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Abstract

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SOUTHEASTERN GEOLOGY

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COMPARATIVE TECTONICS OF SOME ALPINE AND SOUTHERN APPALACHIAN STRUCTURES

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ABSTRACT

Comparative tectonics is strongly influencing the rapidly developing concepts of structure in the crystalline Southern Appalachians. Burchfiel and Livingston (1967) compared the Brevard zone to Alpine root zones. Major structures along a cross-section of the central Swiss Alps are described in order to provide background for a comparison of structures and concepts between the Alps and Appalachians. The Blue Ridge belt in North Carolina must be a thrust sheet that is probably rooted just northwest of the Brevard zone, at least in the vicinity of the Grandfather Mountain window. An early (Early Paleozoic?) phase of isoclinal folding is recognized in this thrust sheet, in part of the Brevard zone, and over large areas in the Inner Piedmont. A later post-thrusting episode of folding (Late Paleozoic) was most intense in the southeastern Blue Ridge belt and probably over most of the Brevard and Inner Piedmont belts. The Brevard zone is probably a great fundamental fracture zone of long and complex history, that may have responded in different ways to changing stress patterns during the Paleozoic evolution of the Southern Appalachians.

INTRODUCTION

Burchfiel and Livingston (1967) compared the Brevard zone of the Southern Appalachians to Alpine root zones and suggested that the Brevard zone is the root zone for large thrust faults in the Blue Ridge province. This idea is a major contribution to interpretation of Appalachian structure. In order to observe the Alpine root zones first hand, I visited the Swiss Alps for about two weeks in the summer of 1969. Most of my reconnaissance was along a cross-section of the Alps from Luzern to Locarno in central and southern Switzerland. This section includes, from north to south, the following major features: (1) Tertiary Molasse, (2) Helvetic nappes, (3) Aar massif, (4) Urseren zone (root zone for Helvetic nappes), (5) Gotthard massif, (6) Pennine

nappes, (7) root zone for Pennine nappes, and (8) Insubric line, a knife-sharp tectonic boundary with the Southern Calcareous Alps that lie mainly in Italy. I am not an expert in Alpine geology, but offer here some general observations and impressions pertinent to the root zone interpretation of the Brevard zone, parts of which I have studied for several years. I do not imply here that the Alps can be used as a model for the Appalachians, or vice versa. The Alps "are neither the paradigm of all mountain chains nor a regrettable accident of the Earth's surface" (Trümpy, 1960, p. 847) and the same may be said of the Appalachians. In the Alps, the long period of intensive study, excellent exposures, widespread occurrences of fossils, and many "marker beds" have aided the deciphering of very complex stratigraphic and structural problems. The Southern Appalachian crystalline region offers a poor comparison in all these respects, and geologists of the region often turn to comparative tectonics for working hypotheses concerning the structural development.

I am grateful to the following people for reading the manuscript and suggesting improvements: David E. Dunn, Villard S. Griffin, Jr., Robert D. Hatcher, Jr., S. Duncan Heron, Jr., Philip S. Justus, and John L. Livingston. Comments by John Livingston were particularly helpful. Robert E. Jackson kindly provided a translation of the paper by Trümpy (1963). The University Research Council of the University of North Carolina at Chapel Hill provided funds to cover part of my expenses while I was in Switzerland.

DEFINITIONS

A nappe is "a large allochthonous, sheet-like tectonic unit that has moved along a predominantly subhorizontal floor " (Dennis, 1967, p. 111). Nappes may therefore be either large thrust sheets or recumbent folds. The Helvetic nappes are characterized by large-scale thrusting and the Pennine nappes are mainly recumbent isoclinal folds (Lombard and others, 1962). In both regions, however, the individual nappes include subsidiary, but sometimes large-scale, faulting and folding.

A root zone is "a zone where successive nappes have their roots, usually closely appressed" (Dennis, 1967, p. 128). The following discussion by Bailey (1935, p. 54) elucidates the concept: "Outliers of thrust-masses are said to have no roots, or, more strictly, to be separated from their roots. A tectonic root is easily defined in connection with an open anticline. In the case of a recumbent anticline, or of a thrust-mass, a root is the backward, downward penetration of the anticlinal core, or of a thrust-mass, as the case may be." Bailey also gives Termier's picturesque metaphor "roots are vertical chimney-stalks from which horizontal nappes of smoke proceed."

The above definition applies mainly to root zones of recumbent-

fold nappes. Trümpy (1963) describes a different type of root zone related to some thrust nappes. In this second type, material of the nappe has been ejected between the jaws of two crystalline massifs. The nappe may move away from the root zone so that it has no connecting units and may consist entirely of higher parts of the stratigraphic section no longer represented in the root zone. This concept is called the "synclinal" origin of nappes, as opposed to the "anticlinal" origin described above. The root zone in the "synclinal" origin would be a strongly compressed zone containing remnants of rock units mainly older than those of the nappe.

Root zones consequently have no uniform geometrical or deformational characteristics, and their recognition depends on stratigraphic and structural evidence that identifies the zone where the nappes have their roots. Post-nappe deformation may make recognition extremely difficult.

It should be emphasized that all structures change character along strike. Some aspects of the changes in the Blue Ridge belt and Brevard zone have been discussed by Burchfiel and Livingston (1967), Livingston and Burchfiel (1969), and Dunn and others (1968). In the Alps, the displacement of some of the large nappes diminishes along strike, and far-traveled nappes may change into autochthonous sequences.

GEOLOGIC SECTION ACROSS THE ALPS

The relationships described here were observed along a section across the central Alps from Luzern to Locarno (Figures 1 and 2), a map distance of about 100 km. This section illustrates many of the principal Alpine structural elements.

Molasse Deposits

Sedimentary rocks of the Molasse formations (Oligocene-Miocene) fill the marginal depression (Swiss Plateau) north of the Alps (Trümpy, 1960, p. 880). The Molasse thickens and coarsens southward toward the Alps. These deposits form rugged mountains along the northern edge of the Alps where they are folded and overridden by the frontal Alpine nappes (Trümpy, 1960, p. 880). Near Luzern, overturned red-to-maroon conglomerates, sandstones, and shales of the Molasse are overridden by dark gray Mesozoic carbonate rocks of the Helvetic nappes.

Helvetic Nappes

The formations of the Helvetic nappes seen at lake-level south of Luzern are mainly sequences of apparently unmetamorphosed

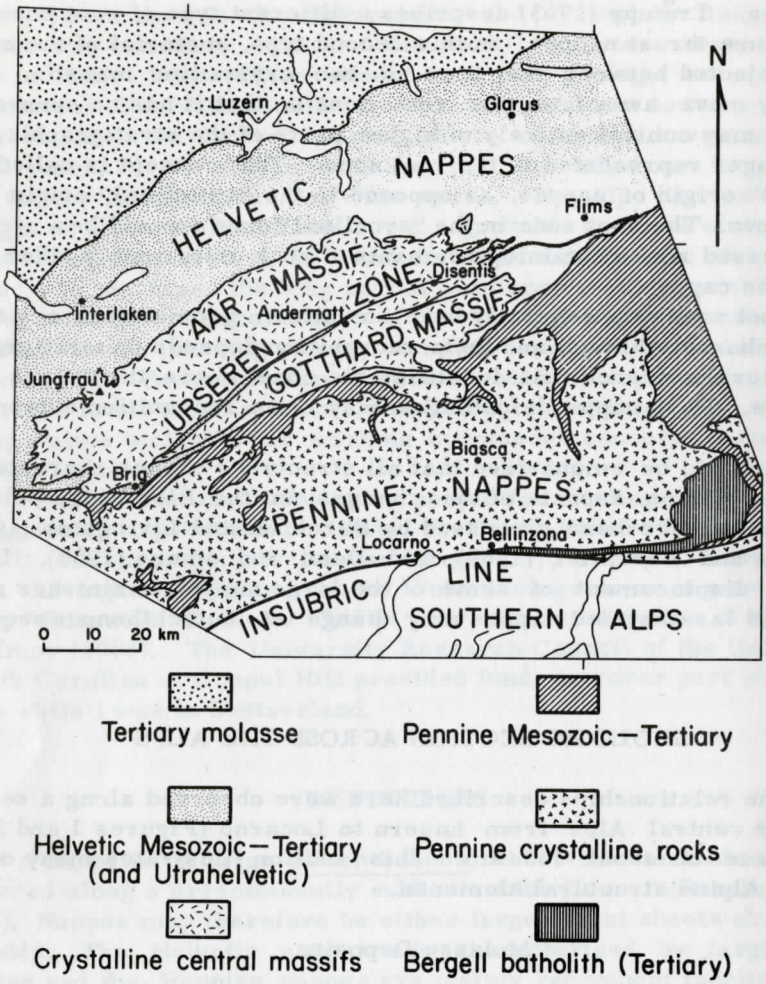

Figure 1. Generalized geologic map of the Central Alps. Modified after map from Lombard (1967).

