these traces probably belong to the ichnogenus Paleophycus (Frey, 1975).

In the Erwin Formation, Rusophycus and Cruziana dominate the trace fossil assemblage of Sf1, although horizontal burrows are present. This subfacies occurs as 1-30 cm caps on top of cross-stratified sandstone and these traces are best seen as hyporelief on the base of the sandstone beds. Rusophycus and Cruziana vertically increase in size from 1-15 cm in width. The overall trends of increasing trace fossil diversity and abundance throughout the Chilhowee Group will be discussed later.

Subfacies Sls

This facies consists of either (1) single sets or cosets of large-scale planar-tabular and trough sets that are about 1-1.5 m thick or (2) thick beds of inversely graded conglomeratic sandstone. Individual set boundaries may be marked by discontinuous, 1-cm-thick siltstone partings. Where this is not the case, the coset weathers as a massive ledge. The flat even base of the sequence is marked by a granule or very coarse-grained quartz/feldspar lag which is typically about 24 cm in length. In some instances, a lag may be present at the top of a

coset. Foresets range in thickness from 1-30 cm and are commonly graded. No small-scale cross-stratification is observable within the foresets. In fact, the internal lamination of an individual foreset parallels the foreset plane. The coset or large-scale set may thin to one-half of its greatest thickness and the relief seen in outcrop on the upper surface reflects either the original bedform or scour of the original bedform (Fig. 21). Subfacies Sls is typically overlain by medium-scale planar-tabular or trough cross-stratified Subfacies Sms or by interlaminated siltstone and sandstone. In one particularly well-exposed example, a large-scale planar-tabular set is covered with megaripples (trough cross-stratification) and each megaripple appears to have some small-scale trough cross-stratification at the top (Fig. 22). In another example, the top of a large-scale set is scoured as evidenced by the truncated foresets. The relief on the scour is 40-50 cm, and thinly interbedded sandstone and siltstone drape and fill the scour trough. In some cases, no scour surface defines the top of Sls and set thicknesses decrease upwards to the predominantly medium-scale cross-stratification of Subfacies Sms (Fig. 23).

Subfacies Sms

Medium-scale, planar-tabular and trough cross-stratified, fine- to medium-grained subarkosic arenite constitutes the greatest thickness of rocks within the Chilhowee Group (177 m out of 362 m). It occurs throughout the stratigraphic section and is associated with the other three facies: G, QA and Shcs. Unlike Subfacies Sls, the individual sets

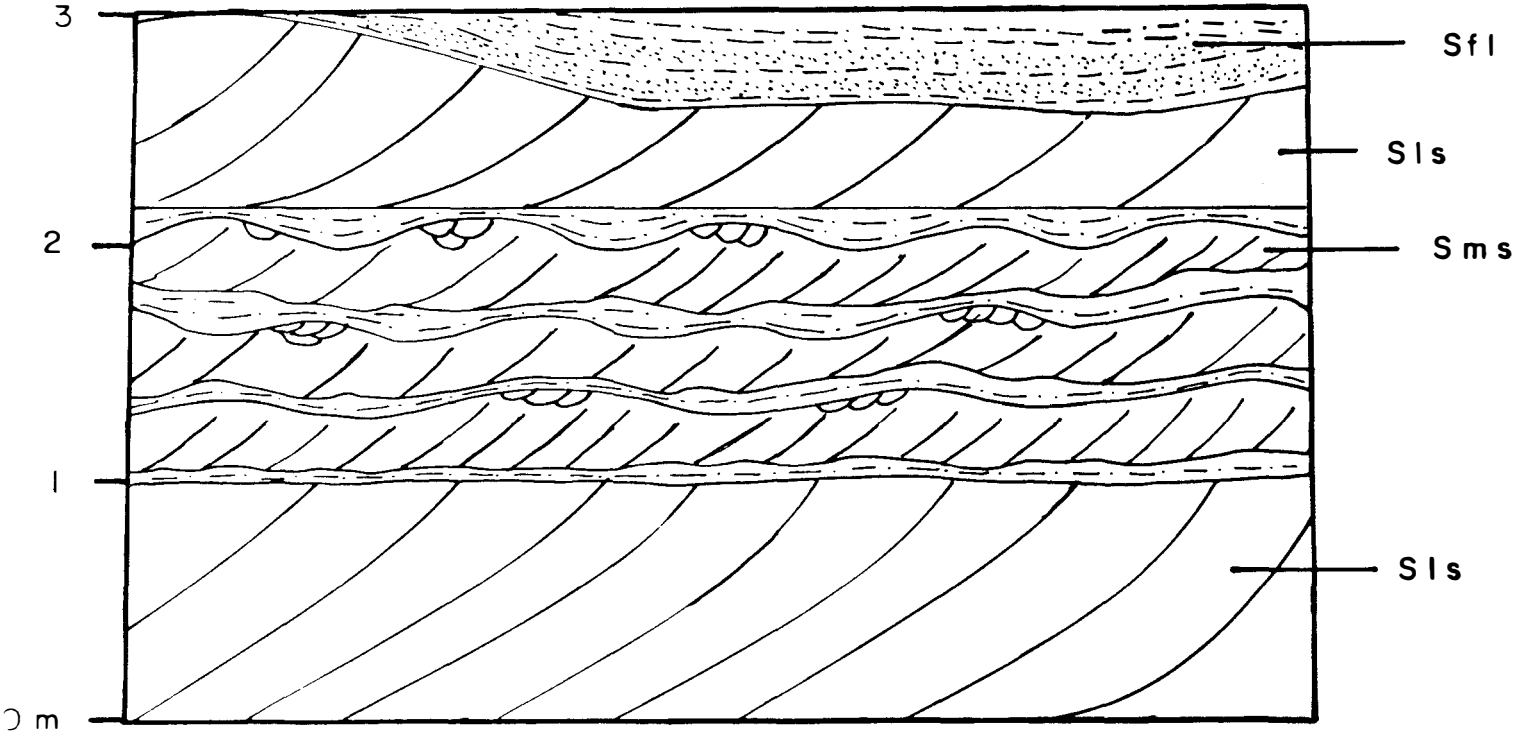
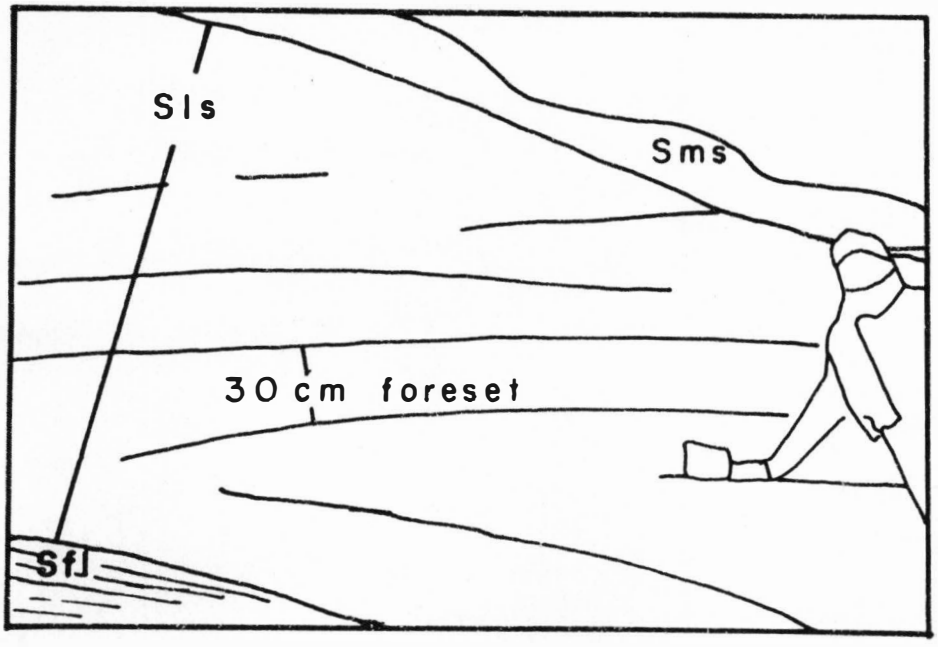


Figure 21. Sketch from field notes of the stratification sequence of Facies S.



A.



B.

Figure 22. Photograph of Subfacies Sls (A) with explanatory sketch (B). Internally, the 30 cm foresets are horizontally laminated, not cross-stratified. Unit 26.

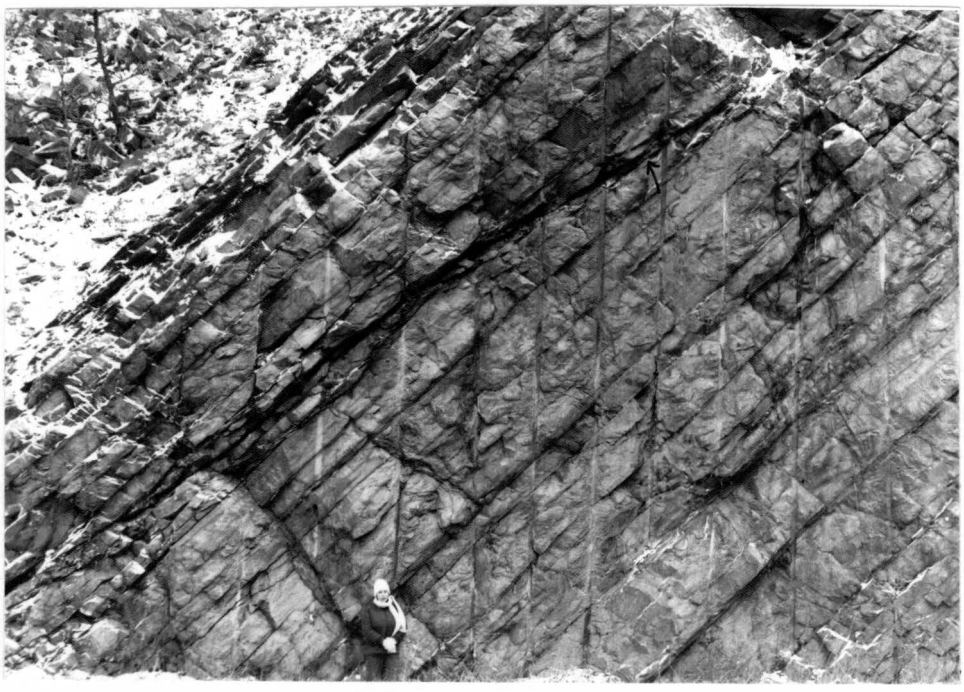
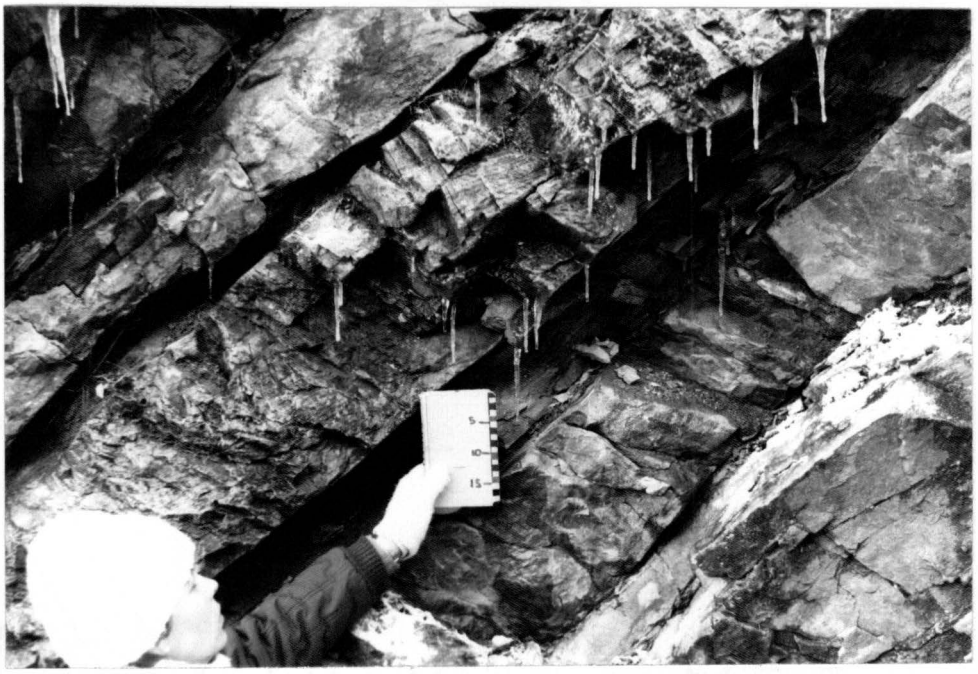


Figure 23. The thinning-upward trend that characterizes Facies S. The surface at a low-angle to bedding is a fault. Unit 16.

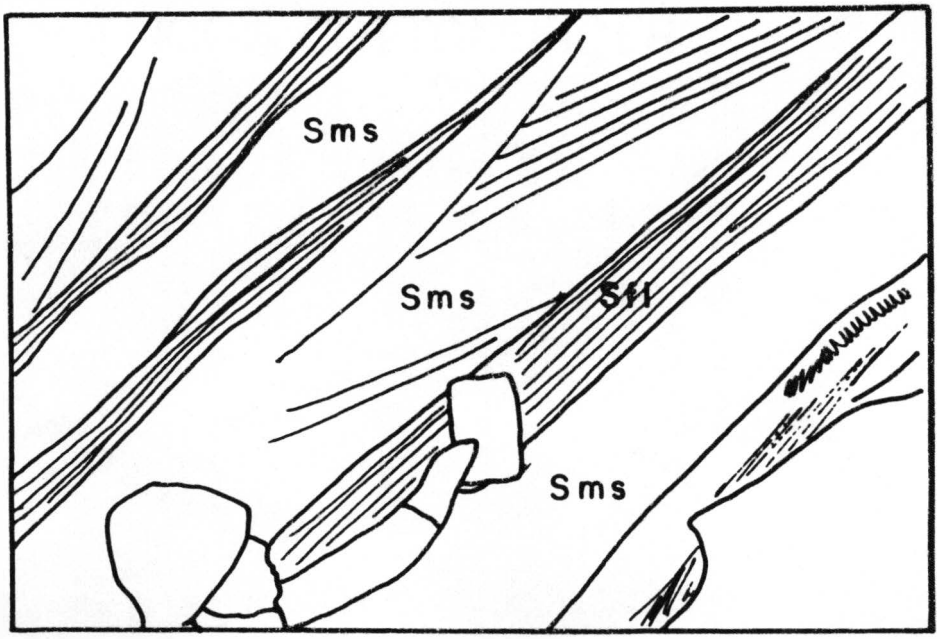
are bound above by a siltstone and/or fine-grained sandstone layer (Subfacies Sf1). These medium-scale sandstone beds occur typically in stacked sets, one to tens of meters thick, in which the individual sets are laterally continuous (30 m) in thickness. In the Unicoi and Erwin Formations, the cross-stratification is predominantly trough and the upper and lower contacts are undulatory (Fig. 24). In the Hampton, planar-tabular cross-stratification is dominant and the capping siltstone, if present at all, is less than 2 cm thick (Fig. 25). The basic repeated sequence is a flat to wavy scoured base → medium-scale cross-stratification → small-scale cross-stratification → laminated siltstone/sandstone cap. No reactivation surfaces were observed.

Whereas small-scale cross-laminations may overlie a medium-scale set, ripple marks per se are not particularly abundant in Sms or Sls. However, when present, the ripples are of the straight-crested, simple to bifurcating wave variety with slightly rounded crests. Their heights are 1 cm and they have a spacing of 10-15 cm. Other crests have been modified to an interference ripple pattern. Because the ripple marks are typically exposed 10-15 m up on the vertical outcrop, the author was unable to observe the internal laminations of the ripples. It is highly possible that these are wave-modified current ripples rather than simple oscillatory ripples.

The trace fossils of Sms in the Hampton Formation are paired or single vertical burrows approximately 5 mm in diameter and are visible only on bedding surfaces (Fig. 26). They occur in thin isolated beds (<4 cm thick) of fine- to medium-grained sandstone. These thin beds



A.

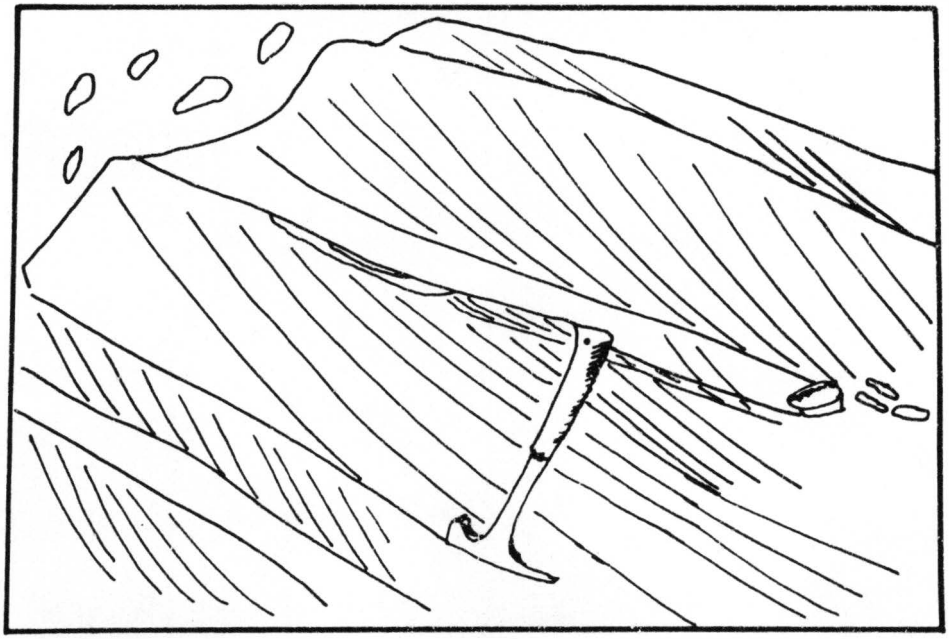


B.

Figure 24. Photograph (A) and explanatory sketch (B) of Subfacies Sms of the Unicoi Formation. Trough cross-stratification with undulatory upper and lower surfaces (Sms) is overlain by laminated siltstone and sandstone (Sfl). Unit 22.



A.



B.

Figure 25. Photograph (A) and explanatory sketch (B) of Hampton Formation illustrating abundant medium-scale, planar-tabular cross-stratification and the lack of fine-grained interbeds. Unit 26.

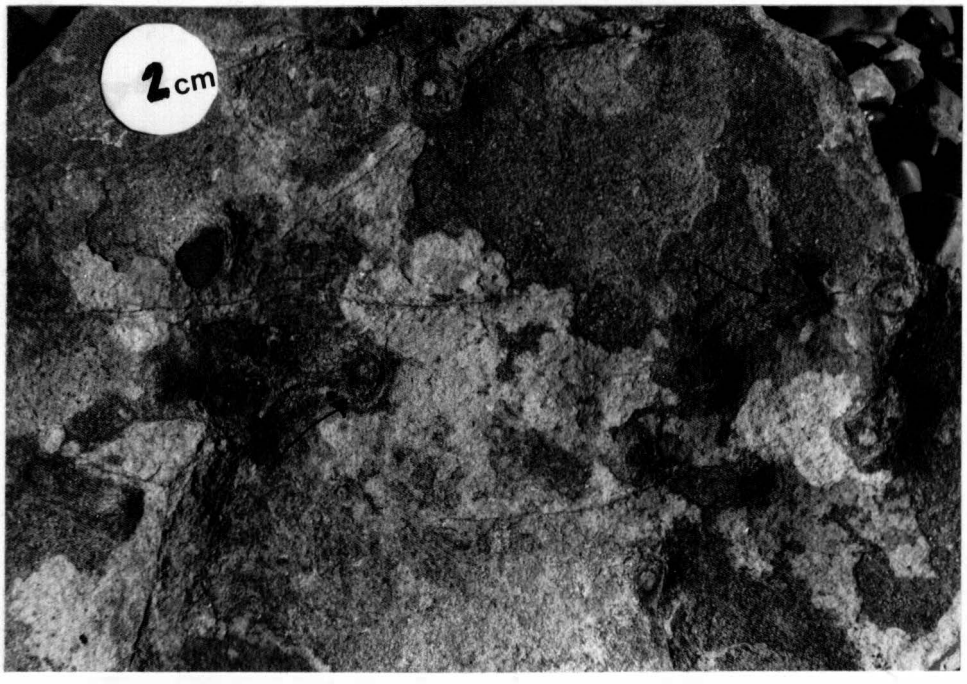


Figure 26. Bedding plane view of paired burrows of Subfacies Sms. Unit 26.

overlie cross-stratified sets and the burrows do not extend into the underlying bed. One planar-tabular cross-stratified bed appeared to have thin (5 mm diameter) Skolithos burrows. The burrows were observed only once and are conspicuously absent elsewhere.

There are a few important sedimentological changes in this subfacies that occur at different stratigraphic levels. First, several beds of Sms in the Erwin contain abundant (40%) well-rounded glauconite grains; glauconite was not observed anywhere in the Hampton or Unicoi Formations. Second, the amount of bioturbation increases upwards; it is absent in the Unicoi and rarely present in the Hampton. Bioturbation is abundant in the Erwin, but is concentrated in the fine-grained layers (Subfacies Sf1) between the cross-stratified sandstone of Subfacies Sms.

Facies Shcs: Hummocky Cross-Stratified, Fine-Grained, Arkosic Arenite

Facies Shcs occurs only within the Erwin Formation (Fig. 6, p. 15) and consists of laminated fine-grained sandstone and bioturbated siltstone. The fine-grained sandstone is composed of equal amounts of feldspar (plagioclase, microcline, and orthoclase) and monocrystalline quartz, about 35-40% each. Other constituents are detrital micas, rare sedimentary rock fragments, tourmaline, zircon, magnetite and ilmenite. Glauconite is locally abundant and no depositional matrix is present. Grains are well-sorted, silica cemented and show evidence of pressure solution. Unlike the Facies G and S, Shcs cannot be subdivided into subfacies.

The thicknesses of individual siltstone and sandstone beds range from 1-20 cm, although sandstone beds are up to 70 cm thick. The typical stratification sequence (Fig. 27) begins with a flat, sharp-based planar-laminated arkosic arenite. Tool or drag marks, parting lineations, or wrinkle marks are commonly visible on the bottoms of sandstone beds. Overlying the planar-laminated zone is a zone of hummocky cross-stratification.

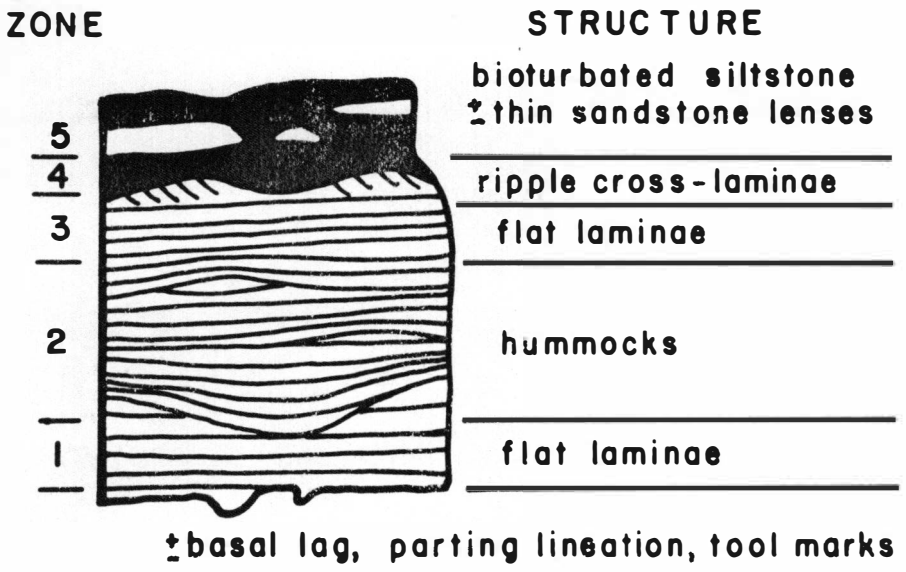
Hummocky cross-stratification has been defined by Harms et al. (1982), to consist of gently dipping laminae with low-angle truncations (Figs. 28 and 29). Salient features are: (1) erosional lower bounding surfaces that dip at angles less than 10°; (2) laminae above the erosional set boundary are parallel to that surface; (3) laminae may thicken laterally and their dip may diminish upwards; (4) dip directions of laminae are random.