Mesozoic carbonate rocks with relatively little lithologic contrast in outcrop. Some large-scale recumbent isoclinal folds hundreds of meters in amplitude are beautifully exposed. Locally, spectacular mountainside exposures show smaller-scale, strongly disharmonic recumbent folds, probably formed during the same episode as larger isoclinal folds. Published sections of the region (Lombard and others, 1962) show several superimposed nappes separated by large faults and internally disrupted by lesser order folds and faults. The interpretation (Trümpy, 1963) is that the Helvetic nappes are the sedimentary cover deposited on top and between the massifs to the south, and that the nappes moved northward during strong compression and relative uplift of the massifs.

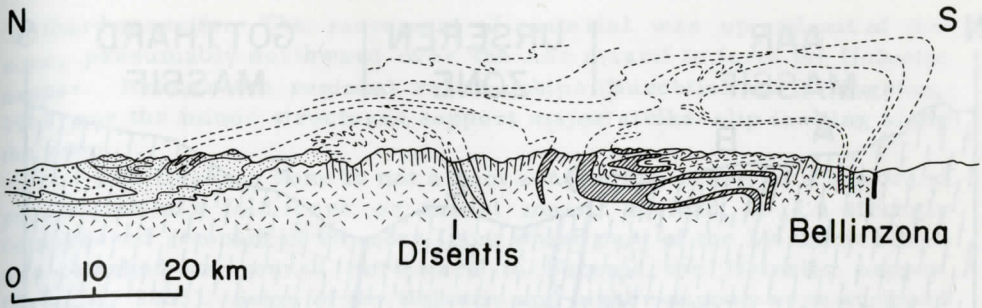

Figure 2. Schematic cross section of the Central Alps. Line of section is approximately from Bellinzona to Disentis to a point east of Luzern. For legend, see Figure 1. Modified after Heim (1922, Plate 27).

Aar Massif

South of the Helvetic nappes, one passes through the autochthonous sedimentary cover of the Aar massif and then into the crystalline rocks of the massif itself. The massif is Hercynian basement that was strongly deformed during the Alpine orogeny. The massif is cored by the Aar Granite and plunges beneath younger rocks and higher structural units to the northeast and southwest (Labhart, 1968, Plate 1). Mica gneiss and schist form the flanks of the massif.

The Aar Granite is a coarse-grained granitic gneiss of relatively uniform composition, with generally prominent foliation but little compositional layering. The Aar Granite may range from massive to schistose in a single outcrop. The foliation is generally nearly vertical and fans across the massif. A very prominent mineral lineation in strongly deformed gneiss plunges downdip in the foliation plane and is therefore generally about vertical. The lineation appears to be mineral elongation and streaking caused by movement in the foliation plane.

Urseren Zone

The Urseren zone is a long, narrow belt of Late Paleozoic and Mesozoic metasedimentary rocks between the Aar massif to the north and the Gotthard massif to the south (Figure 1). The general relationships are summarized by Burchfiel and Livingston (1967). The cross-section near Andermatt (Figure 3) shows that the Urseren zone is an asymmetrical zone of younger rocks (D and E) between older rocks of the massifs, with a zone of intense mylonitization (between C and D) near the northern boundary. The map by Staub (1932) shows the northern boundary as a large fault. Foliation and layering in the Urseren zone generally strike parallel to the zone and dip steeply to the south or southeast. Locally there are steep northerly dips. My field

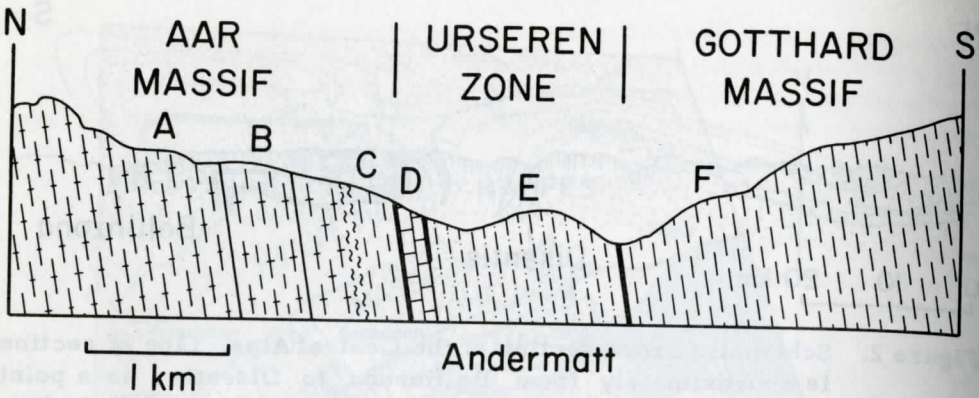

Figure 3. Generalized cross section of the Urseren zone near Andermatt, Switzerland. Modified after De Quervain (1967, Figure 2). A - Aar Granite, B - schistose Aar Granite, C - gneiss, locally intensively mylonitized, D - Mesozoic carbonates and schist, E - Permo-Carboniferous sericite gneiss and phyllite, F - gneiss and schist.

observations and thin-section study suggest that the Urseren zone and the adjacent parts of both massifs have no obvious differences in tectonic style or rank of metamorphism. Rocks of the Urseren zone are in the greenschist facies. Niggli and Niggli (1965) reported chloritoid from several localities along the Zone.

The Paleozoic rocks (Permo-Carboniferous; Fig. 3, zone E) are light greenish gray and range in grain size from very fine to coarse. A lineation defined by elongation of mineral grains plunges approximately downdip in the plane of foliation. Locally, the schistosity is deformed by crinkle folds and kink bands. The crinkle axes strike parallel to the zone and have shallow plunges. Crinkles are caused by a fracture or slip cleavage intersecting the earlier schistosity at an acute angle. The rocks probably represent a sequence of thin-bedded sandstones and shales metamorphosed to sericite and chlorite gneiss and schist. Zones of dark greenish gray rock are probably remnants of an ophiolite sequence of mafic and ultramafic igneous rocks.

Mesozoic rocks of the Urseren zone are mainly thin-bedded, dark-gray metamorphosed limestones interlayered with some dark schist. From vantage points along the zone, one can see the dark bands of limestone float trending for several kilometers parallel to the valley. The rocks in the Urseren Zone have been described as slices (Burchfiel and Livingston, 1967, p. 244), but it should be emphasized that the slices are very elongate and may preserve a consistent succession of units over distances of many kilometers.

The structures of the Urseren zone agree with the concept of intense penetrative deformation caused by squeezing between the Aar and

Gotthard massifs. The movement of material was up and out of the zone, presumably northward over the Aar massif to form the Helvetic nappes. Neither the regional relationships (Burchfiel and Livingston, 1967) nor the minor structures support major strike-slip faulting along the zone.

The Urseren zone is not a root zone in the sense of a backward penetration of a fold core or thrust mass. Instead, it is a strongly compressed remnant of the zone from which part of the Mesozoic cover was detached and moved northward to become the Helvetic nappes (Trümpy, 1963). Some of the Helvetic and higher nappes were stripped from the cover of the Aar, Tavetsch, and Gotthard massifs. The Helvetic nappes are therefore of "synclinal" origin (Trümpy, 1963).

Gotthard Massif

The Gotthard massif includes a variety of crystalline rocks metamorphosed and deformed during the Hercynian orogeny and then again during the Alpine orogeny (Lombard and others, 1962). Two large Hercynian (?) granitic bodies, the Rotondo Granite in the western part of the massif and the Cristallina Granite in the east, locally have discordant contacts(Lombard and others, 1962, p. 51). In some places, the Rotondo Granite is a well-foliated augen gneiss or granitic gneiss resembling the Aar Granite. Much of the schist and gneiss appears to be derived from sandy and pelitic sediments. Interlayered with the schist and gneiss is considerable amphibolite, particularly in the southern part of the massif. Some amphibolite layers in schist have been pulled apart into pods or boudins.

A north-south traverse across St. Gotthard Pass reveals a fan-like array of foliation. In the northern part of the massif, dip is steep to the south and in the southern part the dip is steep to the north. A mineral elongation is well developed in some places, and the lineation plunges approximately downdip.

Rank of metamorphism increases southward across the Gotthard massif. Staurolite and kyanite are found in the southern part of the massif (Niggli and Niggli, 1965), which is progressively metamorphosed to lower amphibolite facies.

Chadwick (1968) has recently described the complex relationships of rocks along the southern flank of the Gotthard massif, in an area about 15 km east of St. Gotthard Pass. The southern flank includes the Mesozoic cover of the massif and some schist-gneiss sequences of uncertain stratigraphic and structural affinity.

Pennine Nappes

The Pennine zone of the Alps is a gigantic pile of recumbent folds each cored by remobilized Hercynian basement rock. Evidence for the recumbent-fold structure is the occurrence of Mesozoic

metasedimentary rocks between the flat-lying cores of older gneisses. Locally, no sedimentary cover remains between the cores, and one gneiss unit rests directly on another (Lombard and others, 1962, p. 60). The bottom limb of the apparently structurally lowest nappe, the Leventina Gneiss mass, is not exposed and the allochthonous nature of the gneiss is questioned by some geologists.

The main highway south of St. Gotthard Pass passes through Leventina Gneiss for about 35 km. The main rock type exposed along the road is homogeneous, light-colored granitic gneiss with good foliation and very little compositional layering. The foliation generally has very shallow dips at road level and the surrounding mountains can be seen to have nearly horizontal structures. Cores of some of the higher nappes include darker-colored biotite gneiss and schist as well as granitic gneiss. The Mesozoic cover of the nappes has been strongly recrystallized to marbles, amphibolites, and rocks composed mainly of lime-silicate minerals.

Rocks of the Pennine nappes are mainly in the amphibolite facies. The northern part of the Pennine nappes east of St. Gotthard Pass is in the lower amphibolite facies (Chadwick, 1968) and kyanite occurs at many localities in the part of the Pennine zone discussed in this paper (Niggli and Niggli, 1965). Metamorphic rank increases southward, and sillimanite is found in and near the Pennine root zone.

Pennine Root Zone

As one approaches the Pennine root zone from the north, dip of foliation and layering changes abruptly from nearly flat to nearly vertical. Cross-sections of the Alps such as Figure 2 show this change, but I was unprepared for the spectacular nature of the change in the field. Along one traverse, there is more than ninety-degree variation in dip in a horizontal distance of about two km. The small isoclinal folds seen in the field and the regional relationships described in previous reports indicate that there is an abrupt bend in axial surfaces of the major nappes. Also, there is extreme attenuation and pinching out of units in the root zone. For example, the Adula nappe is several kilometers thick over a large area of the Pennine zone, but thins to about 100 m. in the root zone north of Bellinzona and may pinch out entirely in a westward direction (Lombard and others, 1962, p. 60). The root zone has been subdivided into a series of subsidiary zones by Swiss and Italian geologists. Some of the units can be traced into or otherwise correlated with specific nappes (Lombard and others, 1962, p. 49). Other sequences in the root zone do not appear to be related to any known nappes. There is controversy over the significance of some structures in the root zone and even over the root-zone concept itself (Lombard and others, 1962, p. 48).

In the root zone near Locarno, there is a wide variety of medium-to coarse-grained crystalline rocks, and the most common types

are biotite gneiss and schist and granite gneiss. Most of the rocks have strongly contrasted compositional layering, and schistosity is generally parallel to layering. Dip of layering is nearly always within twenty degrees of vertical, and in the southern part of the root zone layering dips steeply to the north. Locally, isoclinal and nearly isoclinal folds are spectacularly exposed. The folds have axial surfaces that are parallel to the general attitude of layering and schistosity. Some of the folds have wavelengths of at least twenty meters. The axes of the folds are nearly horizontal, with gentle plunges either to the east or west. A lineation is strongly developed in many outcrops and is approximately parallel to fold axes. The lineation appears as fine corrugations on foliation planes. Inasmuch as the lineation is generally parallel to the intersection of compositional layering and axial surfaces of the folds, it is probably related to development of the folds.

The presence of crests of isoclinal folds presents difficulties to the root zone concept, if the nappes are recumbent anticlines as in the Pennine region. Ideally, the flanks of the fold nappe can be identified in the root zone and the "crest" is far to the north. The root zone concept becomes increasingly untenable as more fold crests are discovered within the Pennine "root zone". My impression is that the root zone concept is generally accepted by Alpine geologists on the basis of regional relationships, but there is much difficulty in the interpretation of structures and sequences in the root zone itself.

The Pennine root zone is bounded on the south by the Insubric line (called the Tonale line along part of its length), a knife-sharp boundary that separates Alpine structures and metamorphism on the north from mainly Paleozoic structures and metamorphism on the south. The Insubric line connects eastward with the Giudicaria and Pusteria lines, so together these lines are the southern boundary of the regions most strongly affected by the main Alpine orogeny. Recently Gansser (1968) has discussed problems concerning the Insubric line in the vicinity of Locarno and Bellinzona. The sharp nature of the line is illustrated by K-Ar ages on hornblende (Gansser, 1968, p. 126). About 100 m. south of the line the apparent age is 326 million years, whereas 500 m. north of the line it is 26 m. y. All of the Pennine nappes must root either north of the line or at the line itself. Along the Insubric-Pusteria line to the east, the Middle East Alpine nappe must root in a narrow zone just north of the line, and the root zone of the huge Upper East Alpine nappe must have been largely engulfed along the line (Burchfiel and Livingston, 1967, p. 248). At a number of places along the tectonic lines, the regional interpretations require that the nappes root in a narrow zone north of the line or in rocks along the line that have disappeared, but the details of the geology and structural mechanisms are poorly understood.

To the north and south of the Insubric line, there are striking differences in structure, lithology, and metamorphic age of the rock formations, none of which have been correlated across the line (Gansser,

1968, p. 138). There are "many indications" of relative vertical movements of considerable extent (p. 139). The kinematic concept of the nappes involves movement of material mainly up and northward out of the root zone (Collet, 1927, p. 18-20), but the structural trends suggest a complex movement pattern. Folding occurred during the Alpine orogeny after the main northward transport of the Pennine nappes; Chadwick (1968) recognized two post-nappe phases of folding just south of the Gotthard massif. Post-nappe deformation may therefore have obscured some earlier relationships in the root zone.

The Insubric line is a major tectonic feature with a very long and very complex history that changes character along strike. There is evidence for some late-stage horizontal displacement along the Insubric line, but the amount if unknown (Gansser, 1968, p. 139). Lithologic changes across the Insubric line and related features to the east indicate structural influence during the Late Paleozoic and Mesozoic (p. 139). The strong contrasts across the Insubric line diminish westward from Locarno (p. 137).

The Insubric line is apparently related in some way to magma generation during the latter stages of the Alpine orogeny. A number of post-nappe granitic intrusions were emplaced on both sides of the Insubric line in Late Oligocene time (Lombard and others, 1962, p. 46, 51). The Bergell batholith (Figure 1) discordantly cuts the Pennine nappes and their root zone. The intrusions are limited to the southern part of the Alps in and near the root zones, or to the region south of the Insubric-Giudicaria-Pusteria lines.

CRYSTALLINE APPALACHIANS IN NORTH CAROLINA

Blue Ridge Belt

Rocks of the Blue Ridge belt can be divided into the following three general sequences: (1) Granitic gneiss, biotite gneiss and associated rocks with ages greater than one billion years (Davis, Tilton, and Wetherill, 1962) to which the name Cranberry Gneiss has been widely applied; (2) Upper Precambrian metasedimentary rocks, locally more than 15,000 m. thick (King, 1964, Hadley, 1970), including the Ocoee Series (King, 1964), Grandfather Mountain Formation (Reed, 1964; Bryant, 1962), Mount Rogers Formation (Rankin, 1967, 1970), and Ashe Formation (Rankin, 1970) and (3) Lower Paleozoic quartzite, phyllite, and carbonate rocks, correlated with the Chilhowee Group and Shady Dolomite (Reed, 1964). Lower Paleozoic rocks of the latter sequence are almost entirely limited to the Tablerock thrust sheet in the Grandfather Mountain window, the Murphy synclinorium, and (possibly) narrow remnants along the Brevard zone. General distribution of the three sequences is shown by Willden and others (1968).

The major structural features in the region discussed here are

the Blue Ridge thrust sheet and the thrust sheets of the Grandfather Mountain window. The northwestern boundary of the crystalline Appalachians is located where Precambrian rocks of the Blue Ridge belt are thrust to the northwest over Paleozoic rocks that are unaffected or very slightly affected by regional metamorphism. Stratigraphic, structural, and the metamorphic discontinuities show that the Blue Ridge belt surrounding the Grandfather Mountain window is allochthonous and has moved northwestward at least 55 km (Bryant and Reed, 1970). Northeastward from the window, the displacement of the thrust sheet diminishes in a manner still poorly known, and the Blue Ridge anticlinorium north of Roanoke, Virginia becomes autochthonous, at least in some places (Burchfiel and Livingston, 1967, p. 251; Espenshade, 1970, p. 208).