Overlying the hummocky zone are ripple cross-laminations and/or wave ripple train. Finally, the cross-stratified sandstone is capped with a bioturbated siltstone.

To find such a sequence completely preserved is exceptional, and a continuum of variations of the ideal sequence is common (Fig. 30). At one end of the continuum, equal amounts of siltstone and sandstone are very thinly to thinly interbedded (Fig. 30A). Sandstone beds generally have wavy bases and tops, and may be unstratified due to intense bioturbation; otherwise, the beds exhibit ripple cross-laminations. Within thicker (10-30 cm) sandstone units, most of the "ideal" sequence is preserved. However, at the other end of the continuum are 70 cm thick



A.



B.

Figure 27. Photograph (A) and explanatory sketch (B) of an ideal hummocky sequence of Facies Shcs. Unit 28.

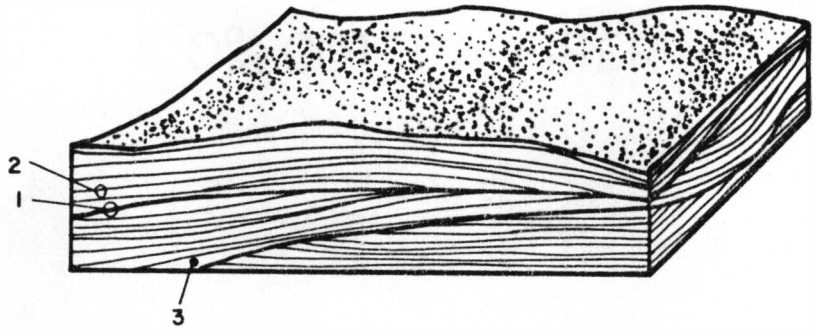
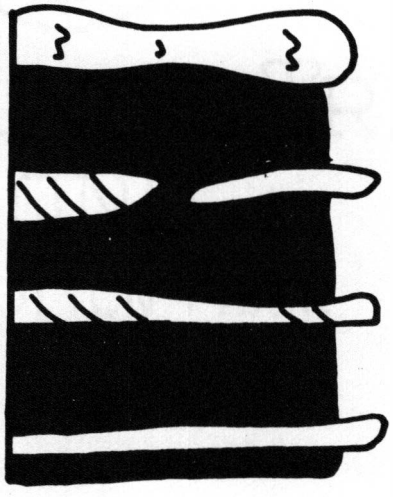


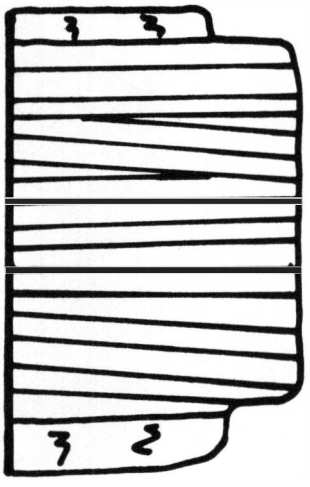
Figure 28. Schematic sketch of hummocky cross-stratification (from Harms et al., 1982) illustrating characteristic features. (1) erosional surfaces truncating the underlying laminae at low angles; (2) laminae above parallel the erosional surface; (3) laminae thicken in trough and thin over crest.



Figure 29. Field photograph of the features that define hummocky cross-stratification. Numbers are keyed to the text and to Fig. 28. Unit 28.



A.



B.



C.

Figure 30. Variations of the ideal hummocky sequence and their vertical occurrence. (A) Distal storm sandstone and fair-weather siltstone, zones 4 and 5; (B) amalgamated hummocky cross-stratification, zones 1 and 2; (C) coarsening-upward sequence of Facies Shcs. Unit 28.

beds of horizontally and evenly laminated sandstone; siltstone and well-developed hummocky zones are conspicuously absent (Fig. 30B).

This spectrum of preservational possibilities occurs within vertical 1-4 m coarsening-upward sequences (Fig. 30C). At the base, the siltstone and sandstone are about equally interbedded. As the siltstone content decreases, the thickness of the sandstone beds increase and the ideal sequence may be preserved. Sharp-based and flat-topped beds of sandstone exhibiting only planar laminations are found with increasing frequency until finally a 50-70 cm thick planar-laminated sandstone caps the whole sequence. The coarsening-upward sequence commences on top of the laminated sandstone with interbedded siltstone and sandstone.

Shcs is also characterized by two types of synsedimentary deformational features: convolute laminae and ball-and-pillow structures. The convolute laminae occur within the laminated sandstone beds of Shcs. They are typically restricted to a single 20 cm bed; beds above and below the deformed layer retain their initial depositional fabric.

The ball-and-pillow structures consist of laminated or unlaminated fine-grained sandstone and are encased by interbedded siltstone and hummocky cross-stratified sandstone. These structures range in diameter from 30-150 cm (Fig. 31--note that the upper half of the pillow is missing, and the erosional surface indicates scouring prior to deposition of the overlying siltstone and sandstone). The silt and sand directly below the ball-and-pillow structures lack bedding due to the soft sediment deformation. One layer of semispherical pillows is strikingly uniform; five pillows are equally spaced and the thickness of each pillow is about 1.5 m.

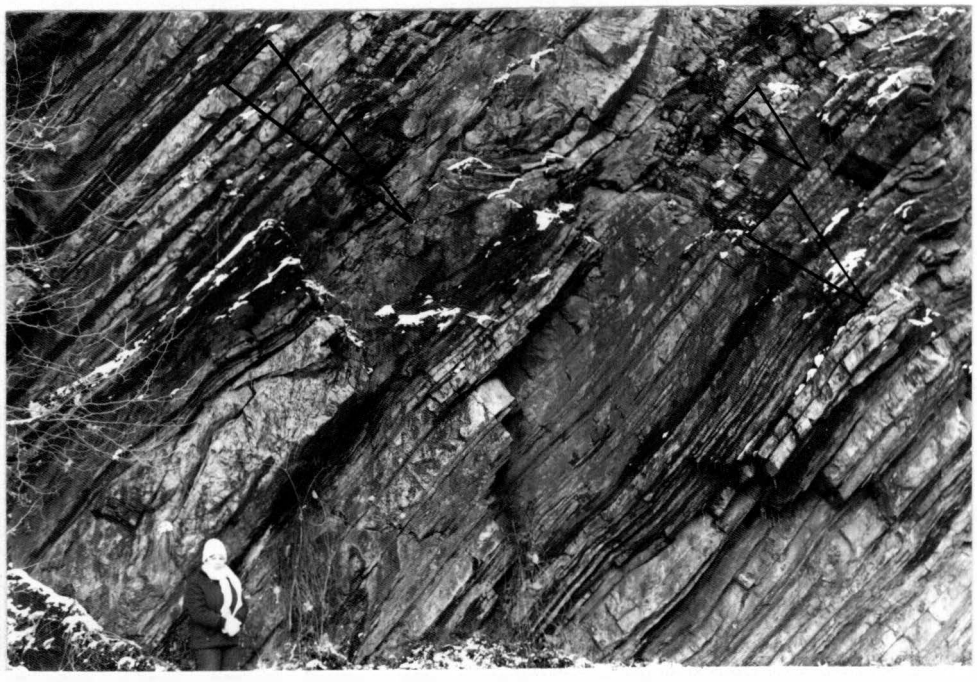


Figure 31. Ball-and-pillow structures of Facies Shcs. Note the repetition of the small-scale (1-4 m) coarsening-upward sequences. Unit 28.

The trace fossil assemblage within Shcs consists of Rusophycus, Cruziana, and horizontal to subhorizontal burrows. The only real difference between Sfl and the siltstone of Shcs is the type of co-occurring stratified sandstone.

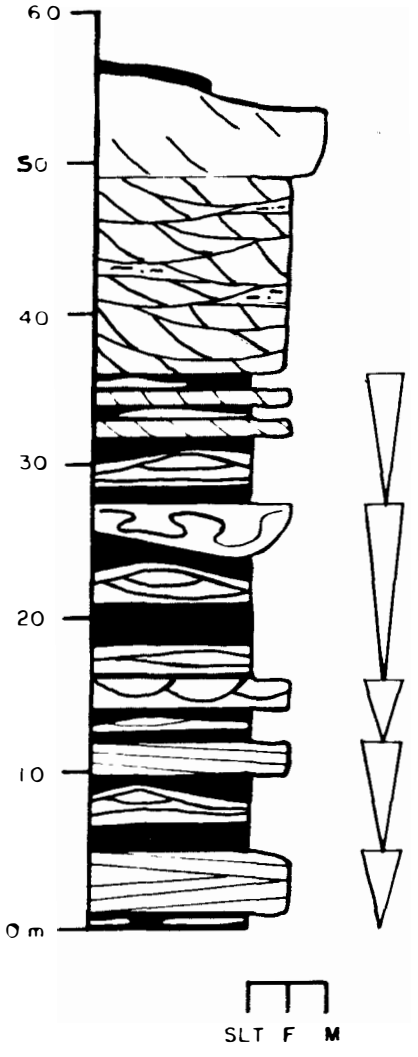
Facies Shcs is interbedded with Facies S and Subfacies QA-Erwin (Fig. 6, p. 15) and has a total stratigraphic thickness of 49.69 m (Units 28, 31, 33 and 36). Unit 28 represents the lowermost occurrence and it abruptly rests upon the medium-scale, cross-stratified sandstone of Unit 27 (Sms). Unit 28 is not a simple package of equally interbedded siltstone and sandstone. As the sequence illustrated in Fig. 30C is repeated vertically, the planar-laminated arenite beds become thicker and the interbedded siltstone and sandstone are thin to absent. Unit 28 coarsens upward as the siltstone fraction decreases.

The cross-stratified arenite beds of Unit 29 also represent a continuing coarsening-upward sequence as the thin siltstone beds become siltstone partings, and Unit 30 (Subfacies QA-Erwin) caps the sequence (Fig. 32) with a medium- to coarse-grained supermature quartz arenite. Units 31-37 record a similar coarsening-upward sequence. The only difference is that Sms occurs sporadically within Facies Shcs rather than forming a thick distinctive unit such as Unit 29.

Facies QA: Cross-Stratified, Supermature Quartz Arenite

General Comments

Facies QA consists of 95-98% well-rounded, well-sorted monocrystalline quartz; minor constituents include zircon, tourmaline, microcrystalline quartz, polycrystalline quartz and less than 5% feldspar. Grain



DESCRIPTIONS	INTERPRETATIONS
UNIT 30 SUBFACIES QA-ERWIN Supermature quartz arenite, large-scale, planar-tabular cross-sets, body capped by a 30 cm bed of shale that contains lenses of granules and glauconite	Large subtidal sand ridges, related to the detachment of shoreface channel-shoal couplets during transgression
UNIT 29 SUBFACIES Sms Medium-scale planar-tabular and trough cross-stratification, discontinuous siltstone interbeds decrease in thickness upwards; glauconite locally abundant.	Cross-stratification is a result of migration of megaripples during tidally-enhanced storm current, siltstone represents periods of normal sedimentation periods, silt probably draped whole bedform but was partially eroded with the ensuing high energy event, megaripples segregated in patches
UNIT 28 FACIES Shcs Sandstones range in thickness 1-70 cm; exhibit horizontal laminations, hummocky cross-stratification, undirection ripple cross-stratification, upper contact is transitional Bioturbated siltstones 1-30 cm thick.	Alternating storm current deposition of hummocky sand and fairweather suspension deposition of silt, bioturbated siltstone indicates the substrate was recolonized by organisms during fairweather times
Five coarsening-upward sequences, each consists of interbedded siltstones and hummocky sandstones overlain by horizontally-laminated or cross-stratified sandstones	1-8 m coarsening-upward sequences reflect the variable proximity of the sand source - sand ridges or patches **INTERPRETATION OF ENTIRE SEQUENCE** Migration of tide/storm maintained sand ridges and patches over storm-dominated areas on a stable shelf

Figure 32. Overall coarsening-upward sequence of the Erwin Formation. Shcs → Sms → QA-Erwin, with brief lithologic descriptions and interpretation of each facies.

size ranges from medium- to very coarse-grained (Fig. 33). The combination of silica cementation and pressure solution makes the rocks of Facies QA particularly well-indurated. The near absence of siltstone beds makes Facies QA thickly to very thickly bedded and results in weathering as massive ledges.

This facies is difficult to describe and interpret because of the pervasive pressure solution, absence of fine-grained layers and the fairly uniform grain size. Internal stratification is only rarely visible and well-defined cross-sets are uncommon. However, the faintly visible stratification features of QA and its relative stratigraphic position together yield enough information to formulate some reasonable interpretations for each of the two subfacies.

Facies QA can be divided into two subfacies based upon internal stratification and facies association. Subfacies QA-Erwin consists of large-scale, planar-tabular cross-beds and is associated with Facies Shcs and S in the Erwin Formation. It occurs only three times as Units 30, 35, and 37 with thicknesses of 5.0 m, 1.0 m, and 9.0 m respectively (Fig. 6, p. 15). On the other hand, Subfacies QA-Unicoi is characterized by low-angle and large-scale, planar-tabular cross-stratification. It is associated with Facies G and Facies S in the Unicoi Formation (Fig. 5, p. 14) and comprises Units 6 and 14, 10.2 and 8.3 m thick respectively.

The similarity between these two subfacies is their supermaturity, which indicates intense reworking within their depositional environments. However, the dissimilar facies association of each subfacies

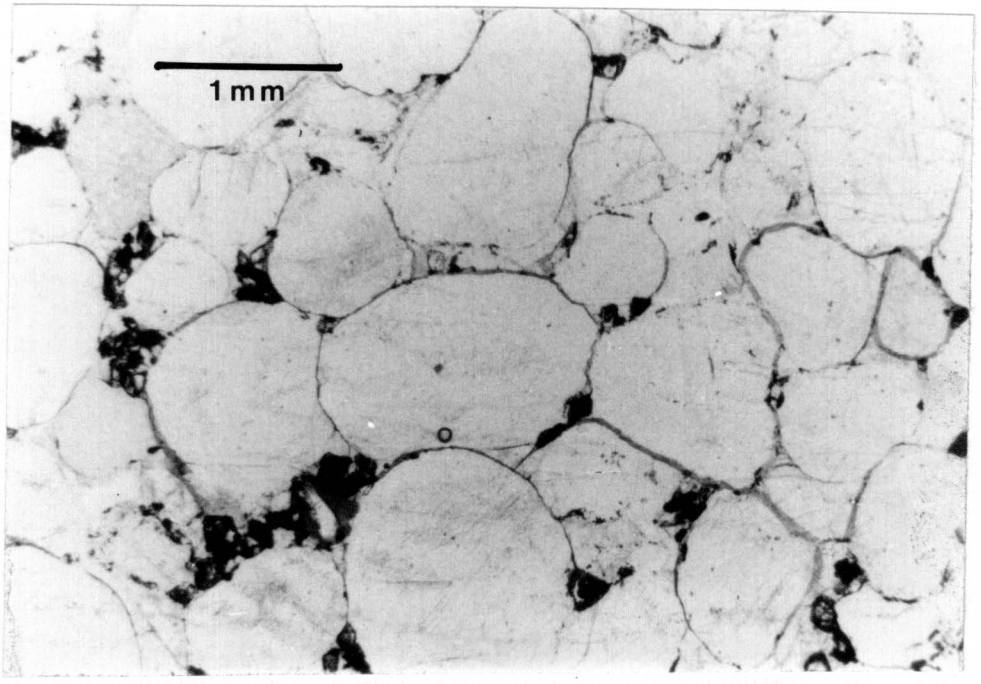


Figure 33. Photomicrograph of a typical Facies QA lithology showing textural and mineralogical maturity. Compare with Fig. 9 (p. 21) the photomicrograph of Facies G lithology. Sample 4-29-3 E. Unit 35.

demands two different environmental interpretations. Two interpretations of one facies seem to represent a violation of the traditional genetic implications of a facies. The logic of grouping Subfacies QA-Erwin and QA-Unicoi into Facies QA lies in their maturity and therefore, the fact that both environments must have been similarly characterized by extensive reworking.

Subfacies QA-Unicoi

This subfacies occurs in Units 6 and 14 (10.2 m and 8.3 m thick) within the Unicoi Formation. Unit 6 overlies Gp of Unit 5 but does not truncate these beds. It contains sets of large-scale, planar-tabular cross-stratification and low-angle cross-stratification (Fig. 34). Unit 6, unlike Unit 14, has several low-relief erosional surfaces that divide the unit into several 0.7-1.0 m beds. Along the scour surfaces are thin, discontinuous layers of red, ferruginous siltstone with thin laminae of heavy minerals and scattered granules. The siltstone commonly drapes the underlying rippled surface, which has symmetrical interference ripple marks. The red color is due to the iron-staining by hematite of diagenetic clays that fill the pores and replace feldspar grains. Above the scour surfaces are thin discontinuous granule lenses. Unlike QA-Erwin, QA-Unicoi is not associated with a glauconitic facies.

The base of Unit 14 is erosional and truncates the underlying beds of Facies S. It shows less than 1 m of relief along a broad scour that



A.



B.

Figure 34. Photograph and explanatory sketch of QA-Unicoi (Unit 6) showing ripple marks, large-scale planar-tabular cross-stratification and low-angle cross-stratification. Unit 6.

extends the length of the outcrop (Fig. 35). Low-angle cross-laminations are present although the dimensions of the sets are indeterminable. Unit 14 weathers as a massive ledge.

Subfacies QA-Erwin

Subfacies QA-Erwin consists of large-scale, planar-tabular cross-stratified quartz arenite. It occurs three times and in association with Facies Shcs and S. The sharp erosive bases of each unit truncate underlying strata at low-angles and these scours extend for 10-15 m with only 0.5 m of relief. Internally, these beds appear massive but from a distance, large-scale, planar-tabular cross-stratification can be seen with 30-50 cm thick foresets. No small- to medium-scale stratification within the foresets is visible. The tops of these three units are smooth and no small-scale bedforms are preserved on them. Each unit thins laterally to as much as one-half of its greatest thickness (Fig. 36). Each of the three units has a less than 5 cm thick, very coarse-grained lag at its top and is overlain with 3-20 cm of nonfissile shale with lenses of glauconite and very coarse quartz sand, and laminae of fine sand.

Summary of Vertical Facies Relationships

Vertical facies relationships are summarized on Fig. 37. At the first outcrop locality, Facies G and Subfacies QA-Unicoi are associated with one another in the lower 150 m of the outcrop. In the upper 50 m of the Unicoi Formation Facies S is the dominant facies, but includes two units (Units 20 and 23) of Facies G occur here.



Figure 35. Unit 14 of QA-Unicoi illustrating broad low-angle scour. Unit 14.



Figure 36. Unit 30 of QA-Erwin illustrating the thickly bedded nature of the unit as well as the apparent lack of cross-stratification and lack of fine-grained beds. Unit 30.

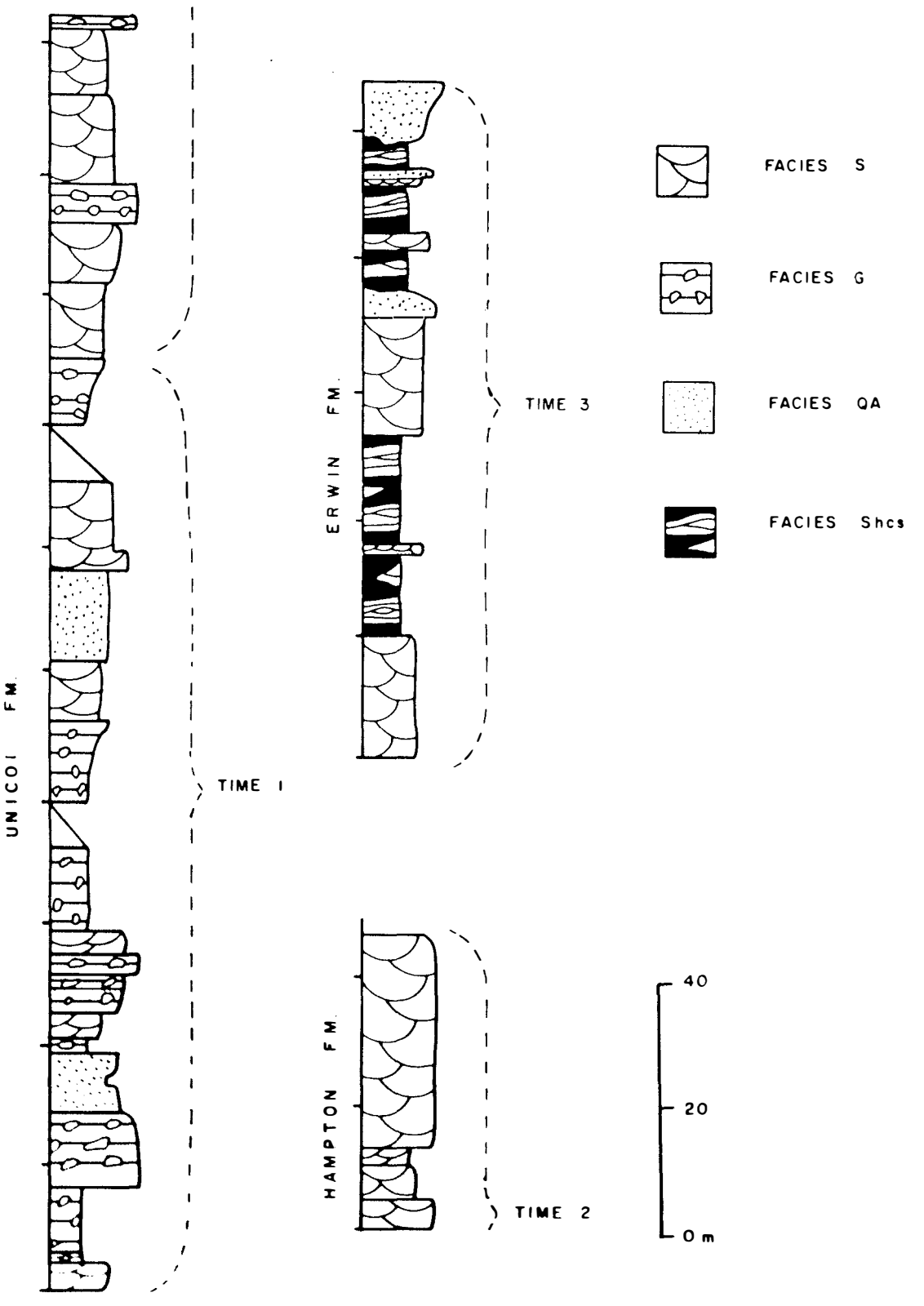


Figure 37. Vertical facies relationships of the Chilhowee Group. Facies symbols are given on Fig. 3 (p. 10).

At the second outcrop locality, Facies S occurs alone in the Hampton Formation. The vertical facies sequence of the Erwin Formation is exposed at the third locality, and an obvious vertical pattern exists. This is in direct contrast with the apparent random vertical sequence of the Unicoi Formation. The Erwin Formation consists of two coarsening-upward sequences: Shcs → Sms → QA-Erwin and Shcs → QA-Erwin. At the base of the Erwin Subfacies Sms is abruptly overlain by Shcs, which begins the coarsening-upward sequence.

III. INTERPRETATIONS

Facies G

General Alluvial Depositional Environment

Although the three subfacies represent specific processes and therefore different subenvironments, a few general interpretations can be made based upon the textural and compositional immaturity of Facies G. Texturally, the abundant matrix and poor sorting indicate an absence of reworking of an initially poorly sorted, mud-rich deposit. Because the most intense sediment reworking is in the shallow-marine or eolian environments, and the least amount of reworking in fluvial or deep-marine environments, it seems likely that Facies G rocks were deposited within a fluvial system. This interpretation will be more fully developed with the process-response interpretations of each subfacies.

Compositional immaturity is reflected in the abundance of unstable mineralogic components such as feldspars and lithic fragments. The presence of feldspar indicates deep weathering and rapid erosion of a sourceland characterized by potassium feldspar-bearing granites and gneisses. The presence of foliated polycrystalline quartz fragments indicates that metamorphic rocks also occurred in the sourceland. In addition, the abundance of fine- to medium-grained sedimentary rock fragments in Subfacies Gh indicates the presence of sedimentary rocks in the sourceland. Thus, based on composition, Facies G is interpreted as the result of erosion of a mixed source terrain. However, the abundance of feldspar is not only related to source area but also the climate within the sourceland (Pettijohn et al., 1973).