At least two phases of folding are recognized in the Blue Ridge belt, including a phase during or after the emplacement of the Blue Ridge thrust sheet. Early isoclinal folds occur in all three lithologic sequences of the Blue Ridge thrust sheet (Bryant, 1962, p. 25; Reed, 1964, p. 37-38; Brobst, 1962, p. 12; Butler, unpublished data). In the Grandfather Mountain window, the early folds change northwestward from tight or isoclinal folds to more open folds (Bryant, 1962, p. 25; Reed, 1964, p. 35). The late phase of folding includes large-scale gentle folds in the northwestern part of the window and adjacent thrust sheet. Faults bordering the southwestern part of the window steepen and become vertical and possibly overturned as they approach the Brevard zone (Reed and Bryant, 1964, p. 1188); therefore, the late phase of folding may have been more intense in the southeastern part of the Blue Ridge belt. The late folding probably is Late Paleozoic, inasmuch as the thrusting may be Mississippian or younger (Reed, 1964, p. 43-44).

Regional metamorphism preceded emplacement of the Blue Ridge thrust sheet and there has been little intrusive igneous activity since Middle Paleozoic time. Most of the Blue Ridge thrust sheet is in the kyanite or sillimanite zone of regional metamorphism, while rocks of lower greenschist facies (biotite-albite zone) are found in the Grandfather Mountain window (Bryant, 1962; Reed, 1964; Carpenter, 1970). The only post-metamorphic igneous plutons of significant size in the Blue Ridge belt are pegmatite and alaskite bodies that are particularly common in the Spruce Pine district, and the Stone Mountain granitic batholith north of Wilkesboro (Espenshade, 1967).

The region where the root zone of the Blue Ridge thrust sheet has been defined best is along the southeastern side of the Grandfather window (Figures 4, 5, and 6). Here the Linville Falls fault, which is the lower boundary of the thrust sheet, and the Brevard zone are less than two km apart for a distance of over 56 km (Reed and Bryant, 1964, p. 1188). The Linville Falls fault in this area probably has a southeast dip that averages 55 degrees. The gneiss in the strip is strongly sheared and retrogressively metamorphosed, but relict grains of kyanite are

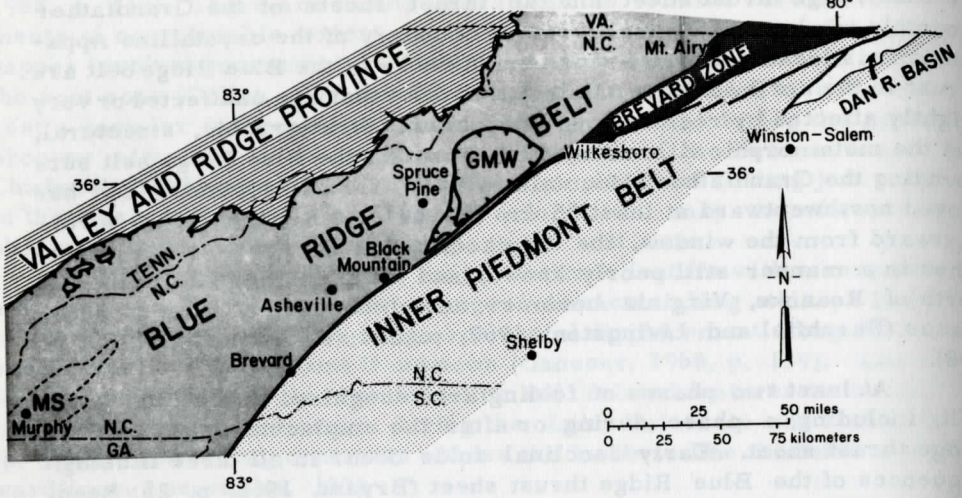

Figure 4. Geologic belts in western North Carolina and easternmost Tennessee. Modified after Willden and others (1968), Butler and Dunn (1968), and earlier references. GMW - Grandfather Mountain window, MS - Murphy synclinorium.

found in some specimens (Reed, 1964, p. 23). The Blue Ridge thrust sheet must root in this strip or somewhere to the southeast, as suggested by Burchfiel and Livingston (1967, p. 251). If the root is to the southeast, it has been overridden or destroyed. Rocks of the Brevard zone adjacent to the strip are very strongly sheared. Least-altered tectonic slices in the Brevard zone either resemble Inner Piedmont rocks or have no known counterparts in the Blue Ridge thrust sheet adjacent to the Brevard zone (Reed, 1964, p. 23-24).

The Blue Ridge thrust sheet is the highest tectonic unit yet recognized in the Appalachians of North Carolina and adjacent parts of Tennessee (Bryant and Reed, 1970, p. 219). Thrust sheets of the Valley and Ridge belt and Unaka belt lie tectonically beneath the Blue Ridge thrust sheet, therefore it is "at the top of the stack and probably originated farthest southeast" (Rankin, 1970, p. 242). Inasmuch as the Blue Ridge thrust sheet is apparently rooted just northwest of the Brevard zone, the Brevard could only be a root zone for structural units tectonically higher than the thrust sheet. No such higher units have yet been recognized. The emplacement of the Blue Ridge thrust sheet and attendant deformation of crystalline rocks probably provided the plunger that transmitted tangential forces to the sedimentary cover in the Valley and Ridge and Allegheny Plateau (Rodgers, 1964).

Brevard Zone

The Brevard zone is a narrow belt of cataclastic and other rocks

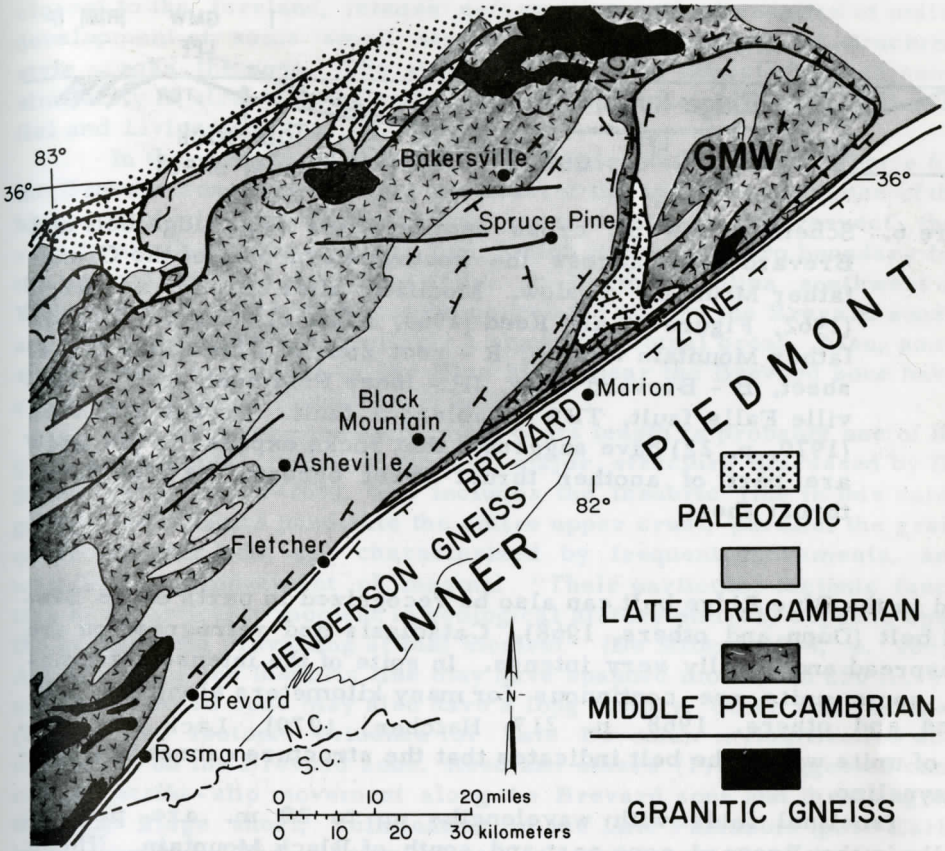

Figure 5. Generalized geologic map of the Blue Ridge belt and Brevard zone in part of western North Carolina and easternmost Tennessee. Modified after Willden and others, (1968) and earlier references, with some unpublished information by the author. GMW - Grandfather Mountain window.

probably more than 1,000 km long (Reed and Bryant, 1964; Burchfiel and Livingston, 1967). Northeast of the Grandfather Mountain window (Figure 4), the belt widens and changes character (Espenshade, 1967; Dunn and others, 1968). The following discussion applies to the segment southwest of Wilkesboro. Along most of the Brevard zone, schistosity and compositional layering dip at moderate angles to the southeast (Dunn and others, 1968, p. 217-218). Near Black Mountain, N. C. (Butler, in press), there are no strong differences in structural style across the zone or in adjacent belts, but locally truncation of map units affords good evidence for major faulting, particularly along the northwest boundary.

Rocks similar in lithology to parts of the two upper sequences

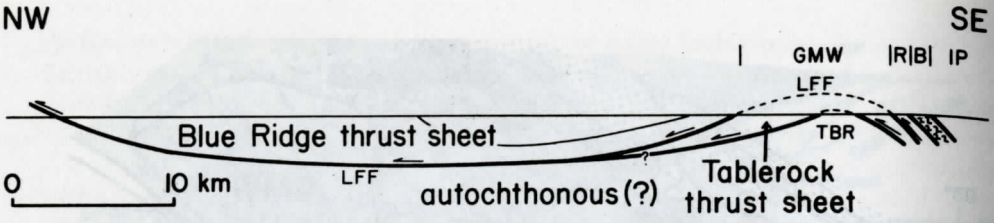

Figure 6. Schematic NW-SE cross section of the Blue Ridge belt and Brevard zone, across the southwestern part of the Grandfather Mountain window. Modified after Bryant and Reed (1962, Figure 3) and Reed (1964, Plate 2). GMW - Grandfather Mountain window, R - root zone of Blue Ridge thrust sheet, B - Brevard zone, IP - Inner Piedmont, LFF - Linville Falls fault, TBR - Tablerock fault. Bryant and Reed (1970, p. 22) have suggested that rocks exposed in the GMW are part of another thrust sheet beneath the Blue Ridge thrust sheet.

found in the Blue Ridge belt can also be recognized in parts of the Brevard belt (Dunn and others, 1968). Cataclasis and retrogression are widespread and locally very intense. In spite of the intense deformation, some units are continuous for many kilometers along the belt (Dunn and others, 1968, p. 217; Hatcher, 1970). Lack of repetition of units within the belt indicates that the structure cannot be a simple syncline.

Isoclinal folds with wavelengths up to 20 m. are present locally in the Brevard zone east and south of Black Mountain. These folds are tentatively correlated with the isoclinal phase of folding in the Blue Ridge belt.

There is some evidence for tectonic thinning and attenuation of units in the Brevard zone. In the vicinity of Wilkesboro (Figure 4) the zone narrows southwestward from 18 km to 2 km (Espenshade, 1967; Justus, 1971) and the cataclasis becomes progressively more intense. Two major units both show progressive thinning southwestward.

On a geologic-tectonic map (Willden and others, 1968) the nearly straight trace of the Brevard zone contrasts sharply with the much more irregular pattern of rock units and faults in the Blue Ridge belt. The Spruce Pine synclinorium is truncated along the northwestern boundary of the Brevard zone, and no equivalent units have been recognized southeast of the zone. These relationships suggest that the Brevard zone is a fault zone of large displacement and is related to a major linear feature that extended to considerable depth (through most or all of the crust).

The Brevard zone is similar to the Alpine Urseren zone in several respects: linear nature, continuity along strike of some units, asymmetric stratigraphy, concentration of faulting along the boundary

closest to the foreland, intense deformation and attenuation of units, development of some small structures, and continuity of structural style across the zone. The similarity of these aspects implies some similarity in structural roles for the two zones, as suggested by Burchfiel and Livingston (1967, p. 255).

In the vicinity of the Grandfather Mountain window (Figure 6), the Brevard zone lies directly southeast of the apparent root zone of the highest tectonic unit (Blue Ridge thrust sheet). In this respect, the zone is similar to the Insubric line, which is the southern boundary for the highest Pennine structural units. In North Carolina southwest of Wilkesboro, no rock units can be correlated across the Brevard zone, and the Insubric line is similarly a sharp structural break. Also, some small-scale structures in the Blue Ridge near the Brevard zone have counterparts in the Pennine root zone.

The Brevard zone along most of its length is probably one of the great fundamental faults, a type of major structure discussed by De Sitter (1964, p. 161-165), who includes the Insubric line in this category. These faults penetrate the entire upper crust, parallel the grain of the orogenic belt, are characterized by frequent movements, and may facilitate movement of magma. "Their particular tectonic function at a particular moment in geological history entirely depends upon the stress field prevailing at that moment" (De Sitter, 1964, p. 165). Activity along the Insubric line may have spanned more than 250 m. y. and the Brevard zone may also have a long history. Odom and Fullagar (1970) cited isotopic evidence for Late Devonian mylonitization and movement on the Brevard zone. Reed and others (1970) suggested concurrent strike-slip movement along the Brevard zone and thrusting of the Blue Ridge sheet, culminating in the Late Paleozoic (post-Early Mississippian). The closing stages of movement along the Brevard may have been in Late Triassic time (Reed and others, 1970, p. 269). One of the segments of the western border fault of the Dan River Basin (Figure 4) appears to be a splay of the Brevard zone that was rejuvenated in Triassic time (Butler and Dunn, 1968, p. 24). The history of the Brevard zone may therefore extend over more than 150 million years.

Inner Piedmont and Other Belts

The Inner Piedmont belt is very poorly known, because detailed geologic mapping has been done only in scattered areas. Much of the belt in North Carolina is underlain by well-foliated biotite gneiss and schist of undetermined age. A major feature is the vast terrain of augen gneiss and associated granitic gneiss in the northwestern part of the belt (Geologic Map of N. C., 1958), which includes the Henderson Gneiss (mainly microcline augen gneiss). Most of the Inner Piedmont belt is in the sillimanite zone of regional metamorphism (Overstreet and Griffitts, 1955), but metamorphic rank decreases to greenschist facies in part of the belt north of Winston-Salem, N. C. (Butler and

Dunn, 1968). No large post-metamorphic plutons have been found in the belt in North Carolina.

In the Inner Piedmont southeast of the Grandfather Mountain window, rootless isoclinal folds are overturned to the northwest and there is evidence for at least two periods of deformation (Reed, 1964, p. 40). Farther northeast, Butler and Dunn (1968) postulated large-scale recumbent isoclinal folding, followed by a phase of more open folding.

In part of the Inner Piedmont of South Carolina, the tectonic style is overturned to recumbent isoclinal folds with axial surfaces having shallow to moderate dips to the southeast, and commonly with sheared fold limbs passing into tectonic slides (Griffin, 1969a, 1969b). A later phase of more open folding is again recognized. The cross sections show a shingled effect of successive folds stacked against one another. The general appearance is quite different from the Pennine nappes, where many of the nappe cores can be traced or projected into a narrow, tightly compressed root zone.

In the southeastern part of the Inner Piedmont, dips are generally steep. The major structural style in the Kings Mountain, Charlotte, and northwestern Carolina slate belts seems to be tight to isoclinal, upright folds (Butler, 1966). Across the slate belt, the folds grade into open, upright folds, then into folds overturned to the southeast (Stromquist and Conley, 1959; Conley and Bain, 1965).

SCALE OF FEATURES

The area of Switzerland is less than one-third the area of North Carolina, and the Swiss Alps cover a consideraly smaller area than the combined Blue Ridge-Brevard-Piedmont belts in North Carolina. Some of the major structures are of the same order of magnitude. The Blue Ridge thrust sheet in the region of Grandfather Mountain window moved at least 55 km relatively northwestward (Bryant and Reed, 1970, p. 218). If the entire Blue Ridge belt southwest of the Grandfather Mountain window (in the segment from Brevard to the Great Smoky Mountains) is a thrust sheet, then the sheet has moved perhaps 120 km. According to Collet (1927, p. 178), the Great St. Bernard nappe of the Pennine Alps is about 95 km wide from root zone to forward edge, and is as much as 15 km thick. In the Eastern Alps, the Upper East Alpine sheet is probably more than 170 km from root zone to forward edge (Oxburgh, 1968, Figure 9), although the relationships are somewhat controversial.

CONCLUSIONS

1. Root zones may exhibit a wide range of features and their

identification generally depends on the deciphering of very complex stratigraphic and structural problems.

2. The Brevard zone has similarities to the Urseren zone and Pennine root zone-Insubric line of the Alps that imply somewhat similar structural roles.

3. The root zone of the Blue Ridge thrust sheet is located just northwest of the Brevard zone, or it has been engulfed, overridden, or sheared off by movement concentrated along the Brevard zone.