Great amounts of feldspar are most likely to be preserved in sedimentary rocks due to one of two possible climate regimes in which feldspar does not completely weather to clay. In arid or cold climates, the chemical weathering process is inhibited such that the incompletely weathered or fresh feldspar would become part of the sediment produced. However, arkosic sands may also be produced in warm, humid areas with high relief, which also experience rapid erosion. In humid areas, water is available to hydrolize feldspar, therefore yielding abundant clay. Under conditions of high relief and rapid erosion, both fresh and partly weathered feldspars are incorporated in the sediments.

The middle Late Cambrian paleogeographical reconstruction places the east coast of the United States within 15-25° south latitude (Ziegler et al., 1979) (i.e., the tropics or subtropics). The sourceland for the Chilhowee Group was therefore probably located within a warm humid climate. The arkosic sediment may reflect the relief (high, proximal) and/or climate (warm, humid) of the sourceland.

Orogenic sediment is generally more feldspathic than sediment related to erosion and deposition within tectonically stable areas. This would lead to the conclusion that a proximal, high-relief source is necessary for the generation of large quantities of feldspar. However, sands of the Mississippi delta still contain 20% feldspar (Pettijohn et al., 1973). Thus, the presence of feldspar may not necessarily imply a short distance of transport.

Another factor that may have the greatest influence on the generation of large quantities of detrital feldspar is the absence of

terrestrial vegetation. When vegetated slopes are stabilized, the feldspars of the bedrock weather to clay, a soil horizon is developed, and erosion is greatly reduced. The abundance of feldspar in the basal Chilhowee may therefore more accurately reflect the climate and degree of erosion in the sourceland rather than its relief and proximity to a depocenter.

The absence of vegetation would also affect the processes within a fluvial depositional system per se, as well as in its sourceland. Cotter (1978) extensively reviewed the literature and documented a mid-Paleozoic shift in fluvial style from nearly all braided to a mixture of braided and meandering. He attributes the shift to the advent of land vegetation (Silurian-Devonian). Vegetation affects stream geomorphology by: (1) retarding erosion; (2) decreasing sediment yield; (3) decreasing total runoff, discharge and flood peaks for a given amount of precipitation; (4) decreasing bedload grain size and enhancing fine sediment production, and (5) by increasing bank stability. All these factors would enhance the tendency of streams to meander.

Development of braided streams, although not well understood, is favored by rapid discharge fluctuations of greater magnitude than in meandering rivers. Braided rivers tend to have steeper gradients, coarser loads and more easily erodable banks than meandering rivers (Rust, 1978).

Facies G is interpreted as a braided stream deposit, based upon its pre-Devonian age and its immature lithology. With the general depositional setting as a frame of reference, each subfacies can be

interpreted in terms of its hydrodynamic origin and its position within a braided river system.

Modern braided rivers consist of broad, shallow channels and bars, with elevated areas active only during floods. Braided streams are developed in areas with large magnitudes of variation in discharge such as glacial outwash areas (e.g., sandurs of Iceland and Alaska, Tana River, Kicking Horse River, South Saskatchewan River); humid alluvial fans (Platte River, Brahmaputra, Kosi River Fan); and wadis of semiarid regions. A wadi is a stream in the desert environment that is subject to violent flash floods (Reineck and Singh, 1980); the channels are braided, and thus their deposits may resemble other braided stream deposits (McKee et al., 1967). From these modern environments, workers have developed models for deposition within a braided system. Important processes include bar formation, channel-floor dune migration, low-water accretion and overbank sedimentation (Miall, 1977).

Subfacies Gp: Transverse Channel Bars

The large-scale, planar-tabular, cross-stratified conglomerate of Subfacies Gp represents migration of large transverse channel bars (i.e., sand waves) of a distal reach in a braided stream system (Reineck and Singh, 1980). Cross-stratified, clast-supported gravel is the dominant lithofacies in distal braided streams, whereas horizontally bedded, imbricate gravels are common in proximal reaches (Rust, 1981). The sharp flat base of the foreset indicates erosion in front of a channel bar as it migrates during high flood stage. However, no scour trough or

gravel lags are associated with the cross-sets of Gp. The poorly sorted and crudely laminated bottom set indicates deposition of silt- to gravel-sized sediment from suspension as well as from bedload. The steeply dipping, graded foresets (dip angle approximately 25°) are characteristic of channel bars which possess active slip faces. Bluck (1974) documents large-scale, planar-tabular cross-bedding in coarse-grained sediment produced by the migration of steep accretionary bar margins with graded foresets in sandur deposits of Iceland.

Sand waves have well-defined avalanche faces, straight crests, a high length-to-height ratio, and are internally planar-tabular cross-stratified. There is an absence of uneven scour in the troughs and stoss sides are flat and featureless or covered with small ripples (Harms et al., 1982). Sand waves develop at a velocity less than that for dunes but greater than that for ripples. In modern braided rivers, the size of sand waves varies even within a particular river. In the Tana River of Norway (Reading, 1978), sand waves range from 200-300 m long, 200 m wide and up to 2 m high. These bedforms are large enough to generate large-scale, planar-tabular cross-stratification similar to that seen in Subfacies Gp.

Minimum flow conditions necessary to form such bedforms are difficult to evaluate based solely on flume studies. Flume experiments exploring the relationships between bedform configuration, sediment size and flow conditions encompass flow depths less than 1 m, flow velocities up to 3 m/sec, and sand-sized sediment (0.1-2 mm) (Harms et al., 1982). Speculation on flow conditions for sand waves at least 2 m high (minimum

bedform height equals the thickness of the cross-set) consisting of coarse to very coarse sand, granules and pebbles must use a combination of (1) conditions extrapolated from flume studies and (2) observed conditions of natural flows. This combination of experimental and observed data may be found in Fig. 2-8 of Harms et al. (1982).

Sand waves with a grain size of 1 mm (1 mm is the maximum grain size from the diagram in Harms et al., 1982) and that are at least 2 m in height form in flow velocities from 0.75-2.0 m/sec and in flow depths from 2-100 m. It is probably more instructive to consider flow depth and velocity in modern braided stream channels in which sand waves form. The depth, velocity and grain size data from several modern braided rivers are summarized in Table 2. Flow depths range from 0.4-15 m and velocities from 0.6-6 m/sec. Sediment size ranges from fine sand to pebbles.

Flow depths necessary to deposit Subfacies Gp most likely ranged from 2-15 m. Depths greater than 15 m are probably unreasonable because braided streams are characteristically shallow.

Estimating flow velocity from depth, grain size and minimum bedform height from Fig. 2-8 of Harms et al. (1982) yields a minimum value because (1) the height of the bedform is a minimum and (2) the size of the sediment transported during peak flooding is limited by the sediment available (Miall, 1977). For example, the 1965 flood on Bijou Creek (McKee et al., 1967) deposited only fine to coarse sand, yet measured flood velocities reached 6 m/sec. Therefore, minimum flow velocity necessary to deposit the sand waves of Gp is estimated to have been between 1-2 m/sec, but was probably much greater.

TABLE 2. Velocity, bankfull depth, and grain size for some modern braided streams.

RIVER	BANKFULL DEPTH (m)	FLOOD VELOCITY (m/sec)	GRAIN SIZE	SOURCE
Brahmaputra	15	2.4	fine-very coarse pebby	Coleman (1969)
Ganges	<10	3.4	—	Singh and Kumar (1974)
Donjeck	3	3.6	fine-pebby	Rust (1972)
Tana	15	—	fine-very coarse pebby	Collinson (1970)
Skeidarasandur distributaries	2	0.6-2.5 depends upon tidal stage	sand	Hine and Boothroyd (1978)
Scott outwash fan, measurements from active stream channels	0.4-0.06	1.5	pebbles (<10 cm)	Boothroyd and Ashley (1975)
Yana outwash fan	—	2	sand-pebbles (<10 cm)	Boothroyd and Ashley (1975)

Subfacies Gmr: Low Flood-Stage Pool Deposits

The close association of cross-stratified pebbly sandstone with laminated siltstone in Subfacies Gmr indicates large and rapid fluctuations in flow regime. Gmr represents deposition on top of and between transverse channel bars, where medium-scale coarse-grained bedforms develop with falling flow stage and large bedforms cease to migrate (Boothroyd and Nummedal, 1978). During low-water stands, pools develop between large bars and here silt may settle from suspension and drape the underlying megarippled surface. This relationship between bedforms and flow stage in braided systems has also been documented by Smith (1970, 1971), Hine and Boothroyd (1978), and Miall (1977) in many modern systems.

Gmr commonly overlies Gp in a vertical sequence or it occurs by itself as stacked megaripples with siltstone caps. Both situations represent sediment accretion as flow stage wanes. The first case results from superimposed bedforms on channel bars (Bluck, 1974; Reineck and Singh, 1980); the second case represents migration of megaripples over each other, possibly in shallow channels on bar surfaces, or on the floors of large channels at low flow stages (Boothroyd and Nummedal, 1978).

Subfacies Gh: Vertical Accretion Deposits

Subfacies Gh comprises 71% of the stratigraphic thickness of Facies G and consists predominantly of horizontally laminated sandstone. Less abundant features include low-angle sets of laminated sandstone, cross-

stratified conglomerate-filled scours, unstratified conglomerate lags and small-scale trough cross-stratification.

Evenly laminated sand may develop in one of several ways: (1) beach swash, (2) upper plane bed, (3) lower plane bed, (4) sedimentation from suspension clouds, (5) migrating wind ripples, and (6) migration of low relief (<1 cm) bedforms (Reineck and Singh, 1980). Deposition on a plane bed within the upper flow-regime is the most plausible mechanism for the thick (up to 8 m) layers of horizontally laminated sandstone. The presence of labile rock fragments and clay matrix in Gh discounts deposition within a high-energy beach swash zone, in which swash and backwash round and sort the sediment, leaving only quartz. Deposition within a wind regime is also unlikely because of the effectiveness of wind as a sorting and rounding agent. Sedimentation from suspension can generate laminated sand, however, normal grading should exist, if there was grain size variation in the suspension cloud initially. However, Gh contains granules and pebbles that define the laminae or are randomly scattered throughout the fine-grained sandstone. The granules and pebbles do not appear to deflect the laminae as if they had settled from suspension, and there are also no graded layers.

Horizontal stratification has been documented as forming by foreset accretion of small amplitude sand waves (<1 cm) in water depths less than 2 cm (Smith, 1971). Internally, the laminae are graded from coarse- to fine-grained; the laminae of Subfacies Gh do not exhibit grading on such a fine scale. In the Platte River (where these small bedforms were

observed), the thickness of laminated beds is only 6 cm, and no appreciable accumulation was observed.

The horizontally laminated sandstone of Gh most likely was deposited on a plane bed in upper flow-regime conditions, with abundant sediment carried in suspension and as bedload. Parallel laminations and low-angle cross-laminations have been documented by Reugg (1977) in the Pleistocene sandur deposits of the Netherlands; by McKee et al. (1967) in modern overbank deposits of Bijou Creek, Colorado; and by Boothroyd and Ashley (1975) on braided outwash fans of the Gulf of Alaska.

Bijou Creek is an ephemeral stream where the dominant sedimentary structure (90-95% of flood deposit) generated by a flash flood event is horizontally laminated, medium-grained sand. Because bridge girders were also carried by flood waters, McKee et al. (1967) attributed the laminated sand to deposition within the upper flow-regime. The laminated sand covers the channel and extends beyond its banks for 420-460 m. Its thickness ranges from 0.7 m to a maximum of 3.5 m. However, horizontally laminated sand is not a major facies in many modern and ancient braided stream systems (Miall, 1977).

Gh is not composed entirely of horizontally laminated sandstone, although a monotonous 8 m thick zone occurs in Unit 12 (Fig. 5, p. 14). The other sedimentary structures present include (a) small-scale trough cross-stratification, (b) unstratified conglomerate lags, and (c) cross-stratified, conglomerate-filled scours. These structures indicate periodic fluctuations in flow regime, that is, either changes in velocity or flow depth, or both.

The typical sequence is about 0.5 m thick and commences with a conglomerate-filled scour that may or may not be cross-stratified. The scour indicates the beginning of a high-energy event and, depending upon flow conditions, medium- to coarse-grained sediment may be winnowed away leaving behind a lag. Alternatively, sediment may be deposited as bed-load within a scour. The presence of scouring currents is also indicated by shale rip-up clasts in Units 12 and 17. The filled scours are overlain by low-angle to horizontally laminated sandstone. Within the low-angle cross-stratified sandstone are low-angle truncation surfaces, which resulted from scour of a previous deposit prior to deposition during the next flood event. The sequence is rarely capped with small-scale trough cross-stratification and the entire unit indicates deposition under a decreasing flow regime.

Rust (1978) documented the presence of horizontally laminated sandstone and sets of low-angle cross-stratified sandstone in the braided stream deposits of the Malbaie Formation of the Northwest Territories. He attributed both structures to upper flow-regime conditions, but indicated that the low-angle sets formed in shallow scours. Miall (1978) attributed the low-angle sets to crevasse-splay deposits as well as deposition within scours. Although this sequence is not common in Subfacies Gh, its presence indicates that Gh was deposited by an amalgamation of flood events. Because structures indicating lower flow-regime would have the lowest preservation potential, thick accumulations of laminated sands could very well develop.

McKee et al. (1967) described lower flow-regime structures (planar-tabular cross-stratification) from the Bijou Creek flood deposits of 1965. These structures were formed at the distal margins of the overbank deposit where flow depth and velocity were less than conditions nearer and within the main channel. They also described climbing ripple cross-lamination and convolute laminae which formed during waning flood stage. So within a single flood event, lower flow-regime structures would also be expected to develop, although they may compose only a small portion of the total stratigraphic thickness of the deposit.

The thick accumulations of Gh therefore are interpreted to be vertical accretion deposits from an alluvial plain, and are the result of upper flow-regime/sheet-flood conditions. The overall coarsening-upward trend of Units 3, 4, 12, and 17 (Fig. 16, p. 29) from interbedded silty sandstone to conglomeratic sandstone may be a reflection of the proximity of an active braided channel. The glacial outwash plains of southern Iceland (Hine and Boothroyd, 1978) extend to the coastline and parts of the plain are inactive, although they may be innundated by wind tides, which consist of freshwater from streams and ground water seepage. As a result of this flooding, wide flat zones are formed, which have lost their original braided stream topographic features. The coarsening-upward trend may reflect the lateral migration of the active braided zone across a wind tidal flat.

The laminated siltstone and ripple cross-laminated sandstone of Unit 3 and at the base of Unit 12 resulted from deposition from suspension and under lower flow-regime conditions in a lagoon or

saltwater pond. Marine and terrestrial processes coexist on the wave-dominated coast of Iceland (Hine and Boothroyd, 1978) and on the Yallahs fan delta of Jamaica (Wescott and Ethridge, 1979); on the southeast coast of Alaska (Boothroyd and Ashley, 1975); and on the Copper River delta (Galloway, 1976). In Iceland, freshwater ponds were noted on the inactive alluvial plain behind wave-worked beach sediment. The Yallahs fan delta has salt water ponds on its flood plain. In Alaska, the subaerial braided fan is separated from offshore bars or barriers by coastal lagoons. Because trace fossils are uncommon within the laminated siltstone and sandstone, the quiet-water environment was probably not chemically hospitable. Such a setting would be a salt water pond that was most commonly brackish due to rain-water dilution, and that gradually filled with flood-plain deposits as the braid channels migrated. The silty sandstone volcanoes represent dewatering of thixotropic sediments, a process which would be highly likely in muddy pond deposits.

Summary of Facies G Environment

The Subfacies of G: Gp, Gmr, Gh, are a result of deposition within subenvironments of a braided stream plain where active channels flow directly into an adjacent marine system. The braid plain was a mosaic of channels with large transverse bars, channel pool areas with megarippled gravels and silts deposited during waning flood stage, amalgamated sheet-flood deposits, and brackish flood-plain pond deposits.

Fig. 5 (p. 14) shows the random order of appearance of Facies G. Due to the rapid migration of bars and channels and the great extent of

the actively braided zone (up to 9 km on the coast of Iceland, Hine and Boothroyd, 1978), fining-upward cycles are not common in Facies G. Fining-upward cycles are common, however, in modern systems with considerable topographic relief on the braid plain in which vegetated high areas trap fine-grained sediment (Miall, 1982).

One of the most intriguing aspects of Facies G is its association with arenite beds of Facies S and QA. The significance of the association will be discussed after each facies has been interpreted.

Facies S

General Marine Depositional Environment

A marine environment is the most viable general interpretation of Facies S based upon its mineralogy, overall maturity and trace fossil assemblage. Facies S lithologies are mineralogically and texturally more mature than Facies G lithologies; although grain sizes range from fine sand to granules, these arkosic and subarkosic arenites exhibit moderate sorting and the grains are subrounded to rounded. Labile components such as rock fragments constitute less than 2-5% of any given sample. Cross-stratification is typically defined by concentrations of heavy minerals or by size grading. Depositional matrix is absent except within bioturbated beds where biogenic activity homogenized siltstone and sandstone beds. The submature to mature character of this facies probably resulted from reworking of fluvial sediment that was deposited directly on the shoreface by tides and/or waves. Less resistant rock fragments did not survive marine reworking, whereas some feldspar grains did.

Other features also indicate a marine origin for Facies S. Facies S of the Hampton contains marine trace fossils (rare Skolithos, paired vertical burrows, horizontal burrows), and within the Erwin, Facies S is locally glauconitic. The units of Facies S occurring within the Unicoi do not exhibit bioturbation and do not contain glauconite, however, their sedimentary structures and sequences are identical to those in the two younger formations.

The subordinate amounts of mudstone (Sfl) that make up 5-10% of the total stratigraphic thickness of Facies S indicate deposition in a setting not only with abundant current agitation but also with periodic fluctuations of current strength; these conditions would most likely be found within a shallow-marine environment. Facies S throughout the Chilhowee Group is interpreted to be of shallow-marine origin. A brief discussion of modern shallow-marine processes and environments is needed to formulate a reference frame in which to place each subfacies.

Broad areas of the shallow-marine environment that are agitated and can accumulate large thicknesses of clean, submature to mature sandstone are at coastlines and on continental shelves. The modern coastline extends to about 10-20 m of water depth (Reineck and Singh, 1980) and may extend several kilometers seaward. Environments of the coastline are varied and include barrier islands and tidal inlets, back-barrier lagoons, offshore or longshore bars, and subtidal channels and shoals. A profile of simplified coastal morphology is illustrated in Fig. 38 and shows five broad zones. The distinction between shoreface, nearshore

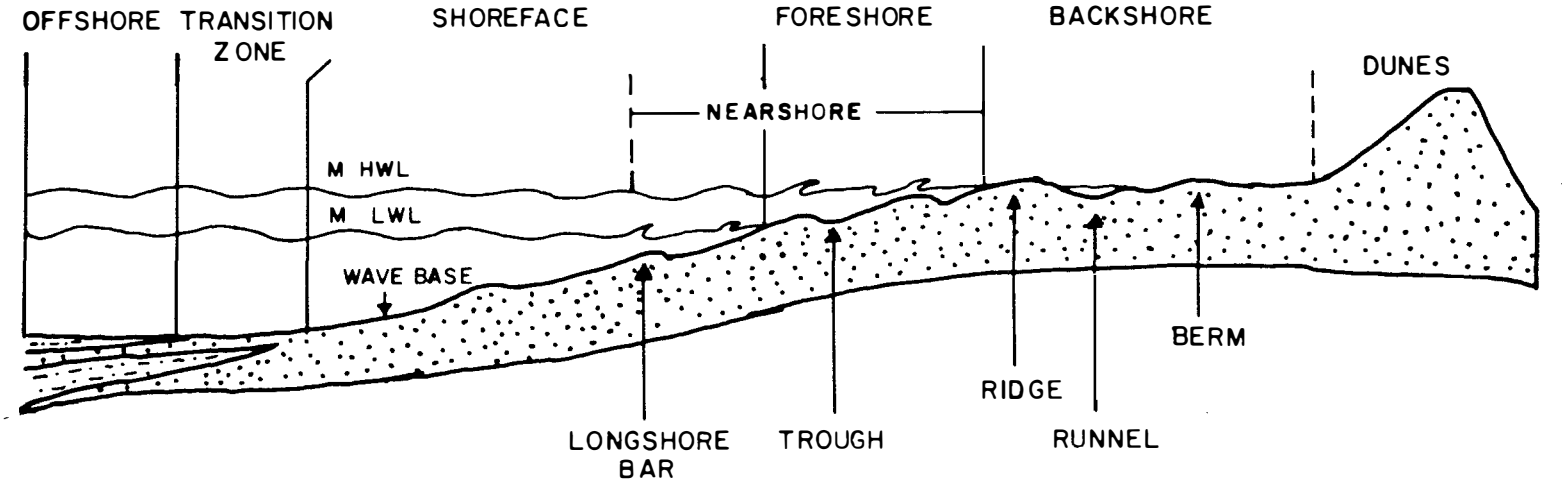


Figure 38. Simplified coastal profile for the purpose of standardizing terminology.

and foreshore is particularly important and requires explanation because those terms will be used frequently in the following discussions.

All the zones are defined based on depositional process rather than absolute water depth or distance from shore. For example, the depth or distance from shore at which waves begin to impinge on the bottom (i.e., boundary between offshore and shoreface) depends upon the width of the shelf and the local wave regime; these zones do not remain spatially stationary through time. During storms, wave base is lowered; thus the nearshore zone is broadened during stormy weather.

The landward-most area of the backshore zone consists of eolian dunes; seaward of the dunes the backshore zone exhibits a mixture of eolian features as well as storm-wave generated features that occur above mean high-water level. The foreshore occurs between mean high- and mean low-water level and is dominated by beach swash features. The shoreface extends from mean low-water seaward to the point where shoreface sands interfinger with offshore silt and clay within the transition zone. The nearshore zone is a subzone of the shoreface and is the area affected by breaking waves. It is characterized by having one or more longshore bars. The foreshore and shoreface are affected by a combination of waves, especially storm waves, and tides.

The shelf is the subaqueous region that extends from the transition zone seaward to the continental slope. The shelf is the offshore zone of Fig. 38 and may range in width from 6 km (parts of the Oregon shelf) to 200 km (the Atlantic shelf off the coast of Georgia) (Harms et al., 1982). Currents acting on the shelf include tidal currents, storm-wave

currents, density currents, and intruding ocean currents (Swift et al., 1972). Due to all the possible processes and combinations of processes that occur on the shelf, it is the most complex of all major siliciclastic environments. Understanding the processes and deposits of modern shelves is hampered by the effects of Holocene sea-level rise and by the lack of accessibility to the shelf. Hence no well-developed facies models have yet been proposed (Walker, 1984). Swift et al. (in press, fide. Walker, 1984), however, distinguish three main types of shelf systems: (1) tide-dominated; (2) storm-dominated; and (3) intruding ocean current-dominated.

The following interpretations of each subfacies will refer to processes and structures observed on modern coastline and shelves and in ancient shallow-marine sequences. Shallow-marine refers to both coastal and shelf environments.

Interpretation of Physical Sedimentary Structures

Subfacies Sfl: slack water deposits. Because Sfl overlies beds of Sms and to a lesser extent Sls, its interpretation needs to be discussed first. Interpretations of the structures of the other two subfacies will incorporate that of Sfl.

The conditions necessary for the generation of the wavy and lenticular bedding that characterizes Sfl are periods of current activity alternating with quiescence (Reineck and Singh, 1980). Waning currents may deposit rippled sand layers, and during a subsequent slack period, silt and clay deposited from suspension drape the ripples.

Slack water provides organisms with an opportunity to recolonize the substrate, resulting in moderate to intense bioturbation.

The character of the alternately deposited silt and sand depends also upon the availability of the appropriate sediment size. Wavy bedding is produced when both silt and sand are available in equal amounts and both layers are preserved in equal amounts. Lenticular bedding, in which sand lenses are isolated vertically within mud, is produced under conditions of meager sand supply so that only incomplete ripples are formed and mud is deposited and preserved (Reineck and Singh, 1980).

The environments where there is a regular change between turbulent and slack water are the intertidal to subtidal zones. Storm events as well as tidal rhythms may provide the alternating flow conditions and sediment supply. The minor stratigraphic abundance of Sfl (less than 5%) and its association with Sls and Sms are features which do not indicate a storm origin.