4. The Brevard zone could only be a root zone for a tectonic unit higher than the Blue Ridge thrust sheet as currently identified. This higher unit has either been eroded away or is present but not recognized.

5. The Brevard zone is a great fundamental fault zone of long and complex history. It probably responded in different ways as the stress pattern changed during evolution of the Southern Appalachians.

6. Northeast of the Grandfather Mountain window in the vicinity of Wilkesboro and Mt. Airy, the Brevard zone changes character. An understanding of these changes would help clarify the nature and role of the Brevard zone.

7. Comparative tectonics is particularly useful when concepts developed in well-exposed and better-known regions can be applied to less fortunate situations. The applications of the root-zone concept from the Alps (Burchfiel and Livingston, 1967) and the concept of stockwork folding from the Greenland Caledonides (Griffin, 1969a, 1969b) are strongly stimulating the development of tectonic theories for evolution of the Southern Appalachian crystalline region.

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SEDIMENT PATTERNS, PHYSICAL CHARACTERS OF THE WATER MASS AND FORAMINIFERIDA DISTRIBUTION IN INDIAN RIVER BAY, COASTAL DELAWARE

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ABSTRACT

The physiographic setting, bottom sediment facies patterns, salinity, temperature, pH, Eh and organic content of the water mass and bottom sediments of the Indian River Estuary have been compared with the distribution of living benthonic Foraminiferida. Twelve genera, including 23 species of Foraminiferida, have been tentatively identified and separated into six geographically defined facies groups.

Indian River Bay and the lower reaches of Indian River, and several smaller associated creeks, comprise an estuary system approximately 12 miles long and 2 miles wide at its maximum. Variation in the physical characters of the water mass and bottom sediment occurs both in diurnal and tidal patterns. Daily variation of bottom water character was observed to be significantly greater at the western edge of tidal intrusion in Indian River tidal creek than in central Indian River Bay. Patterns of sediment distribution in the Indian River Estuary are complex. Bottom sediment type ranges from a dark gray mud in the center and west to a well sorted clean sand in the tidal delta area to the east.

No significant correlation was found between patterns of foraminiferal abundance and the distribution of sediment types. Furthermore, with the exception of an Ammobaculites assemblage occurring in the tidal streams to the west, there is no significant association between species occurrence and the physical parameters observed. The species of benthonic Foraminiferida found living in the Indian River

estuary system appear to be tolerant of highly varied bottom sediment types and physical characteristics of the water mass.

INTRODUCTION

The distribution and abundance of species of Foraminiferida that inhabit Indian River Bay in Atlantic coastal Delaware have been studied in an attempt to relate occurrence patterns to physical aspects of the bottom sediment environments and of the overlying water mass. Comparisons are made with environmental parameters such as temperature, salinity, sediment particle size, oxidation-reduction potential, hydrogen-ion concentration and organic content of the bottom sediment. The coastal sedimentary environments in Delaware are located on the west flank of the Atlantic Coastal Plain-continental shelf geosyncline. At the present time, the coastline is one of inundation, with the Atlantic Ocean transgressing across the Coastal Plain. The Coastal Plain is characterized by large estuarine indentations such as the Delaware Bay and Chesapeake Bay. In addition, the shoreline area is marked by spits, baymouth bars, tidal deltas, lagoons and fringing Spartina marshes that border on and separate low lying headlands of Pleistocene sediment (Figure 1). Indian River Bay, an elongate coastal lagoon is one of a series of very shallow coastal lagoons cut off from the open Atlantic Ocean by baymouth bars and tidal deltas. The shape of the coastal lagoons is in general controlled by a deeply incised topography inherited from the pre-Holocene erosion surface. The Indian River lagoonal system is located in southern Delaware, roughly between latitudes 38°37' - 38°35.30'N and longitudes 75°11' - 75°06.20'W. Essentially, Indian River Bay is an elongate rectangle about six nautical miles long from Indian River Inlet on the east to High Grass Point on Piney Neck to the west. It is about two nautical miles wide in the east and narrows to about 0.8 nautical miles wide in the west. Tidal penetration continues another four nautical miles up river to a dam which is situated at about the natural upper limit of tidal flow.

Acknowledgments

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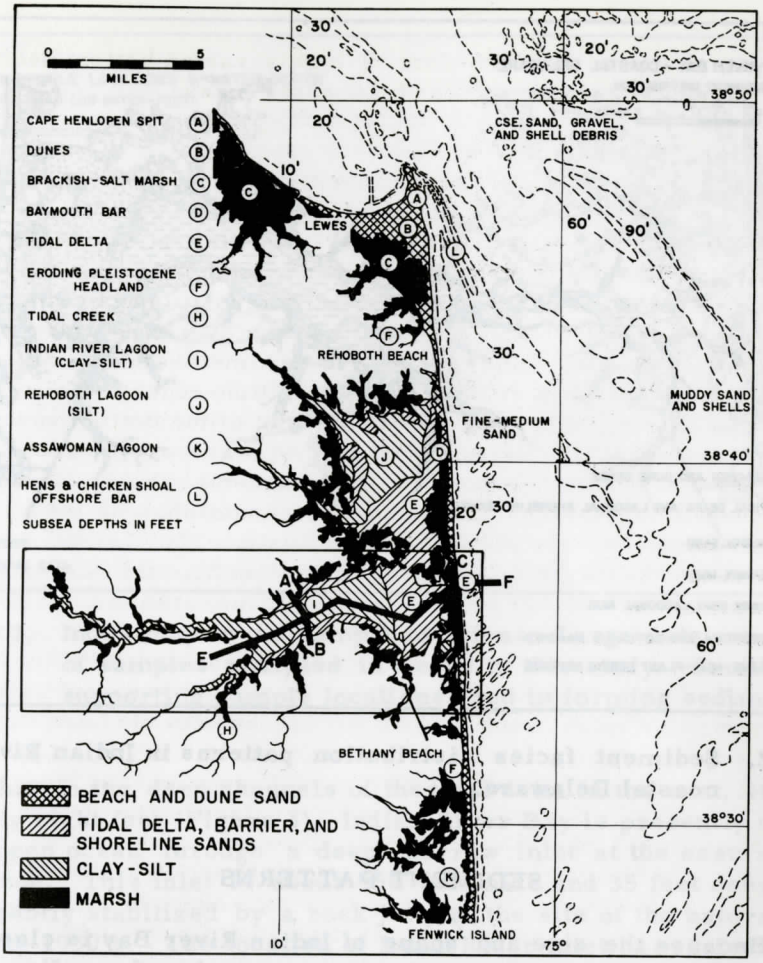

Figure 1. Recent sedimentary environments of the Delaware coast showing the leading edge of the Holocene transgression over the deeply incised and eroding Pleistocene headlands. Lines of section for interpretive cross-section figure 4 (A-B & C-D) for figure 5 (E-F) are shown.

interpretations herein presented. This paper is based in part of a paper presented by Kraft to the NE Section Geological Society of America-Society of Economic Paleontologist and Mineralogists annual meeting in Washington, 1968. A more thorough discussion of the microfauna and their environment tolerance limits may be found in an unpublished Master of Science Thesis by G. Margules, University of Delaware, 1968.

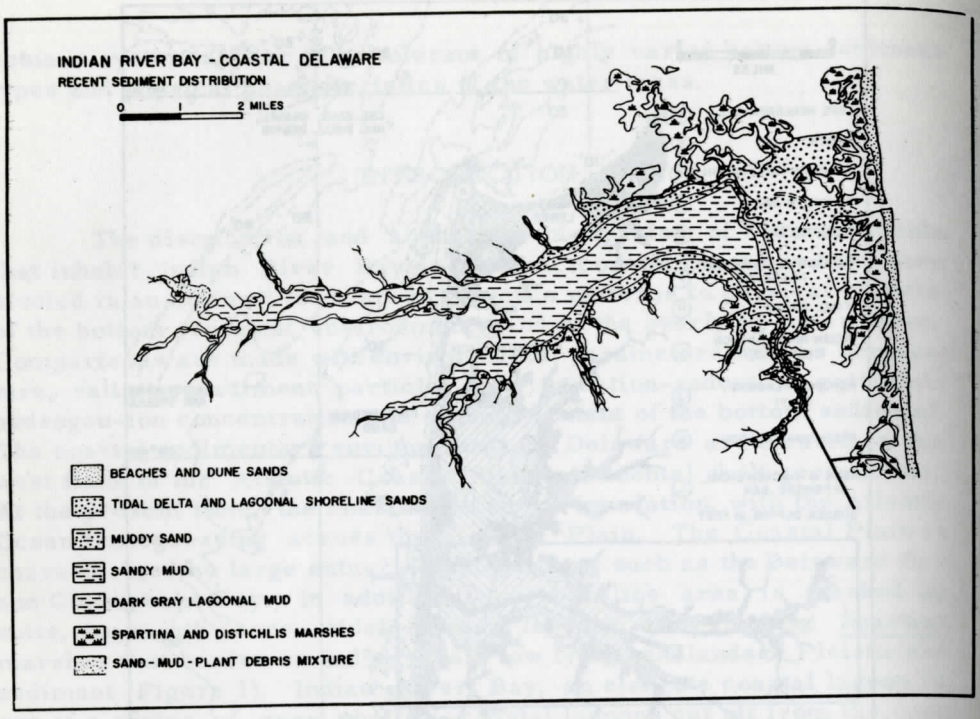

Figure 2. Sediment facies distribution patterns in Indian River Bay-coastal Delaware.

SEDIMENT PATTERNS

Because the size and shape of Indian River Bay is closely controlled by the underlying pre-Holocene topography, the sediment lithosomes within the Bay are formed of materials derived from the immediate surrounding hills as well as transported into the area by fluvial and tidal processes. The sediment patterns in Indian River Bay are as follows: (1) The transgressing beach and dune of the Atlantic coast; (2) A tidal delta forming within the eastern end of the bay through Indian River Inlet with its accompanying smaller tidal delta in the Atlantic Ocean; (3) A peripheral bay-bottom mixture of shoreline sediments comprising partially lagoonal clay-silts and derived Pleistocene sands, gravels and clay-silts; (4) A bay bottom layer of dark gray lagoonal mud typical of many of the smaller coastal lagoons of the Delaware coastal area; (5) A mud and peat formed in the Spartina-Distichlis-Phragmites marshes that surround the lagoonal area and define the limits of the ongoing Holocene marine transgression (Figure 2). Sediments within the marshes vary from silty and sandy organic muds to slightly organic sands in the baymouth bar area. An exceptional type of sediment occurs at the head of tidal intrusion on Indian River and is comprised of a sand, mud and plant-debris mixture. The maximum depth in the lagoon,

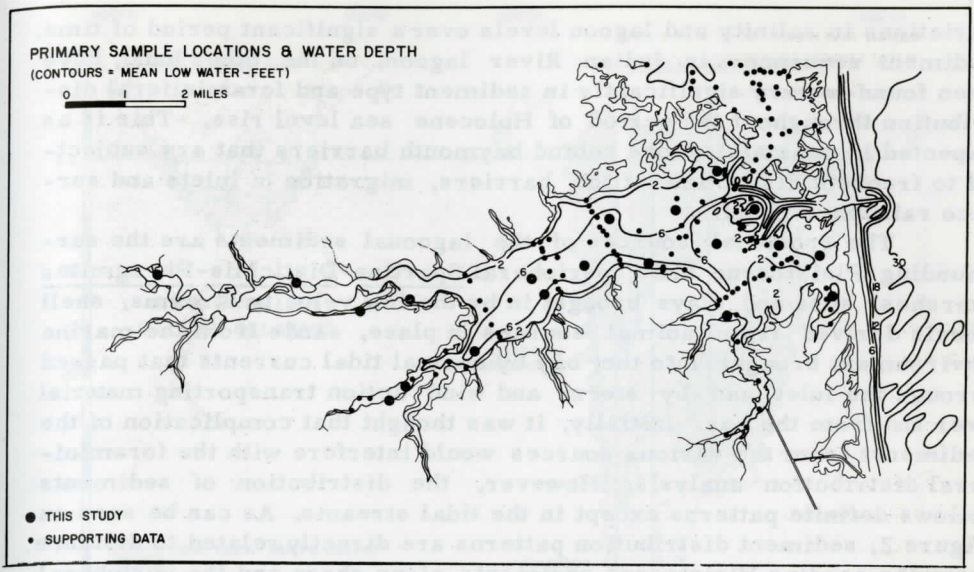

Figure 3. Index map to the sample locations. Large dots are locations of samples analyzed in detail in this study. Small dots are supporting sample locations used in forming sediment maps.

other than in the deep channels of the tidal delta in the east, is approximately eight feet (Figure 3). Indian River Bay is presently connected to the open ocean through a deep, narrow inlet at the eastern edge of the lagoon. This inlet is about 450 feet wide and 35 feet deep and was permanently stabilized by a rock jetty at the site of the natural inlet in the early 1930's. The location of Indian River Inlet has varied considerably over the past several hundred years, and has migrated north and south over a 4 to 5 mile stretch of barrier coast during the last 200 years. The stabilization of Indian River inlet by the U. S. Army Corps of Engineers jetty resulted in comparative subsequent stability of sediment movement and perhaps of physical characteristics of the water mass of Indian River Bay. Another result of the deepening and stabilization of the inlet has been the formation of a large tidal delta in the eastern part of the bay. This delta is cut by deep channels (18-20 ft.) in which there are exceptionally strong tidal currents. Observations by the authors and historical data over the past five decades have shown that salinity variations within the Indian River Bay are significant and dependent upon the amount of rainfall and other seasonal factors. Prior to the stabilization of Indian River Inlet, the inlets of the Delaware Coast would frequently seal up by action of littoral drift. Accordingly, at some times the bodies of water behind the baymouth barriers would become brackish and have at times been reported to be fresh water with levels several feet above the tides of the Atlantic Ocean (Howell, 1931). However, historical data is not precise enough to, at present, map

variations in salinity and lagoon levels over a significant period of time. Sediment sequences in Indian River lagoon, on the other hand, have been found to vary significantly in sediment type and foraminiferal distribution throughout the period of Holocene sea level rise. This is as expected in coastal lagoons behind baymouth barriers that are subjected to frequent disruption of the barriers, migration of inlets and surface rainwater runoff.

The erosional sources of the lagoonal sediments are the surrounding Pleistocene hills, peripheral Spartina-Distichlis-Phragmites marshes, silts and clays brought in by the low velocity streams, shell debris derived from animal sources in place, sands from the marine environment brought into the bay by normal tidal currents that passed through the inlet and by storm and wind action transporting material over and from the bar. Initially, it was thought that complication of the sediments from the various sources would interfere with the foraminiferal distribution analysis. However, the distribution of sediments follows definite patterns except in the tidal streams. As can be seen on Figure 2, sediment distribution patterns are directly related to distance from the eroding Pleistocene sediments along shore and the peripheral marshes. The washover fans and tidal delta forming in the eastern part of the coastal lagoon are essentially part of the baymouth barrier complex. All of the foraminifers listed in this study have been identified as recent benthonic types characteristic of marshes and coastal lagoonal waters.

A peculiar exception in sediment type occurs in the west end of Indian River Bay. At that place, microscopic particles including glass spherules and pieces of ash have been incorporated into the sediments. The glassy debris is derived from a power plant located on Indian River. The occurrence of these microscopic glassy spherules is of particular interest to this study in that certain species of agglutinate foraminifers such as Ammoscalaria sp. actively use the glass spherules as part of their tests.

Figure 3 is an index map to the Indian River Bay area of study. Sample locations for the construction of the bottom sediment map and for study of the geologic history of the Indian River Lagoon are shown as small dot profile lines crossing the lagoon. The larger dots in Figure 3 show the specific sites of detailed foraminiferal population analysis as reported in this study.