Some question has arisen in the literature as to whether the time span of slack tide is sufficient for mud to settle from suspension (Levell, 1980). Wunderlich (1978) documented that a 2 cm layer of mud was deposited within 30 minutes of slack tidal water in Jade Bay of the North Sea. Rapid extensive deposition of mud is facilitated by high concentration of suspended mud (McCave, 1970, 1971). Deposition of silt and clay is also enhanced by flocculation of particles; a mud floc may then behave as a sediment particle with a much greater settling velocity than the individual silt or clay grain.

Subfacies Sfl is therefore interpreted to represent deposition from suspension or under lower flow-regime conditions during slack and waning tidal flow.

Subfacies Sls: subtidal channel fill. Subfacies Sls is a result of deposition within broad subtidal channels dominated by storm-enhanced tidal and/or rip currents. Although no channel-shaped scour surfaces were observed within Sls, a basal coarse-grained lag records the scour over which large bedforms migrated. The large-scale trough cross-stratification that typically occurs as stacked cosets most likely resulted from the migration of megaripples (for terminology see Appendix B).

Single large-scale planar-tabular cross-sets also characterize Subfacies Sls. These features may be attributed to the migration of flow-transverse sand waves located within the deepest parts of a subtidal channel (Kumar and Sanders, 1974; Reineck and Singh, 1980; Reinson, 1984). Thickly bedded, coarse-grained to pebbly beds that coarsen upward probably represent deposition by the highest-energy currents generated from storm-enhanced flow in tidal or rip channels. The crude coarsening is a result of gradual waning of a sediment-charged current. As the competence of the current decreased, coarser grains were left behind (Reineck and Singh, 1980).

These inversely graded and unstratified beds are associated with rocks of Subfacies Gp in the upper part of the Unicoi Formation. Given the proximity of the subtidal zone to an actively migrating channel, it

is highly likely that these sediment-charged flows were associated with unusual flooding and high river discharges. A corollary to this observation and interpretation is that submature sediment of Sls represents immediate reworking of terrestrial sediment that was either deposited on the shoreface or was eroded from the braided coastal plain sediment.

Laminated sandstone and siltstone commonly cap a set or coset of Sls and in rare cases, they are present as a lense along a foreset. Clay drapes are fair indicators of a tide-dominated environment in which flow fluctuations are the norm (Mowbray and Visser, 1984). Reactivation surfaces, erosional discontinuities between foresets, may form as a result of tidal current flow reversals (Klein, 1970; McCabe and Jones, 1977). They were not observed in Sls, but may have been obscured by the coarse-grained and indurated nature of the subfacies. The rare association of Sfl indicates periodic flow cessation.

The paucity of interbedded siltstone and sandstone associated with Sls indicates one of two possible processes: (1) all sediment deposited from suspension was eroded prior to or during the migration of the bedforms or (2) that fines were never deposited because the subtidal dunes were above fairweather wave base and the water column was always in motion. In either case, very fine-grained sediments would have been transported into the offshore zone. Because wave-formed features were not observed in Subfacies Sls, the former is the probable process.

Subfacies Sms: subtidal shoals. Sms is characterized by medium-scale trough and planar-tabular cross-stratification (Figs. 24 and 25, pp. 44, 45) which formed in response to a steady, unidirectional

current. Evidence for unsteady, bidirectional currents, such as reactivation surfaces, clay drapes, or back-flow ripples were not observed. Cross-sets are characteristically continuous across the outcrop and their thicknesses are reasonably constant, indicating constancy of flow conditions through space and time.

Rapid and marked velocity decreases resulted in the fining-upward sequence: Sms → Sfl. This low to zero velocity period could occur during slack tide (Klein, 1977; Reineck and Singh, 1980) or after a tidally or storm-enhanced rip current waned (Davidson-Arnott and Greenwood, 1976). With renewed current activity, the low-energy silt/sand layers are wholly or partially eroded as megaripple migration resumes.

The question that needs to be resolved concerning the origin of Facies S is whether the sediment was deposited in response to tidal currents, longshore and rip currents or a combination of the two types of regimes. Before addressing that question in greater detail, paleocurrent data of Facies S must be considered.

Paleocurrent interpretations. The paleocurrent data of Facies S were measured from foreset azimuths and ripple crest trends within Sms and Sls. There is little difference in paleocurrent direction between the two subfacies, so data from the two are plotted together in Fig. 11 (p. 23).

The data for each formation are unimodal but are widely dispersed between 0-180°. A minor mode is directed to the west within the Hampton. The westward dipping foresets lie above or below foresets whose

azimuths are 180° in the opposite direction. However, herringbone cross-stratification is not particularly common. Rare symmetrical ripple crests indicate bidirectional currents to the southeast and northwest. As previously mentioned, no cross-laminations were observable, therefore these ripple forms could have been produced by either oscillatory currents or by wave modification of current ripples. Because most of the crests are flattened or rounded, modification of the initial bedform is likely. Ripple crests are oriented obliquely to the strike of a subjacent foreset.

In general terms, widely dispersed, polymodal paleocurrent patterns characterize ancient shallow-marine sequences, and may represent the mixing of different current systems (Pettijohn et al., 1973). Tide-dominated settings have bimodal patterns with the largest mode reflecting the relative strengths of the dominant tidal phase (Klein, 1977), although as Klein notes, tidal cycles are characteristically unequal in terms of duration and magnitude of ebb and flood phases of a single cycle. In some instances, ebb and flood currents may follow mutually exclusive paths (Walker, 1984). The resultant paleocurrent pattern would be unimodal. Several ancient thick successions of sandstone interpreted to be tidal deposits, but which are characterized by a unimodal paleocurrent pattern include the Lower Cambrian Duolbasgaissa Formation of Norway, (Banks, 1973); the Cretaceous Lower Greensand of England and France (Narayan, 1971); the Precambrian Jura Quartzite of Scotland (Anderton, 1976); and the Late Precambrian Upper Quartzitic Sandstone Member of the Dakkovarre Formation of Norway (Johnson, 1977).

Paleocurrent patterns of ancient wave-dominated clastic shorelines are directed onshore, offshore and alongshore (Heward, 1981). However, in order to draw definitive conclusions about the paleocurrents, one really needs to know the location of the paleoshoreline, which is difficult to establish for the Chilhowee Group.

Studies by Whisonant (1974) and Schwab (1972) (Fig. 39) show that along its strike belt, the Chilhowee Group (and particularly the Unicoi Formation) exhibit east- and southeast-directed paleocurrent means. Paleocurrents of the braided-alluvial Unicoi Formation are interpreted to be parallel to the regional paleoslope that dips southeast. A north-south or a northeast-southwest trending shoreline may be postulated.

If indeed the shoreline was oriented as such, currents were directed offshore and alongshore, with a very minor mode directed onshore. This pattern is interpreted to reflect a mixing of tidal and wave influences on the shoreface. The tidal influence is recorded mainly by offshore directed azimuths and by infrequent herringbone cross-stratification. Scant evidence for flow reversals in the form of herringbone cross-stratification is not unusual given that tidal cycles are commonly asymmetric. The rarity of herringbone cross-stratification more accurately reflects the great strength or duration of the ebb-tidal phase, relative to the flood tide.

The influence of waves and longshore currents is reflected in the wide dispersion of the data and the modification of tidally formed current ripples. The relative influence of waves on the sedimentary facies is discussed in more detail in the synthesis of Facies S interpretations.

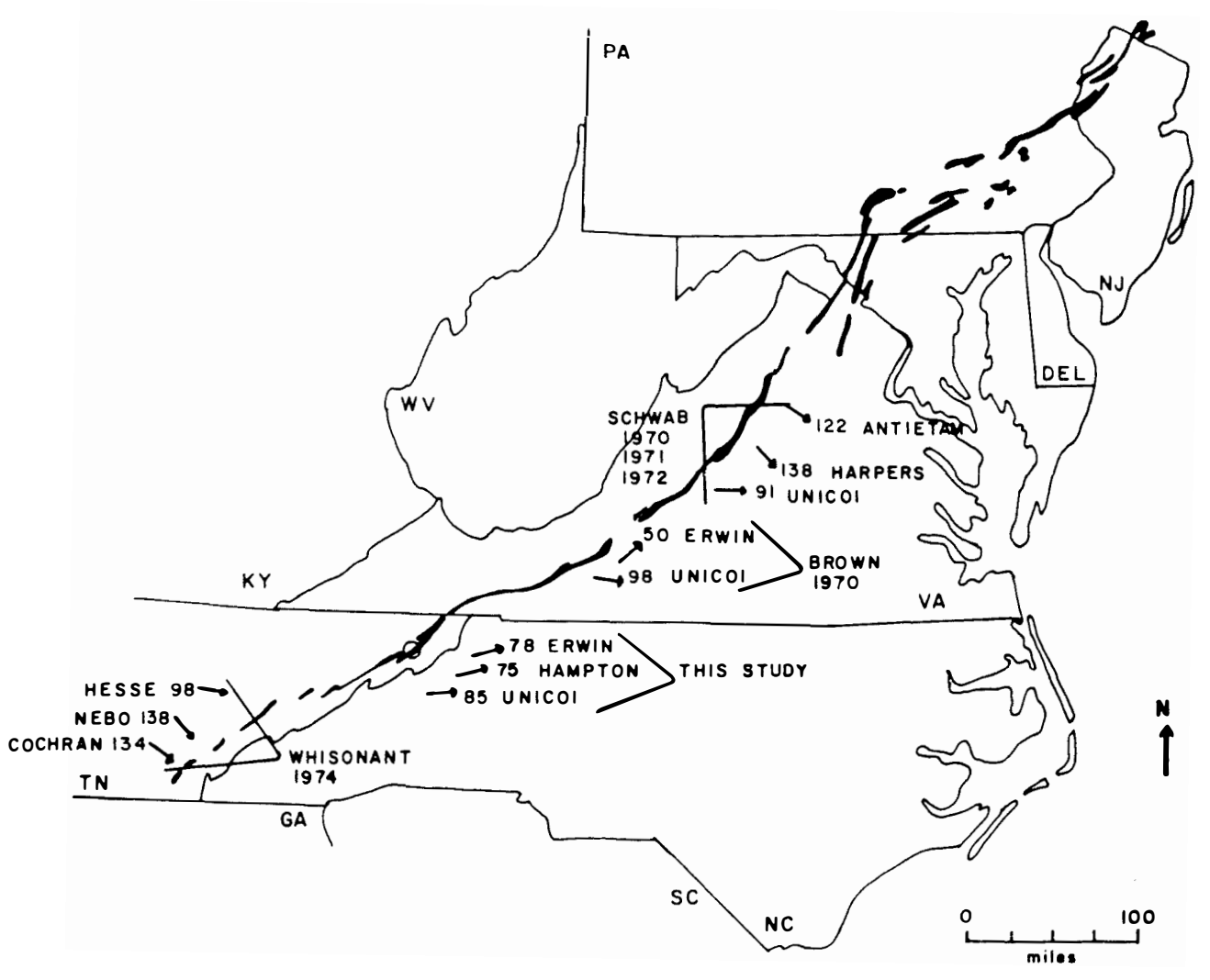


Figure 39. Regional paleocurrent summary for Tennessee and Virginia (modified from Schwab, 1972).

Interpretation of Biogenic Structures of Sms and Sfl

Because Sls does not contain any observable biogenic features, this discussion pertains only to Sms and Sfl. The interbedding of Sms and Sfl is so common that the two subfacies must be considered simultaneously. The stratigraphic ranges of the trace fossils are illustrated in Fig. 40 relative to the facies in which they occur. Ranges are for the Hampton and the Erwin Formations; Facies S of the Unicoi Formation is apparently biogenically undisturbed.

The behavior of organisms is a direct result of the environmental conditions to which they are adapted. Hence physical environments, defined by a set of relative energy conditions and sediment types, may be characterized by a certain ichnofacies or suite of biogenic structures. In this way, biogenic structures may add information to environmental interpretations based upon physical structures. The reverse, of course, is also true, because once physical conditions have been defined, behavioral responses can be postulated. Much environmental interpretation, based upon modern and ancient trace analogues, has been summarized by Frey and Pemberton (1984).

The traces present in the upper part of the Chilhowee Group belong to either the Skolithos or Cruziana ichnofacies (Seilacher, 1978; Frey and Pemberton, 1984). Vertical, cylindrical burrows, characteristic of the Skolithos ichnofacies, occur only rarely in Sms of the Hampton Formation and are absent everywhere else (Fig. 40). U-shaped burrows are probably present as well, although the only evidence for them is paired circular depressions on bedding surfaces. These vertical dwelling

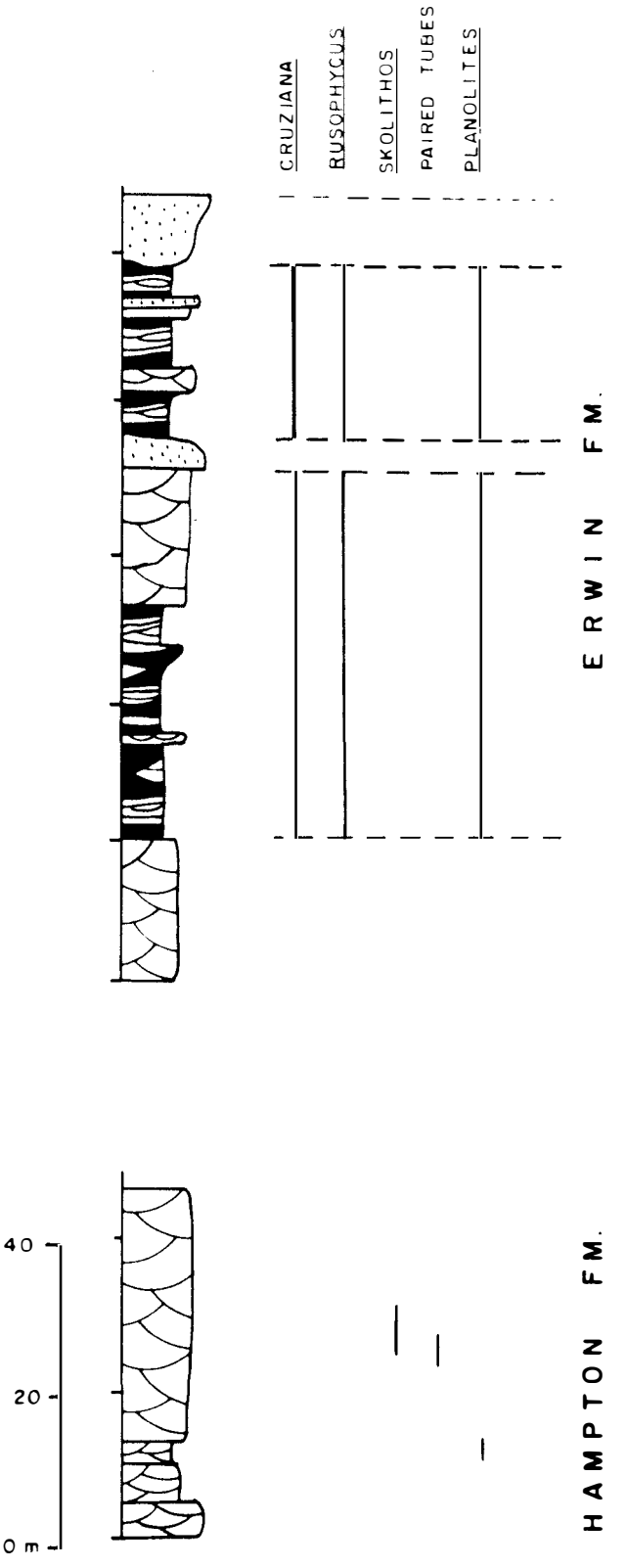


Figure 40. Stratigraphic ranges of trace fossils and their associated facies.

burrows of suspension-feeding organisms are common in ancient cross-stratified sandstones. This association of traces and lithology indicates that the organisms were adapted to a shifting substrate which experienced rapid erosion and deposition. The absence of a well-developed Skolithos ichnofacies in Facies S of the Unicoi and Hampton is probably due to intense physical reworking that obliterated biogenic structures and left a preserved record of physical stratification.

Traces characteristic of the Cruziana ichnofacies--Cruziana, Rusophycus, and horizontal cylindrical burrows--are present within Sfl of the Hampton and within Sfl and Shcs of the Erwin Formation. The resting, crawling, and feeding traces are characteristic of unconsolidated substrate, particularly between fairweather and storm wave base. Lithologies typically associated with this ichnofacies in the rock record are thinly interbedded siltstone and sandstone, indicating moderate energy levels alternating with quiet periods. These trace-forming organisms were adapted to feeding upon the nutrient-rich substrate, rather than a nutrient-rich, constantly circulating water column.

The most important insight the traces lend to the overall environmental interpretation of the Chilhowee Group is the ranges of ichnofacies. Their appearance in the Hampton suggests that marine sandstone overlies nonmarine conglomerate and sandstone (Facies G) of the Unicoi Formation; that is, a transgression has occurred. The shift from a Skolithos-type assemblage to a Cruziana-type assemblage from the Hampton to the Erwin results from increased water depth and slower sedimentation rate through time. Although Sms of the Erwin is similar to Sms of the

Hampton, the bioturbated, fine-grained interbeds of the Erwin contain Cruziana-type traces. During Erwin sedimentation, megarippled sand shoals experienced deeper water and slower sedimentation rates than they were subjected to earlier. Interbedding of Sms with Shcs (hummocky storm sands) and presence of glauconite within the Erwin also indicate that depositional conditions had changed through time.

The lack of Skolithos in the Erwin is a result of unsuitable energy conditions. Lateral lithologic equivalents of the Erwin Formation (Nebo and Hesse Quartzites of East Tennessee) contain abundant, robust Skolithos. This lateral relationship indicates that a fauna was available to colonize suitable substrates in shallower water depths. This is not meant to suggest that the Erwin Formation shallowed only to the southwest (shore-parallel). The Chilhowee Group of eastern Tennessee has been tectonically displaced relative to the Chilhowee Group of northeastern Tennessee and their palinspastic reconstruction is not well known. The lateral variation may just as easily reflect a shoreward shallowing.

Synthesis of Facies S Environments

The facies association Sls-Sms-Sfl needs to be interpreted as a composite environmental setting. As mentioned previously, Facies S is interpreted to be of shallow-marine origin based upon its overall maturity, presence of glauconite, presence of marine traces and its widely dispersed paleocurrent pattern. Its interbedding with Facies G (Fig. 37, p. 62) is indicative of a proximal coastline. As Facies S

becomes the dominant facies in the upper part of the Unicoi Formation, the terrestrial source was no longer exerting a substantial control on the lithologies of Facies S. This is reflected by a gradual upward-maturing of Facies S as the rock fragment component decreases in abundance.

Facies S is arranged in thinning- and fining-upward, 10-20 m thick sequences (Fig. 23, p. 42). Such sequences typify the filling of tidal channels (Reinson, 1984) as longshore currents deposit sediment on the updrift side of the channel and sediment is eroded from the downdrift side (Kumar and Sanders, 1974). The rate of migration of tidal channels or inlets is related to the magnitude of longshore sediment supply. The structures generated within the channel, however, are a result of tidal flow.

Kumar and Sanders (1974) found that sedimentary structures vary with depth in the Fire Island tidal inlet. The subenvironments of the channel include deep channel, shallow channel, spit platform and spit (with beach, berm, and washover features). The deep-channel sediment exhibits lower flow-regime, ebb-oriented cross-stratification with flood-oriented reactivation surfaces; the shallow channel is characterized by upper flow-regime, parallel laminations, analogous to flow within meandering fluvial channels (Walker and Cant, 1984).

Other studies of modern barrier inlets or tidal channels also indicate deposition under different flow conditions for the various parts of a channel. The deep-channel deposits of the mesotidal inlets of South Carolina (Hayes, 1980) have bidirectional, large-scale, planar cross-

beds; the shallow-channel deposits consist of bidirectional, small- to medium-scale trough cross-beds.

Reineck and Singh (1980) described the features present on the floors of subtidal channels of German Bay, in the North Sea. These channels occur within the tidal estuaries of German Bay. The channels are similar in dimensions to tidal inlets associated with barrier islands. The tidal channels of German Bay are as broad as 2 km and as deep as 15 m; Fire Island Inlet is of comparable width and about 10 m deep. The German Bay tidal channels, however, are separated from one another by sand shoals rather than by barrier islands. The water depth over the shoals is about 5 m. Current velocities in the channels are as great as 1.4 cm/sec, whereas the velocities over the shoals are much less. In the channel proper, which in some cases cuts into Pleistocene sand and gravel, large bedforms are well-developed. Such bedforms are sand waves (giant ripples in the terminology of Reineck and Singh, 1980; see Appendix B) and megaripples. Sand shoals between the channels are characterized by megaripple bedding, small-ripple bedding and by laminated sand deposited from suspension.

The best modern analog for Facies S is the tidal channel and sand shoal complex of German Bay. Marine sediment there overlies and cuts into Pleistocene sediment. The Pleistocene exposed on land in the Netherlands is braided glacial outwash (Reugg, 1977). In the German Bay area, transgression places shallow-marine deposits over glacial outwash fan and fluvio-deltaic deposits (Houbolt, 1968).

Lateral migration of the channels and shoals generated a thinning- and fining-upward sequence during the deposition of Facies S. A dominant ebb-tidal phase formed dominantly unidirectional paleocurrent patterns. Fine-grained sediment, deposited during slack tide, formed silt or fine sand layers on top of the bedforms. Subfacies Sf1 does not commonly overlie Sls, thus indicating that slack water deposits in the channel were not preserved.

Because this sequence could have been generated by tidal inlet migration, a single environmental interpretation may be presumptuous. The absence of barrier island and beach sediment cannot be used as evidence against a barrier island/tidal inlet interpretation. One reason is that the sediment deposited within the deepest zone of any depositional setting have the greatest preservation potential, because it is the least likely to be reworked (Heward, 1981). This is particularly important for deposition associated with transgression, where shallow and subaerial deposits are easily eroded and reworked. Because the facies association of the Unicoi Formation records a transgression, one would not necessarily expect a complete barrier sequence to be preserved. Another reason the barrier island/tidal inlet setting is not precluded is that without the stabilizing effect of vegetation during the Cambrian, barrier islands probably would not have been well-developed, much less preserved.

Regionally, a sandstone sequence of tidal inlet fill should be a linear sand body oriented parallel to the shoreline (Reinson, 1984). Tidal channel and shoal migration in a tide-dominated setting such as

the German Bay would probably generate current-parallel linear sand bodies. Because regional geometries of the facies of the Chilhowee Group have yet to be developed, both interpretations are equally viable.

In summary, Facies S represents deposition within subtidal channels or on adjacent shoals. Deposition of Sms in the Unicoi and Hampton Formations occurred above fairweather wave base, as evidenced by wave-modified current ripples. The particularly thick sequences of Facies S in the Unicoi is a result of stacking of tidal-channel fills. Sedimentation must have been keeping pace with subsidence to allow for such a thick accumulation without any evidence for progradation into deeper water. That is to say, none of the thinning-upward sequences overlies offshore or transition zone sediments. The dominance of lateral channel migration instead of sand shoal progradation may be a result of abundant sediment being supplied by longshore currents.

Facies Shcs

Ideal Hummocky Sequence: Lower Shoreface-Offshore Setting

Facies Shcs occurs only in the upper part of the Chilhowee Group and only within the Erwin Formation. The two most environmentally significant characteristics--abundant siltstone and equally abundant hummocky cross-stratified, micaceous, fine-grained sandstone--indicate a storm-dominated offshore or lower shoreface setting. The setting was likely between fairweather and storm wave-base. Glauconite is associated with Shcs and its presence also implies a quiet water (below fair-

weather wave-base) environment dominated by slow sedimentation (Odin and Matter, 1981).

A specific set of processes is indicated by the ideal stratification sequence of the sandstone of Shcs. The sequence is illustrated in Fig. 27 (p. 49) and is (bottom to top): (1) a scoured base (with or without parting lineations or tool marks); (2) a zone of flat to slightly wavy, laminated, very fine-grained sandstone, 1-5 cm thick; (3) a hummocky zone 5-15 cm thick with only 1 or 2 low-angle erosional surfaces and with antiforms spaced 10-30 cm apart; (4) a zone of flat laminae less than 5 cm thick; (5) a zone of ripple cross-laminations and symmetrical ripple forms less than 5 cm; and (6) a layer of bioturbated siltstone, 1-30 cm thick, with or without thin sandstone lenses/layers.

The processes that would generate such a sequence have been debated by Walker (1979), Harms et al., 1982, Dott and Bourgeois (1982), and Swift et al. (1983). They all generally conclude that storm waves provide the energy to scour the seafloor and transport or resuspend sediment. Swift et al. (1983) documented that a hummocky sea floor may be sculpted by a combined flow regime, with both a mean flow that transports sediment offshore and a wave-orbital component, which modifies the resulting bedform.

A unidirectional flow (mean flow) may be generated by (1) wind-forced currents or (2) relaxation currents (storm-surge ebb). Walker (1984) summarized the mechanisms by which these currents are generated. Wind-forced currents are generated as wind blowing across the water

entrains deeper and deeper layers in the ocean until the moving water column affects the ocean floor. The direction of wind-induced flow depends upon the orientation of the shoreline relative to the impinging winds; currents may be directed alongshore, onshore or offshore. Landward-directed winds would generate a coastal set-up (high water elevations) and a seaward pressure gradient. The resultant bottom flow would be directed seaward toward the zone of low pressure and would be simultaneously deflected by the Coriolis effect. The seaward-flowing parallel current would therefore evolve into a geostrophic current flowing parallel to the isobaths. Onshore directed storms will also generate a seaward-returning geostrophic flow (called storm-surge ebb) in a manner similar to currents generated by onshore-directed winds.

On the other hand, winds blowing offshore would create a coastal set-down and a landward pressure gradient. Bottom flow would be directed landward and also evolve into geostrophic flow (see Walker, 1984, Fig. 2).

If indeed a hummocky sequence is emplaced by a combined flow mechanism, the mean flow is probably a combination of wind-forced and storm-surge ebb currents. Modern oceanographic data indicate that geostrophic discharge from wind-induced currents may be 2000-3000 times as great as the storm-surge ebb discharge for a two day storm (Swift, pers. comm., fide. Walker, 1984). Swift (in press, fide. Walker, 1984) also emphasizes that the seaward-directed part of storm-surge ebb is minor compared with geostrophic flow parallel to the isobaths. Tool marks and parting lineations on the bases of hummocky sets of Facies Shcs show an

eastward-directed flow. Due to the poor understanding of the Cambrian regional paleogeography for the Chilhowee Group, it is not known whether the currents flowed seaward or parallel to the isobaths.

A sequence of events that could yield an ideal hummocky set was formulated by Swift et al. (1983) and was developed by comparing storm flow patterns and sea floor responses on the storm-dominated inner Atlantic shelf between Cape Cod and Cape Hatteras. They record the presence of hummocky megaripples with side-scan sonar in areas where tidal currents are less than 10 cm/sec and are inadequate to induce sediment transport. Hummocky megaripples are oval to circular in plan view and lack the asymmetry of current megaripples. Fields of short-crested to sinuous megaripples are developed seaward of estuaries, tidal inlets or on tidal shoals where tidal currents are strong.

Long-term current-meter records indicate that storm-induced currents in excess of 16 cm/sec occur about every 10 days and that major flows, capable of transporting large amounts of sediment occur three to five times a year, primarily in response to winter storms (Swift et al., 1983, p. 1299). Abundant hummocky megaripples have been observed as far landward as the foot of the shoreface.

Boxcores and vibracores of hummocky megaripples indicate that the sediment is fine-grained, moderately to well-sorted sand and internally consists of laminae dipping at low angles and of erosional surfaces that truncate the laminae at low angles. The size of the boxcore (a spade width) precluded retrieval of a definite hummocky set (i.e., no complete synforms or antiforms were recovered). However, the features seen are

most simply explained as laminations and truncations that formed on the side of a hummock. Swift et al. (1983) synthesized their observations into a process-response model that is similar to the ones formulated by Dott and Bourgeois (1982) and Walker (1979), who examined ancient hummocky-stratified sequences.

As a storm intensifies, the seafloor is eroded. The interaction between bottom irregularities and the velocity boundary layer (Richards, 1980, fide. Swift et al., 1983) generate regular patterns of scour-intensity gradients. As a result, a random distribution of scour pits is eroded into the seafloor. Sand settles out as the flow passes, and the seafloor aggrades as mound-like zones rise around the scour pits. As the storm wanes, more sand is carried into the area and the flow can no longer maintain the hummocky megaripple pattern. Turbulence is no longer more intense in the scour pits than on the crests, and deposition occurs preferentially in the pits. If the storm event is characterized by a downstream velocity decrease, the scour pits will become entirely filled with sediment.

After the storm has passed, the sea floor appears to be flat, exhibiting only fairweather oscillation ripples. If the shelf is characterized by a high suspended-sediment input, a layer of mud will accumulate over the sandy sea floor and fairweather burrowing fauna may colonize the mud and some or all of the storm sands.

Other features also point to periodic rapid deposition. These features have also been observed in ancient deposits of hummocky cross-stratification of various ages (e.g., Cretaceous Cardium Formation,

Harms et al., 1982; Cretaceous Fernie-Kootenay transition, Hamblin and Walker, 1979; Cretaceous Cape Sebastian Sandstone, Bourgeois, 1980; Lower Cambrian Andrews Mt. Member, Mount, 1980; Eocene Coaledo Formation, Dott, 1966; Pliocene-Pleistocene of the Makran coast, Harms et al., 1982; Cretaceous Shannon and Sussex Sandstones, Harms et al., 1982).

The deposits typically consist of alternating burrowed and unburrowed intervals. Periods of quiescence alternating with times of erosion and deposition in which the benthic community was disturbed by periodic storms would create such a pattern.

The flat even bases of many of the sandstone beds, together with parting lineations, thin parallel tool marks and wrinkle marks, record the initial erosion and molding of a cohesive substrate by storm currents. Wrinkle marks may also form in very shallow water when wind shear over the water surface wrinkles a slightly cohesive substrate (Reineck and Singh, 1980), and are indicative of nonerosive shear rather than of shallow water-depth. Uncommon granule lags may be present and are a result of scouring currents which remove the finer-grained fraction of the sediment.

Soft sediment deformation features, such as ball-and-pillow structures, are formed by rapid sedimentation of sand on a clay or silt layer. A disturbance of some kind may cause the more dense sand layer to break up into semi-isolated sand bodies, which sink into the underlying yielding mud layer (Reineck and Singh, 1980). Ball-and-pillow

structures are more indicative of rapid deposition rather than a particular environment, for they occur in both shallow-water and deep-water environments.

Several explanations have been proposed for the genesis of convolute laminae; such hypotheses include liquefaction (due to overloading, seismic waves or some other shock), water expulsion, and shearing action of a strong current over a sediment. However, liquefaction of a sediment is the most important factor in the genesis of convolute bedding (Reineck and Singh, 1980). Both ball-and-pillow structures and convolute laminae are evidence for an unstable substrate which may periodically liquefy.

Modern shelf sediment is known to become quick and unstable in response to pressure pulses associated with storm waves (Saxov and Nieuwenhuis, 1982). Synsedimentary deformational features indicate not only rapid deposition but also point to a shelf setting. The fact that the ball-and-pillow structures have been eroded attests to the intense scouring currents present on the shelf.

Variations of the Ideal Hummocky Sequence

The ideal sequence (Fig. 30, p. 51) is rarely completely preserved in Facies Shcs. A review of ancient units characterized by hummocky cross-stratification by Dott and Bourgeois (1982) documented that variations of the ideal sequence occur in nearly all ancient hummocky units, and that the complete hummocky sequence is not the dominant feature.

Close examination of Shcs reveals that the variants occur vertically in a coarsening-upward sequence (Fig. 30C). These patterned

perturbations of the ideal sequence may reflect: (1) fluctuations in relative sand supply; (2) relative water depth; (3) frequency, duration and magnitude of storms; and (4) proximity of strong tidal currents (i.e., magnitude of tidal range). The depositional conditions that generated each end member are discussed below. The two end-member sequences are similar to the micro-hummocky lenses type-sequence and the amalgamated type-sequence of Dott and Bourgeois (1982)

Fig. 30A illustrates the most thinly bedded and the finest-grained end-member. The 0.5-5 cm beds/lenses of sandstone within the bioturbated siltstone represent deposition from pulses of currents generated by distal storms. The occasional presence of symmetrical ripple forms with unidirectional ripple cross-laminations within the form indicates current deposition of sand and oscillatory reworking of the storm layers. Most commonly though, these storm sand layers are structureless as a result of recolonization of the substrate by deposit feeders during fairweather periods. The environmental conditions necessary to deposit this end-member are a deeper-water setting with a more distant sand source relative to the settings for the other sequences. The sand layers therefore reflect the most distal reaches of a storm-dominated shelf.

The other end-member variant (Fig. 30B) consists of planar-laminated, fine-grained sandstone. In contrast to the thinly interbedded sandstone and siltstone, it reflects erosion and deposition in areas located proximally to storm-generated currents. In this case, 50-70 cm beds of planar-laminated sandstone represent an amalgamation of storm

events. Storm and wave currents may generate a complete sequence, but a subsequent event erodes everything except the basal planar-laminated zone. All evidence of fairweather mud deposition and of a hummocky substrate (i.e., antiforms and swales) has been removed. In this manner a series of amalgamated storm events can be generated.

The coarsening-upward sequence may range in thickness from 1-4 m (Fig. 30C). It results from sandstone becoming increasingly dominant upward as offshore mud with intercalated storm sand (Fig. 30A) is overlain by a zone of well-developed hummocky cross-stratification (Fig 27, p. 49), which is gradually overlain by a thick unit of amalgamated, planar-laminated sandstone (Fig. 30B). This sequence reflects the interaction of (1) the proximity and migration of a sand source and more importantly, (2) the proximity, magnitude, duration and frequency of storms.

Source area migration may have been accomplished by either shoreline progradation or by migration of a local sand source in response to storm-surge ebb currents or tidal currents. Levell (1980) suggested that on a sand-starved shelf, sand patches may serve as a local sand source. The abundance of glauconite indicates that the shelf was receiving little sediment influx (Odin and Matter, 1981). Sand may have been effectively trapped in estuaries that formed as a result of the Cambrian transgression. The sand source may also have shifted as the pathways of storm-surge ebb currents and tidal currents changed.

The effects of storm frequency, duration, magnitude and proximity are more difficult to evaluate because they are a function of the

climate and regional and global storm patterns. Whether or not an increased frequency of storm events, for example, is recorded in the sediments is also a function of how much sediment is available. The sedimentary record of storm deposits reflects the interaction of storms and sediment source. The sand source for Shcs is interpreted to be sand patches and sand ridges rather than sediment input from rivers. Sand may also have been eroded from the shoreface and transported offshore. The difference between the two possible interpretations lies in the location of the paleoshoreline, which in this case is unknown. In the first setting, storm sands are probably deposited down current from the ridges and patches. This is similar to the gross sediment distribution patterns in the North Sea (Reading, 1978, p. 219) and in Bristol Bay of the southern Bering Sea (Sharma et al., 1972).

Because Shcs is so intimately associated with Sms and QA-Erwin, the sand source for Shcs must be compatible with the interpretation for the sequences: Shcs → QA-Erwin or Shcs → Sms → QA-Erwin. A discussion of the sequences is presented in the following interpretation of Facies QA.

Facies QA

General Marine Environment

Facies QA is characterized by extreme mineralogical and textural maturity (Fig. 33, p. 57); the extensive reworking that would be required to round, sort and remove fine-grained sediment (silt and clay) and labile constituents (feldspar and rock fragments) most likely

resulted from either nearshore marine processes or long-term transport around shelf sand ridges (Houbolt, 1968).

Subfacies QA-Unicoi: Beach Foreshore-Nearshore Zone

QA-Unicoi, while interpreted to be marine because of its maturity, is intimately associated with Facies G (Fig. 37, p. 62) in the lower part of the Unicoi Formation. Such a setting in which quartz-rich sediment can be intimately associated with immature braided alluvial sediment is in the foreshore and nearshore zone of a wave-dominated coastline.

The sedimentary structures of QA-Unicoi also indicate a nearshore setting (Figs. 34 and 35, pp. 59, 61). The sets of low-angle cross-stratification that are defined by heavy mineral laminae and by graded laminae resulted from wave swash and backwash in the foreshore (Clifton, 1969; Reineck and Singh, 1980; Reinson, 1984). The nearshore zone (area landward of the breakers) and the foreshore are closely associated with each other because both are affected by the complex hydraulic environment of the surf zone. Currents within these zones include shore-normal currents generated by plunging waves and shore-parallel wave-driven currents. The combination of currents results in a set of multidirectional sedimentary structures (Reinson, 1984).

The foreshore and nearshore are also subject to intense erosion by storm waves. Storm activity is reflected in the beach-nearshore profile because, as storm waves erode the beachface, sediment is transported not only offshore, but is also piled into nearshore bars (Davis et al.,

1972; Owens and Frobel, 1977). This storm-generated bar and adjacent trough, called a ridge-and-runnel when located in the intertidal zone, does not remain stationary during fairweather periods. Rather, sediment is gradually returned to the beach during post-storm-recovery as small waves create currents which transport sediment over the bar crest in a landward direction. The ridge thus behaves like a migrating bedform (Davis, 1983).

A shore-parallel runnel and a shore-oblique rip channel form to accommodate the seaward return of water. The runnel is thus floored with ripples (ladderback ripples), which result from processes operating within the runnel at right angles to one another (Davis et al., 1972). Waves spilling over the ridge transport sediment landward in the form of unidirectional ripples, whose crests are oriented parallel to the shoreface. Water returning to the sea generates a set of ripples oriented perpendicular to the shoreline. Mud flasers may be present in the ripple troughs. As the ridge migrates over the runnel, it eventually welds to the beachface and reworking of the welded ridge creates seaward-dipping foreshore deposits.

The structures found in QA-Unicoi indicate not only fairweather swash-backwash processes, which formed seaward-dipping low-angle cross-stratification, but also storm processes. Broad, shallow scours and gravel lags indicate initial beachface erosion. Large-scale, planar-tabular cross-stratification with azimuths oriented from 50-211° may have resulted from landward or obliquely landward migration of storm ridges or longshore bars (Davis et al., 1972). The thin, discontinuous

lenses of laminated silty conglomerate overlying a surface with interference ripple marks is all that remains of a runnel or longshore trough deposit.

Paleocurrents of this facies are not well documented, due mostly to poor three-dimensional exposure. However, the low-angle cross-stratification dips east-southeast, in a seaward direction (see interpretation of Sms for a discussion of the paleoshoreline). The azimuth range of 50-211° for the large-scale cross-stratification and its significance in view of the uncertain shoreline position, is debatable. Therefore, whether or not any of the sets dip directly landward is at this point an unanswerable question. What the paleocurrent pattern does indicate is the complexity of the currents within the nearshore zone due to the interaction of waves, longshore currents and tides with each other and with the nearshore topography.

A modern example of such complex current systems is the nearshore zone of the barriers of Kouchibouguac Bay, New Brunswick (Davidson-Arnott and Greenwood, 1976). There, a ridge-and-runnel is simply a nearshore bar and trough that accretes to the beachface. Davidson-Arnott and Greenwood (1976) discussed the sedimentary structures that form within a submergent nearshore bar system. Bars are emergent for short periods only at spring low tides. The five subenvironments and their distinctive structures are: (1) seaward slope--dominated by shoaling waves, ripples formed parallel to advancing wave crest, and seaward dipping plane bed; (2) bar crest--dominated by shoaling and breaking waves, characterized by plane beds and lunate megaripples with seaward-

and landward-dipping slip faces; (3) landward slope--steep avalanche faces that dip landward and result from sediment being transported over the bar crest during periods of high wave activity; (4) trough--dominated by longshore currents which generate current ripples perpendicular to the shoreline (fine organics may accumulate here); (5) rip channel--unidirectional rip current forms seaward-dipping megaripples. Thus, a wide spectrum of structures with a variety of orientations can form within the nearshore zone.

A general comment that can be made in reference to modern nearshore zones is that high wave-energy areas are dominated by physical sedimentary structures, whereas low to intermediate wave-energy areas allow an alternation of physical and biogenic structures (Reading, 1978) as well as alternating sandy and muddy lithologies (Howard and Reineck, 1972). The absence of finer-grained units as well as the maturity of QA-Unicoi dictates a high-energy nearshore environment. The mechanism for reworking of sand is the interplay of storm and fairweather deposition. Storms remove beach sediment into the nearshore zone and fairweather waves transport sediment landward (Davidson-Arnott and Greenwood, 1976) and in some cases, back to the beach. It is the landward-seaward transportation of sediment grains that would effectively sort and round them.

The association of Subfacies QA-Unicoi with Facies G also supports the nearshore marine interpretation of QA-Unicoi. Facies G is interpreted to be a braided stream deposit, and there are two modern settings in which braided stream sediment is immediately reworked by nearshore wave processes: (1) the Skeidarasandur shoreline (a glacial outwash plain

shoreline) of the southern coast of Iceland (Hine and Boothroyd, 1978); and (2) Yallahs fan delta of Jamaica (Wescott and Ethridge, 1979). The tidal range of the Icelandic coast is 1.2-2.0 m and Jamaican coast is also less than 2 m. In the transition zone between totally fluvial and totally marine environments in both settings, waves impinge directly on fluvial deposits. The features that develop on both shorelines reflect the dominance of waves and longshore processes and the relative unimportance of tides.

The Skeidarasandur coastline is characterized by barrier spits with well-developed berms, nearshore ridge-and-runnel systems, and large subtidal bars. Nummedal et al. (1974) did not examine the internal structures produced within each environment, but rather documented the overall geometries of the fluvial and marine system and the associated wave regime. Hine and Boothroyd (1978) examined the sedimentary structures produced on the barrier spits. They found that barriers were dominated by overwash and swash processes; hence the dominant sedimentary structures are landward-dipping, gently inclined laminae (berm-top sediments) and seaward-dipping laminae (beach sediments).

The processes associated with the nearshore ridge-and-runnel system of Iceland are not unlike those described from Lake Michigan and the Massachusetts coast (Davis et al., 1972). In addition, the wave action over the subtidal bars is similar to the processes documented for the barred system of Kouchibouguac Bay. The most striking difference between the Icelandic coast and these other modern examples is the effect of high sediment influx and rapid river discharge fluctuations on the nearshore system.

Barrier-spit sediment is overlain by tidal-inlet fill (Kumar and Sanders, 1974), lagoonal, or nearshore sediment. The Skeidarasandur shoreline, however,has prograded about 1 km over the past 300 years due to the high sediment influx. Much of the sediment supplied to the nearshore during glacial burst flooding is redistributed by longshore currents on the shoreface. There is both offshore-onshore sediment transport as well as longshore sediment transport (Nummedal et al., 1974).

The transition zone of the Yallahs fan delta of Jamaica is dominated by beaches, both depositional and erosional. The depositional beaches consist of well-developed berms and gently sloping beaches. The presence of well-developed berms, like those on the Icelandic barrier spits, indicates the importance of washover depositional processes resulting from high storm waves. The Yallahs fan delta has no well-developed nearshore or offshore bar system because the alluvial fan progrades directly onto a steep submarine slope, whereas the Icelandic shoreline is prograding onto a continental shelf. Regardless of the two different tectonic settings, the Yallahs fan delta serves as another example in which fluvial deposits are directly modified by high-energy wave regime in a nearshore marine environment.

Subfacies QA-Erwin: Subtidal Sand Ridges

A marine origin for the sediment of QA-Erwin is indicated not only by their maturity but also by their association with Sms and Shcs; Sms and Shcs are both interpreted to be marine based upon the presence of glauconite, marine trace fossils and diagnostic sedimentary structures. QA-Erwin is interpreted to be subtidal because hummocky beds commonly

are found directly above or below the large sand bodies. There is also no evidence for emergent conditions.

The interpretation of QA-Erwin as subtidal sand ridges is based upon sedimentary structures and facies associations. Given the massive appearance of the sandstone bodies, interpretations of QA-Erwin rely heavily on the vertical sequence. Therefore, the first part of the discussion is concerned with the sedimentary features found in modern, subtidal sand bodies and how they compare with QA-Erwin. The second part discusses the significance of abundant glauconite above and below the sandstone bodies and summarizes the depositional setting for QA-Erwin.

Interpretation of sedimentary structures. Large subtidal sand bodies have been termed sand waves by Allen (1980) (see Appendix B), who used the term in a genetic sense. He defined a sand wave to be a flow-transverse bedform associated with reversing tidal currents. Wavelengths vary from 25-1000 m, with heights ranging from 1-25 m. Generally, the wavelength of sand waves increases with water depth, from 25 m in the intertidal and shallow subtidal zones to 1000 m in 100 m of water. The slope of the stoss side is 1-5° and the lee face dips from 6-30°; the larger the bedform, the smaller the angles of dip. Sand waves are found in a variety of tide-dominated environments. They occur in channels of coastal barriers (Kumar and Sanders, 1974) and within estuaries (Visher and Howard, 1974). They are most abundant in restricted seas and seaways where tidal ranges are high (meso- or macrotidal)

and tidal currents are strong (Cook Inlet, Alaska, Bouma et al., 1977; the Bay of Fundy, Klein, 1970; Dalrymple et al., 1978; the North Sea Houbolt, 1968; Caston, 1972; Caston and Stride, 1973).

Allen (1980) developed a model for the formation of sand waves based upon an interpretation of the associated currents. The external and internal structures of sand waves depends upon the strength and time-velocity asymmetry of the associated currents. Symmetrical sand waves, associated with currents of low asymmetry, are unaffected by large-scale flow separation. Internally, these forms are expected to consist of compound cross-stratification with gently inclined master bedding surfaces and thin sets of herringbone cross-stratification; well-developed avalanche bedding is relatively rare.

Asymmetrical currents produce asymmetrical sand waves with one side so steep that a flow separation is produced. This group of sand waves is expected to contain internal: (1) angle-of-repose avalanche bedding separated by reactivation surfaces, representing tidal reversals, or (2) mud drapes or bioturbated zones, representing periods of gentle flow. The predicted internal structures are similar to those in modern sand waves and in ancient sand bodies interpreted to be sand waves.

A spectrum of external forms and internal structures exists between these two end-members, but regardless of where a particular sand wave falls within the spectrum, it is characteristically internally complex.

The importance of this discussion of sand waves is that these large-scale bedforms are defined to be produced by periodic currents (i.e., tides) and in the stratigraphic record are recognized not only by

their lateral geometries, but also by smaller internal structures generated by the periodic (or unsteady) nature of the currents. Mowbray and Visser (1984) discussed similar features found in a subtidal channel deposit of the Oosterschelde sea-arm of the North Sea. The two diagnostic tidal features are clay drapes and reactivation surfaces. Clay drapes are deposited during slack-water periods on the foresets of bedforms that developed in response to the dominant tidal phase. Using clay drapes as an indicator of periodic tidal flow creates problems, because there are some tide-dominated settings that never experience slack water. For example, tidal flow velocity within a channel may approach zero at slack tide. However, on a more open shelf, where tidal current flow is in an elliptical pattern rather than a rectilinear reversing flow, velocities may remain above those needed to allow fine-grained sediment to settle.

In the absence of fine-grained drapes, reactivation surfaces record the unsteadiness of flow conditions. Reactivation surfaces are defined to be gently inclined erosion surfaces that separate adjacent foresets of similar orientations (McCabe and Jones, 1977). They are thought to reflect fluctuations within the flow mechanism or changes in flow direction; they are not restricted to a particular environment (Reineck and Singh, 1980). Reactivation surfaces can be produced by a faster-moving megaripple migrating over a slower-moving one or by erosion of the lee side of a megaripple by a subordinate current (Mowbray and Visser, 1984). They are common in tidal deposits (Klein, 1970), but examples are described from eolian and fluvial deposits as well.

Subfacies QA-Erwin does not contain clay drapes, and no fine-grained or silty interbeds are associated with the large-scale cross-sets. No reactivation surfaces are visible in outcrop, and internally QA-Erwin does not appear to consist of smaller-scale cross-sets. Large-scale planar-tabular cross-stratification is present; but does not appear to be well-developed. Because of negative evidence pointing to an absence of periodic tidal flow, another origin for the massive-appearing sandstone bodies must be considered.

Swift and Field (1981) discussed the morphology and formation of sand ridges on the Delaware-Maryland inner shelf. The sand ridges are large shoreface and shelf features ranging from 1-12 m in height, from 1-3 km in width, and from 2-18 km in length. Their spacing ranges from 1-12 km. The spatial variations of the ridges on the Atlantic shelf is indicative of how the ridges respond to increasing water depths. Swift and Field (1981) developed a projected sequence of sand ridge development through time and in conjunction with the Holocene transgression.

The sand ridges originate as shoreface attached ridges (i.e., as wave-built bars due to beachface erosion in the nearshore zone). Unlike some wave-built nearshore bars, these ridges respond to both a mean storm-generated unidirectional flow as well as wave-orbital motion (oscillatory flow). During the course of transgression, as the water column deepened, the shoreface-attached ridges are not affected as frequently by wave-orbital motion. The relative dominance of the mean flow over wave processes produces megaripples and sand waves between the ridges. Because wave agitation is no longer constant, mud accumulates

in the troughs between ridges. In even deeper water, the ridges become more asymmetric in a seaward direction. The coarsest grains are concentrated on the landward-facing flank as a result of erosion, whereas finer-grained sediment is concentrated an the seaward-facing flank indicating deposition. Grain size gradients occur on all the ridges at all water depths and at all distances from shoreline. Therefore, the ridges 20 km from shore are not relict sediments but are responding to flows.

Werner and Newton (1975) record 2 m high sand waves in the Baltic Sea that have formed in response to wind-driven aperiodic currents. The bedform contains form-concordant internal cross-stratification consisting of single foresets dipping in one direction. They interpret the lee face to be an active avalanche face. In contrast, most tidally produced bedforms have form-discordant internal stratification (Allen, 1980; Mowbray and Visser, 1984). Only the uppermost 20 cm of the Baltic Sea sand waves consists of small ripple bedding and bioturbation produced during inactive periods.

The internal structure of modern storm-generated ridges has not been well-documented. However, high-resolution seismic reflection profiles across the ridges of the Maryland inner shelf appear to contain some seaward-dipping surfaces which indicate growth by aggradation (Swift and Field, 1981). Stubblefield and Swift (1976) also documented internal stratification in a sand ridge on the New Jersey shelf. In both cases, the exact nature of the internal stratification is unknown.

Detached shoreface ridges, as described by Swift and Field (1981) are linear in plan view and are oriented obliquely to the shoreline.

Sand waves, as discussed by Allen (1980) are oriented transverse to the dominant flow direction and their orientation may therefore be independent of the associated shoreline's orientation. In either case, knowing the lateral geometry of the sand body and its orientation relative to the paleoshoreline is essential to accurately interpret QA-Erwin. Unfortunately, lateral variability of QA-Erwin remains to be documented. Due to poor preservation of sedimentary structures and two possible interpretations, the interpretation of QA-Erwin must rely heavily upon its position in the vertical sequence (Fig. 34, p. 59).

Interpretation of facies sequence. QA-Erwin forms the top of two coarsening-upward sequences 55 m and 35 m thick (Fig. 32, p. 55). A single vertical sequence consists of, in ascending order: Shcs → Sms → QA-Erwin or Shcs → QA-Erwin. As discussed in previous sections, Shcs is thought to represent storm deposition of sand in the offshore zone between fairweather and storm wave base. Sms represents megarippled subtidal sand shoals formed by a combination of storm flows and tidal currents. However, no definitive evidence for a dominantly tidal origin for Sms was observed. The interbedding of <1 m thick units of Sms with Shcs indicates that Sms may have been formed by storm-generated currents. Swift et al., (1979) documented that up to 1 m high megaripples can migrate in response to storm flow, thereby producing angle-of-repose cross-bedding.

The coarsening-upward sequence as a whole represents migration of subtidal sand ridges over adjacent megaripple patches and storm-sand/

fairweather-silt patches. Erosion associated with active migration of large sand bodies is shown by the broad scoured bases. These sand bodies do not represent relict sediment associated with a transgression, but rather they are bedforms actively responding to dominantly storm-induced currents that may be enhanced by tidal currents.

Field and Roy (1984) described large inner-shelf sand bodies, up to 30 m thick, several kilometers wide and tens of kilometers long. The bodies are prograding into deeper water (50-60 m) and they believe that the sediment is transported from the shoreface to the inner shelf by storm-generated currents. The East Australian Current modifies the shape and texture of the sand bodies. Besides high-energy conditions to erode the shoreface, the other major factor needed to generate large piles of sand is long period of stable sea level during which large amounts of shoreface sediment are transported offshore.

The extremely mature mineralogy of QA-Erwin therefore probably resulted from sediment reworking first on the shoreface and then continued reworking by storm currents during renewed transgression.

Interpretation of glauconite. The upper surface of QA-Erwin contains a thin granule/pebble lag, which indicates strong winnowing by currents. The sand body is overlain by 30 cm of shale with lenses of conglomerate. This may represent a period of rapid transgression during which the shelf was swept only rarely by large, long period waves that segregated coarse-grained sediment into lenses within the shale. Komar et al. (1976) documented that such waves can ripple sediment to water

depths of 200 m on the Oregon continental shelf. During rapid transgression, deposition was therefore dominated by sedimentation from suspension.

The lenses of conglomerate are glauconitic. Glauconite is not a useful environmental indicator because it is widespread on continental shelves between 50° S to 60°N latitude and in water depths between 50-500 m (Odin and Matter, 1981). Its abundance does indicate slow sedimentation rates that favor a suitable chemical microenvironment within the substrate.

Glauconite forms by either: (1) adding potassium and iron to an iron-rich mica, such as biotite (Burst, 1958a, 1958b; Hower, 1961); or (2) authigenic crystal growth within the pores of a substrate and simultaneous alteration and replacement of the substrate (Odin and Matter, 1981). The substrate may consist of allochemical carbonate particles, argillaceous fecal pellets, mineral grains or rock fragments.

According to Odin and Matter (1981), the chemical condition that is optimum for the formation of glauconite is semiconfinement within a substrate. This provides a slightly reducing microenvironment in which potassium and iron are removed from seawater. The exchange of ions between substrate pores and ambient seawater is facilitated by nondeposition or slow sedimentation. In an environment receiving little sediment, periodic bottom currents would bring in fresh supplies of ions as well as in segregating transgressive lags into lenses. Marine transgressions seem to provide favorable conditions for widespread glauconite formation because coarse sediment deposited on the shoreface is

submerged to greater depths where water agitation and sedimentation rates are greatly lessened (Odin and Matter, 1981). The presence of a porous substrate and slow sedimentation seem to be the prerequisites for glauconitization.

In summary, QA-Erwin is interpreted as subtidal sand ridges on a storm-wave dominated shelf (offshore zone) where tidal currents were of secondary importance. The origin of the sandstone bodies is hypothesized to be similar to the origin of the shoreface ridges of the storm-dominated Atlantic Shelf (Swift and Field, 1981). This is by no means a definitive interpretation, because it is based primarily upon the lack of tidally formed structures. It has been noted that the similarities between tide- and storm-dominated sand ridges would make it difficult to distinguish one from the other in the geologic record (Reading, 1978).

It is particularly noteworthy that many of the shelves affected by winter storms, including the eastern coast of the United States, the Bering Sea and the North Sea are characterized by linear sand ridges (Marsaglia and Klein, 1983). These ridges are thought to have been formed by ebb-flood channel systems, spiral flow systems, and rotary tidal current systems (Swift et al., 1979). On these shelves, intense storms tend to amplify tidal currents.

IV. DEPOSITIONAL MODEL FOR THE CHILHOWEE GROUP

The Chilhowee Group represents an overall transgressive sequence. Transgression is most strikingly evident in the upward textural and compositional maturing trend of the rocks. This observation, in conjunction with the appearance of marine trace fossils, indicates that marine processes became dominant over fluvial processes. Worldwide, the Early Cambrian has been documented to be a time of sea-level rise, due perhaps to simultaneous rifting on several continental landmasses (Ziegler et al., 1981).

The mineralogy of the lowermost units of the Chilhowee Group (Facies G) at this locality, reflects the nature of a proximal sourceland. The sourceland most likely consisted of sedimentary, metamorphic and plutonic rocks, and crystalline rocks must have been rich in feldspar. Rapid erosion of the sourceland within a warm humid climate accounted for the abundant immature detritus that accumulated on a braided coastal plain.

The depositional model for Early Cambrian sedimentation spans a spectrum of processes and environments. Deposition commenced with braided alluvial deposition on a coastline. Terrestrial sediment intertongued with marine sediment as waves eroded and redeposited coastal plain sediment into nearshore bar systems. With time, and perhaps concomittant with widening of the continental margin, tidal processes became important (Klein, 1982) and nearshore sedimentation processes were dominated by tidal channel migration. When transgression reached the point of effectively trapping sediment along the coastline,

tidally enhanced storm currents reworked and redistributed the sediment of a sand-starved passive continental margin into sand patches and ridges.

The most effective way to understand this depositional model is to consider a time sequence diagram of lateral facies relationships for three relative time intervals (see Fig. 37, p. 62). Fig. 41 is a schematic representation in which each diagram illustrates the facies relationships for a time interval characterized by a particular facies. Table 3 summarizes the characteristics of each facies and subfacies.

Time 1 (Fig. 41A) is dominated by Facies G and therefore by alluvial sedimentation. Random interfingering of Facies G with Subfacies QA-Unicoi records the interaction of sediment supply and alluvial processes with sea level position and marine processes.

An analogous situation may be found on the mesotidal Skeidarasandur shoreline of Iceland (Hine and Boothroyd, 1978). Here the interface of the braided alluvial plain and the marine environment is characterized by narrow barrier spits, which are either backed by active streams or by wind-tidal flats and inactive sandur deposits. The unique characteristic of this storm-wave dominated, high fluvial-sediment-discharge coast is that the many coastal features associated with barrier islands (e.g., recurved spits, tidal deltas, lagoons and estuaries) are absent (Hine and Boothroyd, 1978). Sections of the coastline that are frequently flooded and those sections backed by active stream channels have narrower spits than those sections that have not been flooded for an extended

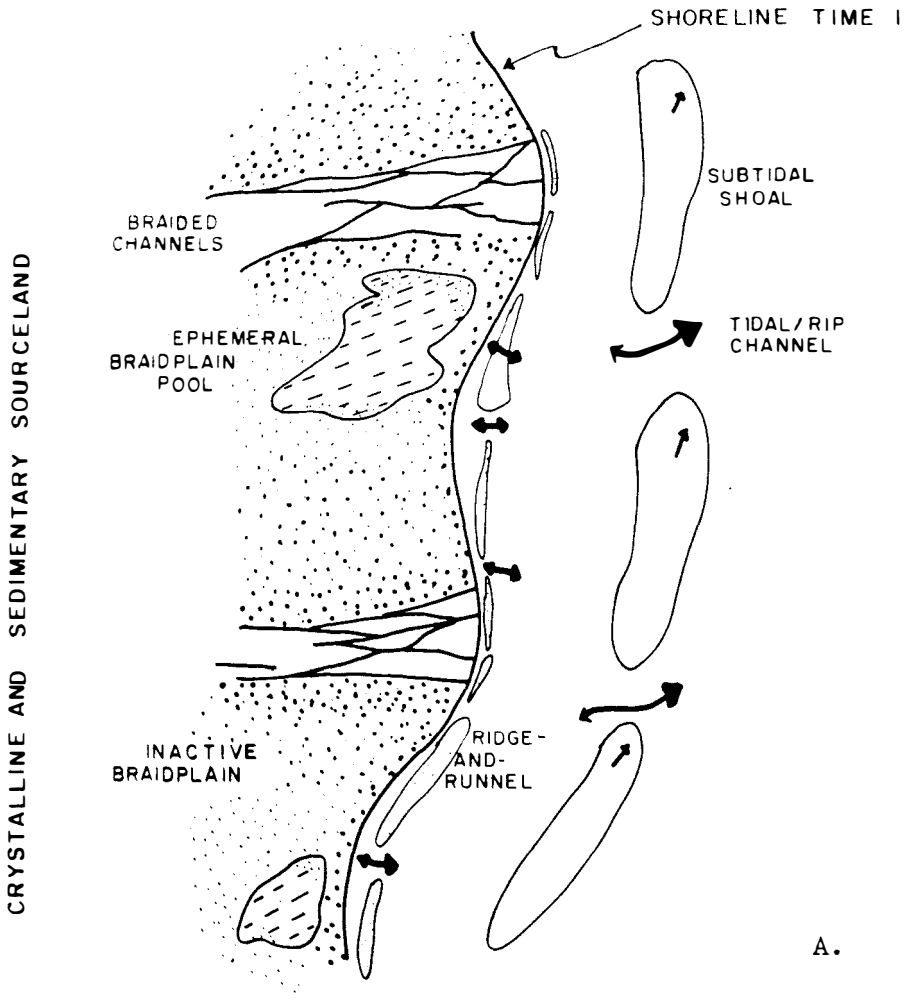
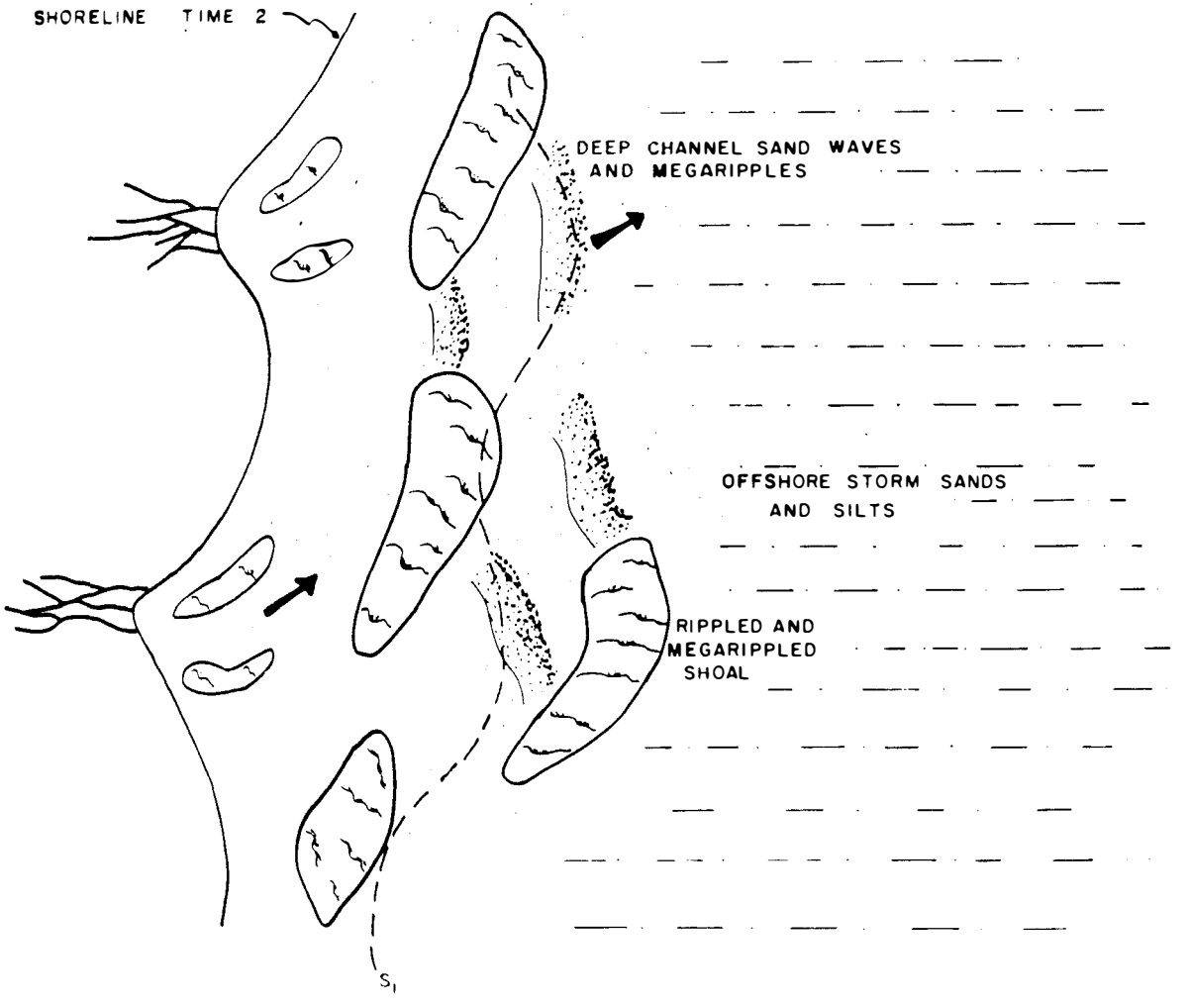
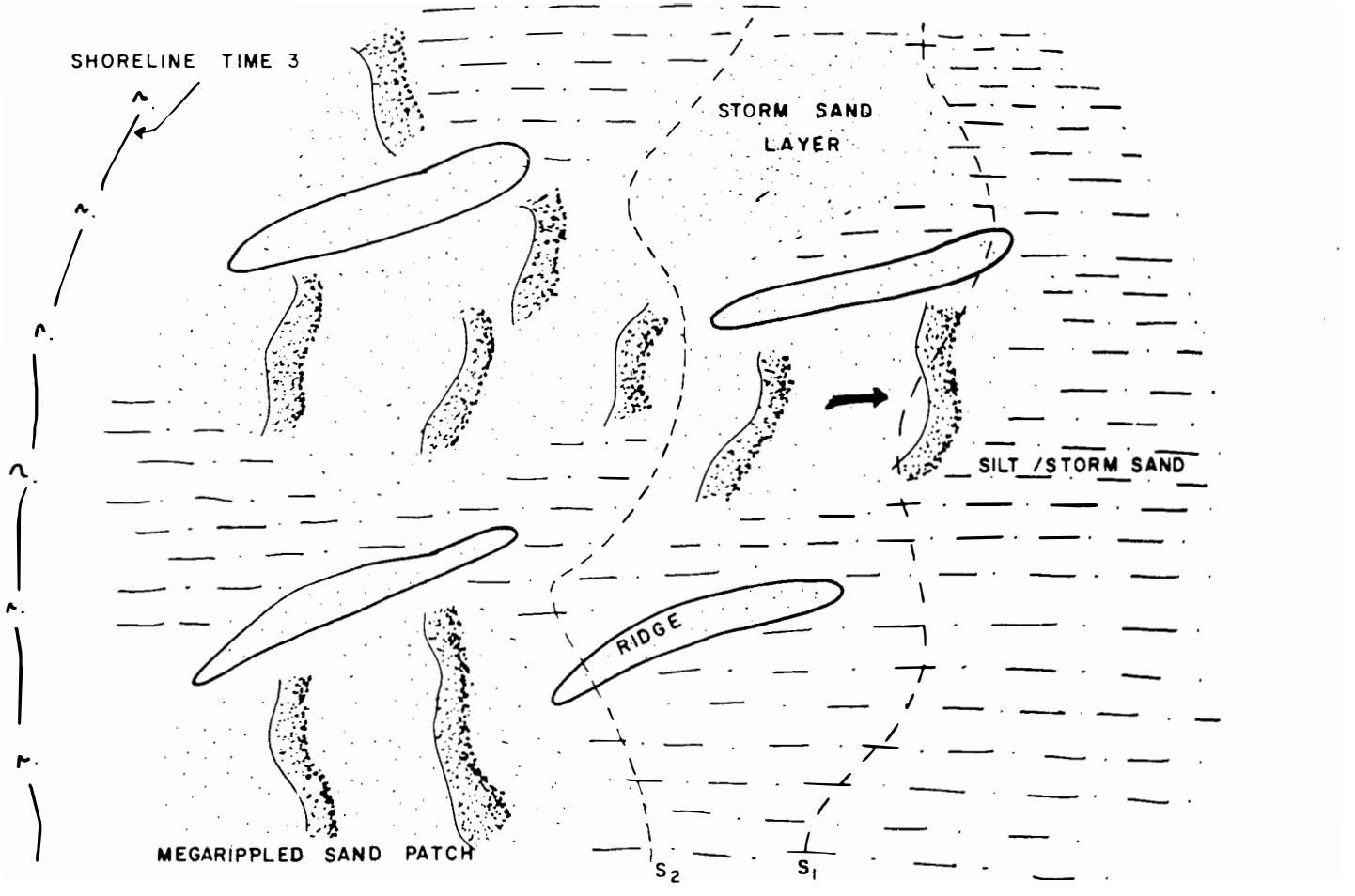


Figure 41. Schematic depositional model for the Chilhowee Group at three relative times. (A) Time 1 is dominated by braided stream deposition. (B) Time 2 is dominated by shoreface tidal channel-shoal migration. (C) Time 3 is characterized by storm/tidal deposition on a sand-starved shelf. The orientation of the shoals and channels in (B) and the ridges in (C) has not been documented in the field and is therefore hypothetical.



B.

Figure 41 (continued)



C.

Figure 41 (continued)

TABLE 3. Summary of facies and subfacies and their interpretations.

FACIES	SUBFACIES	BRIEF DESCRIPTION	PROCESS	ENVIRONMENTAL INTERPRETATION
C		v.f.-pebbly, arkosic lithic wacke	no reworking	braided stream system
	Cp	large-scale, planar-tabular cross-strat.	lower flow regime, high flood stage	transverse channel bars
	Ch	horizontal laminae, granule laminae, small-scale cross-strat.	upper flow regime, local variations in flow conditions	vertical accretion on flood plain
	Cmr	cross-strat. megaripples with siltstone draping the preserved bedford	megaripple migration during flooding, suspension deposition during low discharge periods	interbar pool deposits
S		m.-granules, subarkosic/ark. arenite	moderate reworking	subtidal channels and shoals
	Sfl	interbedded bioturbated sandstone and ripple x-laminated sandstone	suspension deposition, weak currents ripple sand	slack water between tidal phases
	Sls	large-scale, planar-tabular and trough cross-strat., granule lags	erosion and subsequent deposition by ebb-tidal currents	deep channel
	Sms	medium-scale, planar-tabular and trough cross-strat., rare symmetrical ripples	deposition by ebb-tidal currents, above wave base	shallow channel or shoal
Shcs		v.f.-f., subarkosic/ark. arenite interbedded hummocky sandstone and bioturbated siltstone, locally glauconitic, symmetrical ripple forms w/ undirectional x-laminae, Cruziana ichnofacies	deposition between storm and fairweatber wave base; sands deposited by storms, fairweather suspension deposition and recolonization of substrate by enfauna	storm shelf-deposition between sand ridges or at the end of a storm/tidal current transport path
QA		m.-c., supermature quartz arenite	reworking important	—
	QA-Unicoi	low-angle cross-strat., large-scale planar-tabular cross-strat., broad shallow scours, symmetrical ripples	intense wave reworking of fluvial sediment, swash and backwash important	ridge-and-runnel
	QA-Erwin	large-scale planar-tabular cross-strat. erosional lower surface, granule lag at top, sandstone bodies 1-9 m thick, NO FINES	deposition and reworking by tidally enhanced storm currents	subtidal sand ridge

period of time. Flooding prevents the development of wide barrier spits.

Seaward of the barrier spits is a system of ridges and offshore bars (Nummedal et al., 1974). The nearshore ridges become welded to the beachface by constructional waves and sediment is transported to the offshore bar during erosional phases. This onshore-offshore transport of sediment has a lateral motion superimposed on it due to longshore currents.The nearshore and offshore bars are the seafloor expression of the 100 m thick depositional wedge that widens in the direction of longshore currents. This widening reflects the importance of wave processes.

Progradation along the Skeidarasandur shoreline has proceeded very slowly due to the wave energy that has been expended at the shoreline. In some areas of the shoreline, a net recession may have occurred (Nummedal et al., 1974). Thus, in the case of Facies G and Subfacies QA-Unicoi, fluvial sediment may have intertongued with marine sediment as the shoreline moved seaward or landward in response to sediment supply and wave regime.

During Time 1, a system of shoals or bars and channels was located seaward of the shoreline because Facies S occurs in association with Facies G and Subfacies QA-Unicoi. Facies S, however, dominated Time 2. The transition between the two intervals was such that wave processes no longer dominated and tidal processes became more important.

During Time 2 (Fig. 41B), sedimentation in migrating subtidal channels by tidal or rip currents was the dominant process. Because

paleocurrents were directed both longshore (NE) and offshore (SE), the subtidal channels and shoals probably comprised a mixed wave- and tide-dominated shoreline.

The exact geometry and lateral orientation of the channels and shoals of Time 2 are unknown. In comparison, the shoreface bars of the coast of Iceland are long (13 km) and parallel to the shoreline. The shoreface bars and channels of the German Bight of the North Sea are oriented parallel to tidal currents rather than having a distinct orientation relative to a shoreline (Reineck and Singh, 1980). Because the shoals of Facies S are interpreted to have been influenced by both waves and tides, they were probably oriented obliquely to the shoreline.

Channel-shoal couplets (i.e., ridge-and-swale topography) are common on the Middle Atlantic Bight (Swift and Field, 1981) and are oriented obliquely to the shoreline. It has been hypothesized that tidal- and storm-generated linear sand ridges originated as channel-shoal couplets which became detached from the shoreline during transgression (Swift, 1975). This will be an important consideration in the interpretation of Time 3.

During Time 2, Facies S gradually fined from very coarse-grained to predominantly medium- to coarse-grained and lost its labile components; this indicates increased marine reworking of sediment deposited on the shoreface and shoreline. Rare occurrences of Facies G early in Time 2 as well as the initial coarse-grained nature of Facies S, however, indicate that sediment influx was still high high during Time 2, although the shoreline had retreated.

During Time 2 sediment accumulated on the shoreface in submerged shoals or bars. Rocks representing Time 2 in the upper part of the Unicoi Formation therefore consist of stacked subtidal channel-fill sequences. The stacked nature of the fill sequences probably reflects the abundant sediment supply and rapid migration of channels in response to the sediment supply and current regime.

Mention should be made at this point that Time 2 spans the large covered interval of the Hampton Formation. The question should therefore arise as to the significance of those covered intervals and how they fit into the proposed depositional model. The nearly covered units at this locality consist of alternating (1) interbedded siltstone and fine-grained sandstone and (2) thick sandstone ledges. The ledges can be seen on the hillsides such that (1) and (2) occur in fining- or coarsening-upward sequences. However, the exact nature of the contacts is unknown.

Schwab (1971) examined the Harpers Formation (Hampton equivalent in Virginia) and documented a vertical repetition of fine-grained intervals and thick sandstones. The fine-grained intervals (30-90 m) are interpreted as alternating turbidite sandstone and shale accumulating in a distal deltaic to deep shelf setting. The sandstone units (20-50 m) are interpreted to have been deposited in offshore bars and in the various subenvironments of the shoreface.

If the covered intervals of the Hampton Formation (which represent the later part of Time 2) were deposited in offshore environments, then

one of three things was happening: (1) Time 2 experienced pulses of rapid transgression and the vertical sequence represents a coarsening-upward trend that occurred in response to shoreline progradation and a slowed rate of transgression; (2) periods of progradation alternated with periods of gradual transgression, thus generating a fining-upward sequence; or (3) the alternation of coarse- and fine-grained units represents the gradual detachment of channel-shoal couplets and their migration into deeper-water shelf silt and sand (Swift and Field, 1981). No one option can be chosen at this point in the study of the Cambrian of east Tennessee. However, the model for Time 3 constrains the depositional model for Time 2.

Time 3 (Fig. 41C) is interpreted as stable, sand-starved shelf over which sand ridges migrated in response to tidally enhanced storm currents. The two modern analogues for Time 3 are the tidal current ridges of the North Sea (Houboult, 1968) and the storm-maintained ridges of the Maryland inner shelf (Duane et al. 1972; Swift, 1975; Swift and Field, 1982). A storm-dominated shelf is proposed for Time 3 based upon the facies association of the Erwin Formation (Fig. 32, p. 55) and an absence of tidally formed structures. The shelf must have been segregated into hydraulically different areas consisting of large sand ridges, patches of megarippled sand and areas affected only by storm currents. The sand ridges and patches originated as the channel-shoal couplets of Time 2 and were maintained by a mean flow generated by a combination of tidal and storm currents. Modern shelf sand ridges are

thought to have originated as channel-shoal couplets (Swift and Field, 1981) and are therefore relicts of transgression. However, tidal and storm currents may continue to follow the channels once the couplet has become detached; therefore the bedforms are maintained by the hydraulic regime present on the shelf (Swift, 1975).

In the North Sea (Houbolt, 1968) and the Celtic Sea (Belderson and Stride, 1966), fine-grained sediment accumulated near the end of a tidal current transport path. Mud appears to be accumulating in areas that experience a low frequency of wave activity (McCave, 1970, 1971) such as between sand ridges (Walker, 1984) or down-current from ridges. Anderton (1976) also suggested that different bedforms developed along a tidal current transport path during the deposition of the Late Precambrian Jura Quartzite.

The hydrodynamics of sand ridges and their maintenance is not fully understood. However, several workers have pointed out that original topographic highs tend to grow upward (Figueirdo et al., 1981; Parker et al., 1982). Huthnance (1982) formulated a mathematical model of formation for the tidally maintained ridges of the North Sea. In general terms it states that when flow is oblique to the ridge crest, the cross-ridge component of flow must accelerate due to decreasing cross-sectional area. Thus erosion occurs on the up-current side of the sand ridge and aggradation on the down-current side (Swift and Field, 1981); thus the up-current slope develops a coarse lag and fines accumulate on the down-current slope and within the troughs between ridges (Walker, 1984). Seafloor irregularities, such as drowned shoreface shoals that

are originally oriented obliquely to geostrophic flow or to tidal flow, will tend to grow and be maintained in a shelf setting.

There are many ancient sequences which are comparable to any one or two facies of the Chilhowee Group (e.g., the Cape Sebastian Sandstone (Bourgeois, 1980) is a good ancient analogue for Facies Shcs.) However, few workers in the ancient have documented the transgressive sequence from fluvial/marginal marine environments to storm/tide dominated shelf environments. Such ancient examples are summarized in Table 4. Most of the sequences cited represent an overall transgression (Driese et al., 1981; Hereford, 1977; Saller and Dickinson, 1982; Cotter, 1983; Houseknecht and Ethridge, 1978; Watchorn, 1978). A few sequences are interpreted to represent an interfingering of terrestrial and marine deposits and the nature of the overlying strata was not discussed (Eriksson, 1978: Daily et al., 1980).

Note that most of these sequences are pre-Devonian and there is no mention of barrier island development. This is more than likely related to the role of vegetation in stabilizing coastal deposits. The shoreface of these ancient sequences is dominated by tidal channel and/or tidal flat sedimentation rather than barrier island development.

TABLE 4. Summary of several ancient transgressive sequences.

FACIES	ENVIRONMENTS
Mt. Simon Fm., Upper Cambrian (Driese et al., 1981)	
Medium- to large-scale planar-tabular cross-stratified pebbly sandstone, minor siltstone and shale, sparse Skolithos and Arenicolites.	Braided fluvial-marine foreshore
Small- to medium-scale planar-tabular and trough cross-stratified sandstone, Skolithos and Arenicolites; with siltstone and shale, Cruziana ichnofacies.	Tidal channel and adjacent flat of the lower tidal flat.
Bioturbated, structureless sandstone dessiccation cracks, Skolithos, brachiopod Obolus.	Midtidal flat
Tapeats Sandstone, Cambrian (Hereford, 1977)	
Large-scale cross-stratified sandstone, with compound cross-stratification and reactivation surfaces, herringbone cross-stratification; parallel stratification; fining-upward cycles.	Beach and tidal channel
Arkosic pebble conglomerate, no trace fossils.	Braided fluvial
Moodies Group, Archean (Eriksson, 1978)	
Matrix- and clast-supported conglomerate.	Proximal alluvial plain, mass flows
Medium- to large-scale trough cross-stratified sandstone capped with ripple cross-lamination and siltstone, desiccated shale-partings.	Midalluvial plain, channel bars
Small- to medium-scale trough cross-stratified sandstone, fining and thinning upward cycles, desiccated shale partings.	Distal alluvial plain, channel fill sequences
Plane-bedded, well-sorted sandstone with lenses of well-rounded pebbles.	Swash bar
Lower Cambrian of Kangaroo Island, South Australia (Daily et al., 1980)	
Interbedded siltstone and minor sandstone with ripple cross-lamination and flat lamination, flaser and lenticular bedding, trilobite traces, desiccation cracks.	Subtidal or intertidal deposit
Horizontally bedded, cobble to boulder conglomerate, trough cross-stratified sandstone and conglomerate.	Alluvial fan-braided stream complex
Trough cross-stratified and planar-laminated sandstone.	Subtidal channels

TABLE 4 (continued)

FACIES	ENVIRONMENTS
Battle Fm., Pennsylvanian (Saller and Dickinson, 1982)	
Clast-supported conglomerate.	Proximal braided stream system
Sandy conglomerate with rare cross-stratification.	Midbraided stream system.
Cross-stratified conglomerate.	Distal braided stream
Interbedded calcareous sandstone, conglomerate and mudstone.	Tidal and deltaic origin
* * *

V. CONCLUSION

This study examined a single nearly complete exposure of the Chilhowee Group in order to define and understand the facies that are present in northeastern Tennessee. Understanding lateral facies relationships was not a goal of this study due to the structural complexity and poor exposure of the Lower Cambrian outcrop belt, and because of the lack of facies information. Field description of the rocks exposed along the Doe River revealed that four facies are present in the Chilhowee Group. Reconaissance work in eastern Tennessee has shown that these four facies are present at other outcrops, however, one or two facies may dominate to the exclusion of the others.

The four facies are:

1. Facies G--immature, conglomeraticarkosic lithic wacke with large-scale cross-statification as well as thick horizontally stratified, fine-grained lithic wacke.

2. Facies S--submature, medium- to very coarse-grained subarkosic to arkosic arenite with medium- to large-scale cross-stratification interbedded with thin, discontinuous laminated siltstone and sandstone.

3. Facies Shcs--micaceous, very fine- to fine-grained subarkosic to arkosic arenite with hummocky cross-stratification interbedded with bioturbated siltstone and sandstone.

4. Facies QA--supermature medium- to coarse-grained quartz arenite with either low-angle cross-stratification or large-scale, planar-tabular cross-stratification.

Each facies represents a depositional setting in which-subenvironments commonly are represented by subfacies. Facies G is interpreted to have been deposited within a braided stream-alluvial plain environment. Its subfacies are channel bar deposits, interbar pool deposits, and vertical accretion deposits. The fluvial system deposited sediment on a coastal plain because Facies G and Subfacies QA-Unicoi are associated with each other in the terrestrial-marine transition in the Unicoi Formation. Subfacies QA-Unicoi represents either a ridge-and-runnel system or a nearshore bar system.

Facies S is interpreted to consist of subtidal channel-fill sequences, although the exact nature of the bars between which the channels were located is unknown because complete bar sequences were not preserved. Facies S subenvironments include both shallow and deep channel deposits. Slack-water periods are evident by the fine-grained laminated units which drape the cross-stratified channel sandstones.

Facies Shcs is interpreted to represent storm-deposition in the offshore region of a shelf. Because Facies Shcs is intimately associated with both Facies S and Subfacies QA-Erwin, those facies relationships are important to this interpretation.

Facies QA-Erwin is thought to have been deposited on a sand-starved shelf on which sand-sized relict sediment was swept into large sand ridges. The development of the sand ridges was probably related to the detachment of channel-shoal couplets of Facies S. The "drowned" channels served as pathways for tidally enhanced storm currents on the shelf and shoals were reworked into sand ridges by the existing hydraulic

regime. All the sand was not swept into ridges, because Facies S is commonly interbedded with Facies Shcs. Therefore, both sand ridges (Subfacies QA-Erwin) and sand patches (Facies S) existed on the shelf.

In areas of the shelf not affected by tides, only storm-generated processes were important. Such areas include troughs between sand ridges and at the end of the tidal/storm current transport path. Vertical repetition of the interbedding of Facies S, Facies Shcs, and Subfacies QA-Erwin is a result of the migration of the sand ridges and patches over hummocky storm sand and fairweather silt.

REFERENCES

REFERENCES

Alpert, S. P., 1975, Planolites and Skolithos from the upper Precambrian-lower Cambrian, White Inyo Mountains, California: Jour. Paleontol., v. 49, p. 509-521.

Allen, J. R. L., 1980, Sandwaves: a model of origin and internal structure: Sedimentary Geology, v. 26, p. 281-328.

Anderton, R., 1976, Tidal-shelf sedimentation an example from the Scottish Dalradian: Sedimentology, v.23, p. 429-458.

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

Belderson, R. H., and Stride, A. H., 1966, Tidal current fashioning of a basal bed: Marine Geology, v. 4, p. 237-257.

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

Boothroyd, J. C., and Ashely, G. M., 1975, Processes, bar morphology and sedimentary structures on braided outwash fans, northeastern Gulf of Alaska, in Jopling, A. V., ed., Glaciofluvial and Glaciolacustrine Sedimentation: Soc. Econ. Paleo. Miner., Spec. Pub. 22, p. 193-222.

Boothroyd, J. C., and Nummedal, D., 1978, Proglacial braided outwash: a model for humid alluvial fan deposits, in Miall, A.D., ed., Fluvial Sedimentology: Canadian Soc. of Petrol. Geol., p. 641-668.

Bouma, A. H., Hampton, M.A., and Orlando, R.C., 1977, Sandwaves and other bedforms in lower Cook Inlet, Alaska: Marine Geotechnology, v. 2, p. 291-308.

Bourgeois, J., 1980, A transgressive shelf sequence exhibiting hummocky stratification: the Cape Sebastian Sandstone (Upper Cretaceous), southwestern Oregon: Jour. Sed. Pet., v. 50, p. 681-702.

Brown, W. R., 1970, Investigations of the sedimentary record in the Piedmont and Blue Ridge of Virginia, in Fisher, G. W., Pettijohn, F. J., Reed, J. C., Jr., and Weaver, K. N., eds., Studies of Appalachian Geology: central and Southern: New York, John Wiley-Interscience, p. 335-349.

Burst, J. F., 1958a, "Glauconite" Pellets: their mineral nature and applications to stratigraphic interpretations: Bull. Am. Assoc. Pet. Geol., v. 42, p. 310-327.

Burst, J. F., 1958b, Mineral heterogeneity in glauconite pellets: Am. Miner., v. 43, p. 481-497.

Byers, C. W., and Dott, R. H., Jr., 1981, SEPM research conference on modern shelf and ancient cratonic sedimentation--the orthoquartzite-carbonate suite revisited: Jour. Sed. Pet., v. 51, p. 329-347.

Byers, C. W., and Dott, R. H., Jr. , 1982, Cambrian sedimentology of the northern Midwest-search for a depositional model: Geol. Soc. Am., Abst. with Programs, v. 14, p. 457.

Caston, V. N. D., 1972, Linear sand banks in the southern North Sea: Sedimentology, v. 18, p. 63-78.

Caston, V. N. D., and Stride, A. H., 1973, Influence of older relief on the location of sand waves in a part of the southern North Sea: Estuarine Coastal Marine Sci., v. 1, p. 379-386.

Clifton, H. E., 1969, Beach lamination: nature and origin: Marine Geol., v. 7, p. 553-559.

Coleman, J. M., 1969, Brahmaputra River channel processes and sedimentation: Sedimentary Geol., v. 3, p. 129-239.

Collinson, J, D., 1970, Bedforms of the Tana River, Norway: Geogr. Ann., v. 52A, p. 31-55.

Cotter, E., 1978, The evolution of fluvial style with special reference to the Central Appalachians, in Miall, A. D., ed., Fluvial Sedimentology: Canadian Soc. of Pet. Geol., p. 361-381.

Cotter, E., 1983, Sea level fluctuations in the origin of the Tuscarora Formation: Jour. Sed. Pet., v. 53, p. 25-50.

Crimes, T. P., and Harper, J. C., eds., 1977, Trace Fossils 2: Seel House Press, Liverpool, 351 p.

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

Daily, B., Moore, P. S., and Rust, B. R., 1980, Terrestrial-marine transition in the Cambrian rocks of Kangaroo Island, South Australia: Sedimentology, v. 27, p. 379-399.

Dalrymple, R. W., Knight, R. J., Lambiase, J. J., 1978, Bedforms and their hydraulic stability relationships in a tidal environment, Bay of Fundy, Canada: Nature, v. 275, p. 100-104.

Davidson-Arnott, R. G. D., and Greenwood, B., 1976, Facies relationships on a barrier coast, Kouchibouguac Bay, New Brunswick, Canada, in Davis, R. A., Jr., and Ethington, R. L. (eds.), Beach and Nearshore Sedimentation: Soc. Econ. Paleo. Miner., Spec. Pub. no. 4., p. 148-168.

Davis, R. A., Jr., 1983 Depositional Systems, a Genetic Approach to Sedimentary Geology: Prentice Hall, Inc., Englewood Cliffs, NJ, 669 p.

Davis, R. A., Jr., Fox, W. T., Hayes, M. O., and Boothroyd, J. C., 1972, Comparison of ridge-and-runnel systems in tidal and nontidal environments: Jour. Sed. Pet., v. 32, p. 413-421.

Dickinson, W. R., 1970, Interpreting detrital modes of graywacke and arkose: Jour. Sed. Pet., v. 40, p. 695-704.

Dott, R. H., Jr., 1964, Wacke, graywacke and matrix--What approach to immature sandstone classification?: Jour. Sed. Pet., v. 34, p. 625-632.

Dott, R. H., Jr., 1966, Eocene deltaic sedimentation at Coos Bay, Oregon: Jour. Geol., v. 74, p. 373-420.

Dott, R. H., Jr., and Bourgeois, J., 1982, Hummocky stratification: significance of its variable bedding sequences: Geol. Soc. Am. Bull., v. 93, p. 663-680.

Duane, D. B., Field, M. E., Meisburger, E. P., Swift, D. J. P., and Williams, S. J., 1972, Linear shoals on the Atlantic inner continental shelf, Florida to Long Island, in Swift, D. J. P., Duane, D. B., and Pilkey, O. H., eds., Shelf Sediment Transport: Process and Pattern: Dowden, Hutchinson and Ross, Stroudsburg, PA, 630 p.

Figueiredo, A. G., Swift, D. J. P., Stubblefield, W. L., and Clark, T. L., 1981, Sand ridges on the inner Atlantic Shelf of North America: morphometric comparisons with the Huthnance stability model: Geo-Marine Letters, v, 1, p. 187-191.

Field, M. E., and Roy, P. S., 1984, Shallow-marine sand bodies formed during stillstand conditions, southeastern Austrailia, in Abstracts: First annual midyear meeting, Soc. Econ. Paleo. Miner., p. 31.

Folk, R. L., 1968, Petrology of Sedimentary Rocks: Hemphills Bookstore, Austin, Texas, 170 p.

Frey, R. W., ed., 1975, The Study of Trace Fossils: A Synthesis of Principles, Problems, and Procedures in Ichnology: Springer-Verlag, New York, 562 p.

Frey, R. W., and Pemberton, S. G., 1984, Trace fossil facies models, in Walker, R. G., ed., Facies Models, Second Edition: Geoscience Canada, Reprint Series 1, p. 189-207.

Goodwin, P. W., and Anderson, E. J., 1974, Associated physical and biogenic structures in environmental subdivision of a Cambrian tidal sand body: Jour. Geol., v. 82, p. 779-794.

Hamblin, A. P., and Walker, R. G., 1979, Storm dominated shelf deposits in the Fernie-Kootenay (Jurassic) transition, southern Rocky Mountains: Can. Jour. Earth Sci., v. 16, p. 1673-1690.

Häntzschel, W., 1975, Trace fossils and problematic, sin Treatise on invertebrate paleontology, Part W, Miscellanea, Supplement 1: Lawrence, University of Kansas Press and Geol. Soc. Am., 269 p.

Harms, J. C., Southard, J. B., and Walker, R. G., 1982, Structures and Sequences in Clastic Rocks, Soc. Econ. Paleo. Miner. Short Course no. 9, not consecutively numbered.

Hayes, M. O., 1980, General morphology and sediment patterns in tidal inlets: Sedimentary Geology, v. 26, p. 139-156.

Hereford, R., 1977, Deposition of the Tapeats Sandstone (Cambrian) in central Arizona: Geol. Soc. Am. Bull., v. 88, p. 199-211.

Heward, A. P., 1981, A review of wave-dominated clastic shoreline deposits: Earth-Science Reviews, v. 17, p. 223-276.

Hine, A. C., and Boothroyd, J. C., 1978, Morphology, processes and recent sedimentary features of a glacial-outwash plain shoreline, southern Iceland: Jour. Sed. Pet., v. 48, p. 901-920.

Houbolt, J. J. H. C., 1968, Recent sediments in the southern bight of the North Sea: Geol. Mijnb., v. 47, p. 245-273.

Houseknecht, D. W., and Ethridge, F. G., 1978, Depositional history of the Lamotte Sandstone of southeastern Missouri: Jour. Sed. Pet., v. 48, p. 575-586.

Howard, J. D., and Reineck, H. E., 1972, Physical and biogenic sedimentary structures of the nearshore shelf, Senckenber. Mar., v. 4, p. 217-222.

Hower, J., 1961, Some factors concerning the nature and origin of glauconite: Am. Miner., v. 46, p. 313-334.

Huthnance, J. M., 1982, On one mechanism forming linear sand banks: Estuarine, Coastal, Shelf Science, v. 14, p. 79-99.

Ingram, R. L., 1954, Terminology for the thickness of stratification and parting units in sedimentary rocks: Geol. Soc. Am. Bull., v. 65, p. 937-938.

Johnson, H. D., 1977, Shallow-marine sand bar sequences: an example from the Late Precambrian of north Norway: Sedimentology, v. 24, p. 245-270.

Keith, A., 1895, Description of the Knoxville quadrangle: U.S. Geol. Survey Geol. Atlas, Folio 16.

Kerr, P. F., 1977, Optical Mineralogy: McGraw-Hill Book Company, New York, 492 p.

King, P. B., and Ferguson, F. W., 1960, Geology of northeasternmost Tennessee : U.S. Geol. Surv. Prof. Paper 311, 136 p.

Klein, G. deV., 1970, Tidal origin of a Precambrian quartzite--the Lower Fine-grained Quartzite (Middle Dalradian) of Islay, Scotland: Jour. Sed. Pet., v. 40, p. 973-985.

Klein, G. deV., 1977, Clastic Tidal Facies: Continuing Education Publication Company, Champaign, Illinois, 149 p.

Komar P. D., Neudeck, R. H., Kulm, L. D., 1972, Observations and significance of deeper water oscillatory ripple marks on the Oregon Continental Shelf, in Swift, D. J .P., Duane, D. B., Pilkey, O. H., eds., Shelf Sediment Transport: Process and Pattern: Dowden, Hutchinson and Ross, Stroudsburg, Pa., p. 601-619.

Ksiazkiewicz, M., 1970, Observations on the ichnofauna of the Polish Carpathians, in Crimes, T. P., and Harper, J. C., eds., Trace Fossils: Geol. Jour. Spec. Issue 3, Seel House Press, Liverpool, 283 p.

Kumar N. and Sanders, J. E., 1974, Inlet sequence: a vertical succession of sedimentary structures and textures created by the lateral migration of tidal inlets: Sedimentology v. 21, p. 491-532.

Levell, B. K., 1980, Evidence for currents associated with waves in Late Precambrian shelf deposits from Finnmark, north Norway: Sedimentology, v. 27, p. 153-166.

Mack,G. H., 1980, Stratigraphy and depositional environments of the Chilhowee Group in Georgia and Alabama: Am. Jour. Sci., v. 280, p. 497-517.

Maher, L. J. and Dott, R. H., Jr., 1979, Program for analysis of directional orientation data (for generalized Basic): unpublished computer program, University of Wisconsin, Madison, Wisconsin.

Marsaglia, K. M., and Klein, G. deV., 1983, The paleogeography of Paleozoic and Mesozoic storm depositional systems: Jour. Geol., v. 91, p. 117-142.

McCave, I. N., 1970, Deposistion of fine-grained suspended sediment from tidal currents: Jour. Geophys. Res., v. 75, p. 4151-4159.

McCave, I. N., 1971, Wave effectiveness at the sea bed and its relationship to bedforms and deposition of mud: Jour. Sed. Pet., v. 41, p. 89-96.

McCabe, P. J., and Jones, C. M., 1977, Formation of reactivation surfaces within superimposed deltas and bedforms: Jour. Sed. Pet., v. 47, p. 707-715.

McKee, E. D., and Weir, G. W., 1953, Terminology for stratification and cross-stratification in sedimentary rocks: Geol. Soc. Am. Bull., v. 64, p. 381-390.

McKee, E. D., Crosby, E. J., Berryhill, H. L., Jr., 1967, Flood deposits, Bijou Creek, Colorado, June 1965: Jour. Sed. Pet., v. 37, p. 829-851.

Miall, A. D., 1977, A review of the braided-river depositional environment: Earth Science Reviews, v. 13, p. 1-62.

Middleton, G. V., and Southard, J. B., 1978, Mechanics of sediment movement: Soc. Econ. Paleon. Miner. Short Course 3, 246 p.

Mount, J. F., 1982, Storm-surge-ebb origin of hummocky cross-stratified units of the Andrews Mt. Member, Campito Formation (Lower Cambrian), White-Inyo Mountains, Eastern California: Jour. Sed. Pet., v. 52, p. 941-958.

Mowbray, T. de, and Visser, M. J., 1984, Reactivation surfaces in subtidal channel deposits, Oosterschelde, southwest Netherlands: Jour. Sed. Pet., v. 54, p. 811-824.

Neuman, R. B., and Nelson, W. H., 1965, Geology of the Western Great Smoky Mountains, Tennessee: U.S. Geol. Surv. Prof. Paper 349-D, 81 p.

Narayan, J., 1977, Sedimentary structures in the Lower Greensand of the Weald, England and Bas-Boulannais, France: Sedimentary Geology, v. 6, p. 73-109.

Nummedal, D., et al., 1974, Recent migration of the Skeidarasandur shoreline, southwest Iceland: Final Report for Contract # NG0921-73-6-0258, Naval Ordinance Lab., Dept. of Geol., Univ. of South Carolina, 183 p.

Odin, G. S., and Matter, A., 1981, De glauconiarum origine: Sedimentology, v. 28, p. 611-641.

Owens, E. H., and Frobel, D., 1977, Ridge and runnel systems in the Magdalen Islands, Quebec: Jour. Sed. Pet., v. 47, p. 191-198.

Parker, G., Lanfredi, N. W., and Swift, D. J. P., 1982, Seafloor response to flowin a southern hemisphere sand ridge field: Argentine inner shelf: Sedimentary Geology, v. 33, p. 195-216.

Pemberton, S. G., and Frey, R. W., 1982, Trace fossil nomenclature and the Planolities-Paleophycus dilemma: Jour. Paleo., v. 56, p. 843-881.

Pettijohn, F. J., Potter, P. E., and Siever, R., 1973, Sand and Sandstones: Springer-Verlag, New York, 618 p.

Phillips, H. E., 1952, The geology of the Starr Mountain area, southeast Tennessee: Univ. of Tennessee unpublished Master's thesis, Knoxville, TN.

Rackley, R. I., 1951, Geology of Bean Mountain area, southeast Tennessee: Univ. of Tennessee unpublished Master's thesis, Knoxville, TN.

Ragan, C. M., 1973, Structural Geology, An Introduction to Geometrical Techniques: John Wiley and Sons, Inc., 208 p.

Reading, H. G., ed., 1978, Sedimentary Environments and Facies: Elsevier, New York, 557 p.

Reineck, H. E., and Singh, I. B., 1980, Depositional Sedimentary Environments: Springer-Verlag, New York, 549 p.

Reinson, G. E., 1984, Barrier island systems, in Walker, R. G., ed., Facies Models Second Edition: Geological Assoc. of Canada, Waterloo, Ontario, p. 57-74.

Ruegg, G. H. J., 1977, Features of Middle Pleistocene sandur deposits in the Netherlands, Geol. Mijnb., v. 56, p. 5-24.

Rust, B. R., 1972, Structure and process in a braided river: Sedimentology, v. 18, p. 221-246.

Rust, B. R., 1978, Depositional models for braided alluvium, In Miall, A. D., ed., Fluvial Sedimentology: Canadian Soc. Pet. Geol. p. 605-625.

Rust, B. R., 1981, Coarse alluvial deposits, in Walker, R. G., ed., Facies Models, Geoscience Canada, p. 9-21.

Saxov, S., and Nieuwenhuis, J. K., eds., 1982, Marine Slides and Other Mass Movements: Plenum Press, New York, 353 p.

Saller, A. H. and Dickinson, W. R., 1982, Alluvial to marine facies transition in the Antler overlap sequence, Pennsylvanian and Permian of north-central Nevada: Jour. Sed. Pet., v. 52, p. 925-940.

Schwab, F. L., 1970, Origin of the Antietam Formation (Late Precambrian-Lower Cambrian), central Virginia: Jour. Sed. Pet., v. 40, p. 354-366.

Schwab, F. L., 1971, Harpers Formation, central Virginia: a sedimentary model: Jour. Sed. Pet., v. 41, p. 139-149.

Schwab, F. L., 1972, The Chilhowee Group and the Late Precambrian-Early Paleozoic sedimentary framework in the central and southern Appalachians, in Lessing, P., Hayhurst, R. I., Barlow, J. A., and Woodfork, L. D., eds., Appalachian Structures: Origin, Evolution and Possible Potential for New Exploation Frontiers, a Seminar: West Virginia Univ. and West Virginia Geol. and Econ. Survey, p. 59-101.

Seilacher, A., 1978, Use of trace fossil assemblages for recognizing depositional environments, in Basan, P. B., ed., Trace Fossil Concepts: Soc. Econ. Paleo. Miner. Short Course 5, p. 185-201.

Sharma, G. D., Naidu, A. S., and Hood, D. W., 1972, Bristol Bay: a model contemporary graded shelf: Bull. Am. As soc. Pet. Geol., v. 56, p. 2000-2012.

Singh, I. B., and Kumar, S., 1974, Mega- and giant ripples in the Ganga, Yamuna, and Son rivers, Utter Pradesh, lndia: Sedimentary Geology, v. 12, p. 53-66.

Smith, N. D., 1970, The braided stream depositional environment: comparison of the Platte River with some Silurian clastic rocks, north-central Appalachians: Geol. Soc. Am. Bull., v. 81, p. 2993-3014.

Smith, N. D., 1971, Pseudo-planar stratification produced by very low amplitude sand waves: Jour. Sed. Pet., v. 41, p. 69-73.

Stubblefield, W. L., and Swift, D. J. P., 1976, Ridge development as revealed by sub-bottom profiles on the central New Jersey shelf, Marine Geology, v. 20, p. 315-334.

Swift, D. J. P., 1975, Tidal sand ridges and shoal-retreat massifs: Marine Geology, v. 18, p. 105-134.

Swift, D. J. P., Duane, D. B., and Pilkey, O. H., eds., 1972, Shelf Sediment Transport: Process and Pattern: Dowden, Hutchinson and Ross, Stroudsburg, PA, 656 p.

Swift, D. J. P., and Field, M. E., 1981, Evolution of a classic sand ridge field; Maryland sector North American inner shelf: Sedimentology, v. 28, p. 461-482.

Swift, D. J. P., Freeland, G. L., and Young, R. A., 1979, Time and space distribution of megaripples and associated bedforms, Middle Atlantic Bight, North American Atlantic Shelf: Sedimentology, v. 26, p. 389-406.

Swift, D. J. P., Tillman, R. W., Siemers, C. T., and Rine, J., in press, Fluid and sediment dynamics on a modern shelf: implications for models of ancient shelf sedimentation: Sedimentology.

Swift, D. J. P., et al., 1983, Hummocky cross-stratification and megaripples: a geological double standard?: Jour. Sed. Pet., v. 53, p. 1295-13.

Swingle, G. D., 1949, Petrography of the Chilhowee Group, near Walland, Tennessee: Univ. of Tennessee unpublished Master's thesis, Knoxville, Tennessee.

Visher, G. S., and Howard, J. D., 1974, Dynamic relationships between hydraulics and sedimentation in the Altahara Estuary: Jour. Sed. Pet., v. 44, p. 502-521.

Walker, R. G., ed., 1979, Facies Models: Geoscience Canada, Reprint Series 1, Geol. Assoc. Canada, 211 p.

Walker, R. G., 1984, Shelf and shallow-marine sands, in Walker, R. G., ed., Facies Models Second Edition: Geoscience Canada, Reprint Series 1, Geol. Assoc. Canada, p. 141-170.

Walker, R. G., and Cant, D. J., 1984, Sandy fluvial systems, in Walker, R. G., ed., Facies Models Second edition: Geoscience Canada, Reprint Series 1, Geol. Assoc. Canada, p. 71-89.

Watchorn, M. B., 1980, Fluvial and tidal sedimentation in the 300 MA Mozaan basin of South Africa: Precambrian Research, v. 13, p. 27-42.

Werner, F., and Newton, R. S., 1975, The pattern of large-scale bedforms in the Langeland Belt (Baltic Sea): Marine Geology, v. 14, p. 29-59.

Wescott, W. A., and Ethridge, F. G., 1980, Fan delta sedimentology and tectonic setting-Yallahs Fan Delta, southeast Jamaica: Am. Assoc. Pet. Geol. Bull., v. 64, p. 374-399.

Whisonant, R. C., 1974, Petrology of the Chilhowee Group (Cambrian and Cambrian (?)) in central-eastern and southern Tennessee: Jour. Sed. Pet., v. 44, p. 228-241.

Wunderlich, F., 1978, Deposition of mud in the giant ripples of inner Jade, German Bight, North Sea: Senckenbergiana marit., v. 10, p. 256-267.

Ziegler, A. M., and Scotese, C. R., McKerrow, M. E., and Bambach, R. K., 1979, Paleozoic paleogeography, Ann. Rev. of Earth and Planetary Sciences, v. 7, p. 473-502.

APPENDIXES

APPENDIX A

MEASURED SECTION DESCRIPTIONS

The study area (36°17' latitude and 82°11' longitude) is located on Rt. 19E along the Doe River between Valley Forge and Hampton, in northeastern Tennessee. The Chilhowee Group at this locality is a complete section except for the lower part of the Unicoi Formation which has been removed by the Iron Mountain Fault. However, the section is not completely exposed.

The sections examined include the upper 205 m of the Unicoi Formation, the middle 45 m of the Hampton Formation and the uppermost 110 m of the Erwin Formation. The contact between the Erwin Formation and the overlying Shady Dolomite is transitional and consists of interbedded shale and sandstone of the Helenmode Member of the Erwin Formation. The badly weathered nature and fine-grained character of the Helenmode made study impossible. For that reason, only the rocks at the base of the Helenmode were given a cursory examination.

The following descriptions begin 180 m above the basal fault contact of the Unicoi Formation and continue through the base of the Helenmode Member at the top of the Erwin Formation. Thicknesses of covered intervals are from descriptions and mapping of King and Ferguson (1960). Unit designations were used to simplify field study and are based on lithology and weathered appearance (i.e., ledge vs. recess). Bedding thickness, cross-set thickness and grain size conventions are listed below. The rock classification used is that of Folk (1968) and Dott (1964).

Bedding thickness and cross-stratification set thickness (terminology after McKee and Weir, 1953, as modified by Ingram, 1954):

Bedding	Cross-Stratification	Thickness m
Very thick-bedded		>1.0
Thick-bedded	Large-scale	0.3-1.0
Medium-bedded	Medium-scale	0.1-0.3
Thin-bedded	Small-scale	0.03-0.1
Very thin-bedded		0.01-0.03
Laminated		0.003-0.01
Thinly laminated		<0.003

Grain size descriptions (Wentworth, 1922):

Size Class	Diameter mm
Pebbles	4.0-64.0
Granules	2.0-4.0
Very coarse sand	1.0-2.0
Coarse sand	0.5-1.0
Medium sand	0.25-0.5
Fine sand	0.125-0.25
Very fine sand	0.0625-0.125
Silt	<0.0625

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
Unicoi Formation			
		180.00	Covered (measurements from King and Ferguson, 1960).
1	4.80	184.80	Medium- to coarse-grained, rounded, moderately sorted subarkosic arenite; occasional granules; large-scale planar-tabular cross-stratification.
			SAMPLE: 11-12-1 U.
2	1.30	186.10	Interbedded (1) very-fine to very coarse grained, subrounded to rounded, poorly sorted feldspathic lithic wacke with planar-tabular cross-stratification; megaripple bedform preserved and (2) laminated siltstone; siltstone drapes megaripple; fining-upward sequence is 5-30 cm thick.
			SAMPLES: 11-17-1 U, 11-18-1 U.
3	4.90	191.00	Very fine-grained, chloritic feldspathic lithic wacke with thin siltstone beds; lenses of conglomerate at base; ripple cross-laminations in sandstone; dewatering structures; rare Paleophycus.
4	5.79	196.79	Fine- to medium-grained feldspathic lithic wacke with scattered laminae of granules and pebbles (quartz and feldspar); graded laminae, low-angle cross-stratification, cross-stratified granule and pebble lenses, small-scale trough and planar-tabular cross-stratification.
			SAMPLE: 12-11-1 U.

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
5	11.63	208.42	Interbedded (1) fine-grained to pebbly feldspathic lithic wacke and (2) thin beds of laminated silty fine-grained sandstone; large-scale planar-tabular cross-stratification in 1.5-3.0 m sets. SAMPLES: 12-11-2 U, 12-12-3 U, 8-22-2 U.
6	10.22	218.64	Medium-grained, rounded to well-rounded quartz arenite; faint large-scale planar-tabular cross-stratification and low-angle cross-stratification; sets laterally discontinuous and set boundaries are indistinct; broad shallow scours with <0.5 m of relief; symmetrical ripple forms; thin, discontinuous, hematitic, silty conglomerate may mark the scour surfaces. SAMPLES: 12-13-1 U, 12-13-3 U.
7	2.61	221.25	Fine- to coarse-grained, subrounded, subarkosic wacke; low-angle cross-stratification that is graded; 0.5 m beds that coarsen upward and are overlain by thinly interbedded siltstone and conglomerate. SAMPLE: 12-13-5 U.
8	4.39	225.64	Fine- to medium-grained, rounded, micaceous arkosic arenite; medium-scale trough and planar-tabular cross-stratification; granule lags common at base of cross-sets. SAMPLE: 12-13-8 U.
9	6.31	231.95	Interbedded 10-30 cm thick (1) coarse-grained to pebbly arkosic wacke which appears massive but

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
			exhibits inverse grading and (2) medium- to coarse-grained subarkosic arenite with planar-tabular cross-stratification; laminated is continuous siltstone drape both coarse-grained lithologies. SAMPLE: 12-16-3 U.
10	3.09	235.04	Coarse-grained to pebbly lithic arkosic wacke; large-scale sets of low-angle cross-stratification, graded foresets 4 cm thick.
11	3.54	238.58	Medium- to very coarse-grained subarkosic arenite; large-scale sets of low-angle cross-stratification; scattered sets of medium-scale, planar-tabular cross-stratification.
12	33.57	272.15	Fine- to medium-grained, subangular to subrounded, arkosic lithic wacke; low-angle cross-stratification with medium-scale sets bounded by low-angle erosion surfaces, granule laminae and lenses; unit gradually coarsens upward from siltstone with granule lenses at base to a laminated sandstone with no siltstone or conglomerate lenses to a laminated sandstone with abundant laminae and lenses of granules and pebbles. SAMPLES: 1-7-2 U, 2-18-2 U.
13	10.30	282.45	Medium- to coarse-grained, subrounded, well-sorted, arkosic arenite; granule laminae common, medium-scale trough and planar-tabular cross-stratification. SAMPLE: 1-7-5 U.

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
14	14.35	296.80	Medium- to very coarse-grained, rounded well-sorted quartz arenite; graded low-angle cross-stratification; heavy mineral laminae abundant; base of unit marked by a low relief scour. SAMPLE: 1-7-6 U.
15	2.83	299.63	Fine- to very coarse-grained subarkosic arenite; 30-50 cm coarsening-upward beds capped with laminated siltstone.
16	10.93	310.56	Interbedded (1) medium- to very coarse-grained subarkosic arenite with abundant heavy minerals; medium- to large-scale trough and planar-tabular cross-stratification; top of set may be marked by discontinuous lenses of siltstone; and (2) black, laminated siltstone with laminae and thin lenses of sandstone, i.e., lenticular bedding; lenses of sandstone may exhibit ripple cross-laminations; the two lithologies are arranged in a fining-upward sequence as the siltstone overlays a cross-stratified sandstone.
17	19.50	330.06	Very fine-grained to fine-grained chloritic quartz wacke; horizontal to low-angle cross-lamination; lenses of granules; lenses of shale clasts (intraformational?); lenses of siltstone; occasional small-scale trough cross-stratification; granule laminae, which increase in abundance up-unit. SAMPLE: 4-7-1 U.
18	12.68	342.74	Interbedded (1) fine- to medium-grained subarkosic arenite and

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
			(2) interlaminated siltstone and sandstone which occurs in beds 30-60 cm thick; stratification sequence commences with a granule lag which is overlain by large-scale trough and planar-tabular cross-stratification, large sets overlain by medium-scale cross-sets; cross-sets within the thinning-upward sequences commonly overlain by interlaminated siltstone and sandstone.
19	9.26	352.00	Interbedded (1) medium- to very coarse-grained subarkosic arenite exhibiting medium-scale trough cross-stratification, very coarse-grained lags at the base of troughs and (2) thinly interbedded siltstone and coarse-grained sandstone with wavy to lenticular bedding.
20	6.79	358.79	Fine-grained to pebbly, poorly sorted, subarkosic wacke; 0.5-1.0 m beds with faint low-angle cross-stratification; granule lags at the base of beds, inverse grading within the uppermost 10-20 cm of a bed; tops of beds are undulatory and are overlain by a laminated siltstone.
21	13.73	372.52	Coarse- to very coarse-grained subarkosic arenite; 30-100 cm beds with faint low-angle cross-stratification; beds may coarsen upward. SAMPLE: 4-28-2 U.
22	10.17	382.69	Medium- to very coarse-grained subarkosic arenite; medium-scale trough cross-stratification; cross-sets overlain by interlaminated siltstone and sandstone; grain

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
			size and set thickness decrease up-unit to fine-grained and small- to medium-scale sets.
23	2.31	385.00	Fine-grained to pebbly, poorly sorted, arkosic wacke; 30 cm massive-appearing beds.
			SAMPLE: 1-7-10 U.
Hampton Formation--Cardens Bluff Shale Member at Base			
	140 m	525.00	Covered (measurements from King and Ferguson, 1960).
24	5.00	530.00	Fine- to very coarse-grained, rounded, well-sorted, subarkosic arenite; large-scale sets of low-angle cross-stratification with graded foresets; medium-scale trough cross-stratification in the uppermost 0.5 m of the unit.
			SAMPLE: 2-19-3 H.
25	7.82	537.82	Medium- to coarse-grained subarkosic arenite; 30-40 cm beds homogenized by biogenic activity; no recognizable traces or stratification; sandstone overlain by thin beds of interlaminated siltstone and fine-grained sandstone which contain horizontal burrows (Planolites).
26	34.36	572.18	Medium- to coarse-grained subarkosic arenite; medium-scale trough and planar-tabular cross-stratification overlain by very thin beds of siltstone; set thickness gradually decreases up-unit to predominantly 10-15 cm; siltstone decreases in thickness upwards from 5 cm to <<1 cm; paired vertical burrows and Skolithos at 55 m.
			SAMPLE: 4-1-2 H.

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
	107.00	679.18	Covered (measurements from King and Ferguson, 1960).
Erwin Formation			
	205.00	884.18	Covered (measurements from King and Ferguson, 1960).
27	19.44	903.62	Interbedded (1) very fine- to fine-grained, purple subarkosic to arkosic arenite and (2) very thin to thin beds of bioturbated siltstone; glauconite locally abundant at 13, 14.5, 15.6 and 18 m above the base of the unit; low- to moderate-angle, medium-scale trough and planar-tabular cross-stratification, each set is capped with a siltstone bed; bases of the sandstone show hypichnial relief of horizontal burrows. SAMPLES: 1-8-1 E, 2-5-3 E.
28	32.10	935.72	Interbedded (1) very fine- to fine-grained subarkosic arenite, beds 1-70 cm thick, with hummocky cross-stratification or low-angle to horizontal lamination, current ripple cross-lamination, symmetrical ripple forms; and (2) moderately bioturbated siltstone; base of the sandstone may show either wrinkle marks, tool marks, casts of Rusophycus and Cruziana; traces are 0.3-1.0 cm wide; glauconite is abundant at the base of the unit; ball-and-pillow structures occur at 907.62 m (30 cm thick) and at 928.62 m (1.5 m thick), convoluted laminae occurs rarely within the laminated sandstone. SAMPLE: 1-8-3 E.

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
29	18.88	954.60	Interbedded (1) fine- to coarse-grained subarkosic and quartz arenites, medium-scale trough cross-stratification and (2) bioturbated siltstone with thin ripple cross-laminated sandstone lenses, siltstone beds range in thickness <1 cm-30 cm; glauconite abundant at 952.20 m; width of Rusophycus and Cruziana 3-10 cm. SAMPLES: 1-8-4 E, 4-1-1 E.
30	4.91	959.51	Medium- to coarse-grained, well-rounded, well-sorted quartz arenite; no siltstone partings so the unit weathers as a massive ledge; faint large-scale planar-tabular cross-stratification; unit thins laterally to about 3 m; upper and lower contacts are sharp; uppermost 30 cm of unit consists of nonfissile shale with lenses of quartz granules and glauconite. SAMPLE: 1-8-5 E.
31	5.67	965.18	Very fine- to fine-grained subarkosic arenite with hummocky cross-stratification interbedded with subordinate amounts of bioturbated siltstone. SAMPLE: 4-29-1 E.
32	3.5	968.68	Fine-grained quartz arenite; medium-scale planar-tabular and trough cross-stratification; very coarse-grained laminae define the foresets.
33	6.70	975.38	Very fine-grained subarkosic arenite; 30-70-cm beds of horizontal laminations; thin beds exhibit hummocky stratification; equally

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
			interbedded with bioturbated siltstone and sandstone. SAMPLE: 4-29-2 E.
34	1.33	976.71	Thinly interbedded (1) siltstone and (2) glauconitic, bioturbated, medium-grained quartz arenite with small-scale planar-tabular cross-stratification; uppermost 8 cm of unit consists of nonfissile shale.
35	0.85	977.56	Medium- to very coarse-grained, well-rounded, well-sorted, quartz arenite; internally unit is faintly cross-stratified although the exact nature of the stratification is indiscernable, very coarse-grained lag at top of unit; upper and lower contact are sharp; upper surface is undulatory. SAMPLE: 4-29-3 E.
36	5.10	982.66	Interbedded (1) fine-grained, glauconitic quartz arenite, 10-30 cm sets of horizontal to low-angle laminae, occasional hummocky stratification and (2) bioturbated siltstone; trains of straight-crested symmetrical ripple marks preserved on the bases of sandstone beds.
37	9.40	992.06	Medium- to coarse-grained, well-rounded, well-sorted quartz arenite; faint large-scale, planar-tabular cross-stratification; sharp base with low-relief scour; sharp upper surface with winnowed lag; uppermost 10 cm consists of nonfissile silty shale. SAMPLE: 1-8-7 E.

UNIT	UNIT THICKNESS (m)	CUMULATIVE THICKNESS (m)	LITHOLOGIC DESCRIPTION
Helenmode Member of the Erwin Formation			
38	2.29	994.35	Interbedded (1) glauconitic, feldspathic quartz arenite, 30-40 cm thick massive-appearing beds and (2) white-beige weathering shales in 10-20 cm beds.

APPENDIX B

BEDFORM NOMENCLATURE

Bedform nomenclature in the literature is rather confusing, so three classification systems have been summarized in the following table (Table 5).

For this thesis, the term sand wave refer to those structures as defined by Allen (1980). For bedforms whose size is larger than small ripples but smaller than sand waves, the term megaripple is used.

TABLE 5. Nomenclature of bedforms based on spacing and height.

CHARACTERISTICS			Harms et al., 1982	Reineck and Singh, 1982	Allen, 1980
spacing	height	other			
<60 cm	<4 cm		small ripples	small ripples	current ripples–unidirectional flow
>60 cm	4 cm–1 m	straight crests sinuous crests	large ripples (megaripples) / sand waves (2-d) / dunes (3-d)	2-dimensional megaripple 3-dimensional megaripple	dunes
25–1000 m	<1–25 m	flow depth > 4 m superimposed megaripples generally straight-crested, undulatory to bifurcating	not discussed	flow transverse: giant ripples flow parallel: tidal current ridges	sand waves

VITA

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