Indian River Lagoon is at the leading edge of the ongoing Holocene transgression caused in part by coastal erosion and in part by late Holocene sea level rise. Continuing studies in the nearshore marine area clearly show that the present transgression of the sea is occurring by a process of submerged beach face "scarp retreat". This erosion has resulted in the formation of a nearshore marine sand and gravel lag deposit overlying truncated remanents of Holocene coastal sedimentary environments. Sedimentary environments are controlled in location in part by the shape of the laterally and underlying pre-Holocene

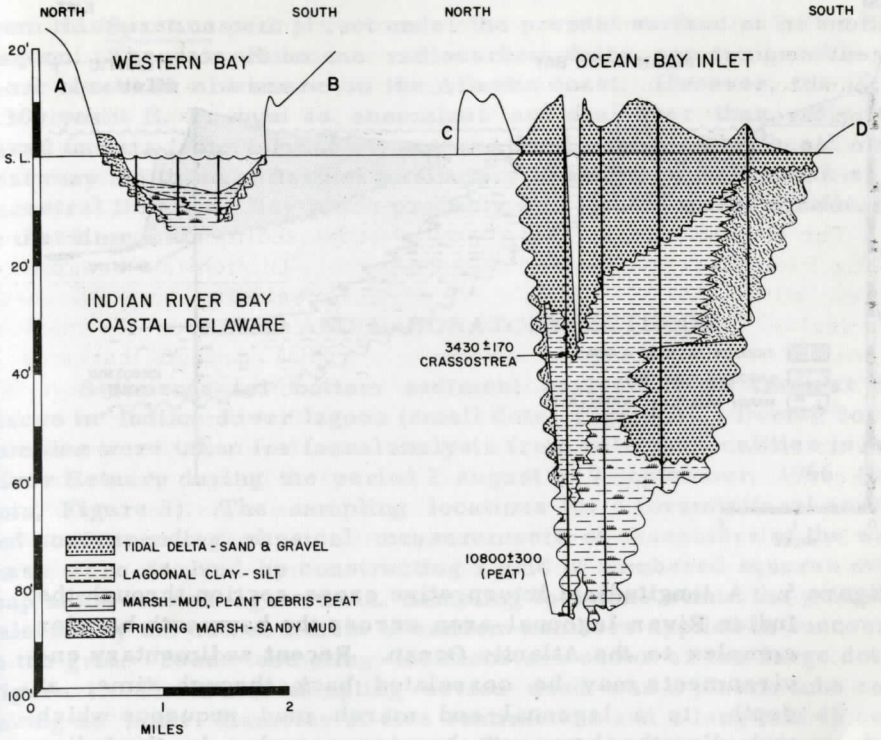

Figure 4. Interpretive cross-sections of the Holocene sediment sequence in western Indian River lagoon and beneath the baymouth barrier sequence between Indian River lagoon and the Atlantic Ocean. Interpreations from drill hole data indicate a deeply incised valley in pre-Holocene time. This valley has been infilled by coastal environments as the Holocene transgression advanced across the coastal plain to its location. The radiocarbon date at 10,800 ± 300 years before present is from a basal marsh peat located at the pre-Holocene unconformity. The 3430 ± 170 years before present date is based on several pairs of Crassostrea virginica shells in growth position within a lagoonal clay-silt sedimentary sequence. The present depth of Indian River Inlet is artificially maintained by the U. S. Army Corps of Engineers' protective jetties.

surface incised on Pleistocene coastal sediments. Figure 4 shows a cross-sectional interpretation of Indian River Lagoon at the ocean-bay inlet area and in the western part of the bay approximately six miles west of the Atlantic Ocean shoreline. As can be seen from Figure 4, the present lagoonal and marsh environments that occur surrounding and in the bottom of Indian River Bay are at deeper levels in the cross

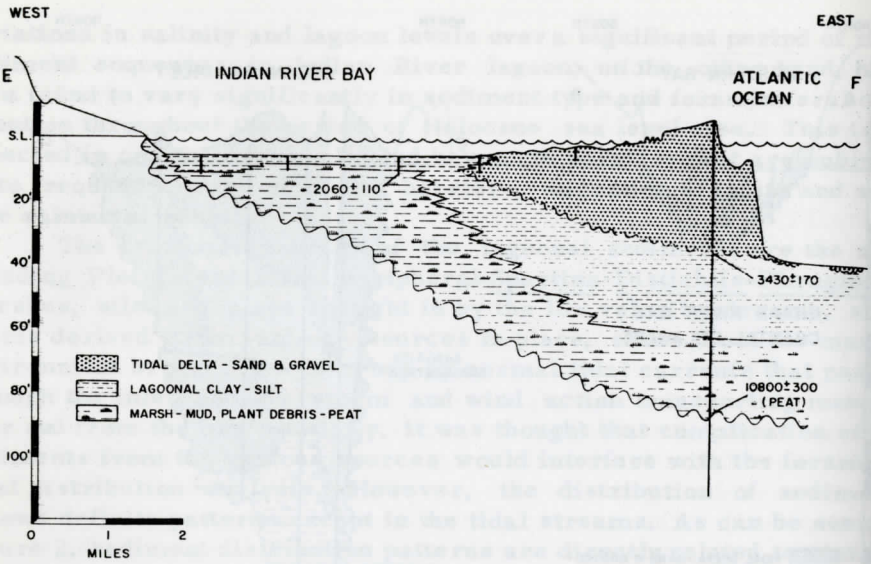

Figure 5. A longitudinal interpretive cross-section through the Indian River lagoonal area across the baymouth barrier complex to the Atlantic Ocean. Recent sedimentary environments may be correlated back through time, at depth, to a lagoonal and marsh mud sequence which underlies the baymouth barrier complex in the Indian River Inlet area. Figure 5 shows a cross-section of the Holocene transgressive advance in the Indian River area over the past 10, 000 + years.

section in mid and late Holocene time. Several radiocarbon dates shown in Figure 4 provide a frame of reference for time involved in the westerly migration of the coastal lagoon. Figure 5 is a longitudinal cross section from western Indian River Bay to the Atlantic Ocean. A direct correlation of present lagoonal and marsh environments with those of similar transgressing environments in the late Holocene is obtained from the log of a deep well at Indian River Inlet. Radiocarbon dates show that there was a simple mid-Holocene to present transgression marked by rising sea level. The earliest dates in the area of study were from a marsh grass and wood fragment peat found at the pre-Holocene unconformity at approximately 87 feet below present sea level and dated 10, 800 ± 300 years B. P. Several pairs of Crassostrea virginica Gmelin in growth position in lagoonal mud at approximately 38 feet below present sea level, underlying the baymouth barrier sequence at Indian River Inlet were dated 3430 ± 170 years B. P. A shallow core taken under Middle Indian River Lagoon penetrated a thin, lagoonal clay-silt equence and a fringing Spartina peat and mud sequence. A radiocarbon date of 2, 060 ± 110 years B. P. was obtained

from this Spartina peat 11 feet under the present surface of Indian River Lagoon. The late Holocene radiocarbon dates are comparable with those observed elsewhere on the Atlantic coast. However, the 10,800 ± 300 years B. P. date is anomalous and shallower than anticipated based on data from elsewhere on the Atlantic shelf. This basal marsh peat may indicate the head of a tributary to a large estuary such as the ancestral Delaware Bay which probably was located on the middle shelf at that time.

FIELD AND LABORATORY METHODS

Sediments for bottom sediment analysis were taken at many places in Indian River lagoon (small dots, Figure 3). Twenty bottom samples were taken for faunal analysis from selected localities in Indian River Estuary during the period 2 August to 9 September, 1966 (large dots, Figure 3). The sampling locations for a foraminiferal analysis and corresponding physical measurements of characters of the water mass were derived by constructing a grid of numbered squares over a map of the coastal lagoon area. Sampling locations within the grid were selected by the use of a table of random numbers applied in succession to the grid. These sampling locations are shown as the large dots on Figure 3. The basic sampling device used was a plastic tube corer having an inner diameter of five centimeters and a length of 45 centimeters. The corer was set into a holder attached to a steel pipe fitted at one end with a circle of prongs to enclose the top of the corer. A rubber ball freely floating within the circle of prongs acted as a valve to close the corer when a sample was taken (R. Priddy, oral communication). The sample was obtained by plunging the corer 6 to 8 inches into the bottom sediments and withdrawing it. When the sample was brought aboard the boat, a tightly fitting cork was inserted into the bottom of the core to hold the sediment in the tube when the valve was removed. Water above the core was siphoned off to about 2 inches above the sediment. The sample was then set aside for about 10 minutes to allow the foraminifers to "recover" from their removal from the bottom. Undisturbed samples are desirable as the foraminifers tend to withdraw into their tests when they are disturbed and thus defeat efforts to apply protoplasmic stain to them. Twenty drops of 10 per cent chlorohydrate were added to the remaining water to narcotize the animals (Aves, 1958). After another interval of 10 minutes to allow the chlorohydrate to take effect, the top 2 inches of the core and the water above it were placed in a plastic bag. A 50 percent solution of ethyl alcohol was added to the sample which was then removed to the laboratory for further study.

Data for the physical parameters were collected at the sampling sites by various means. Temperature and salinity values were taken in vertical profiles at 1 ft. intervals with an electrodeless induction

salinometer made by Industrial Instruments Corporation. Values for pH and Eh (mv) were obtained from the surface water, bottom water, sediment at the water-sediment interface and sediment 2-4 centimeters under the interface. The instrument used to take these data was a portable Sargent pH meter. All Eh readings were made at 2 minutes, and the pH readings were made 5 minutes after the sample was taken aboard. Portions of the samples were used for organic content analysis. Larger amounts of bottom sediment were obtained at the sampling site with a modified Forster-Anchor dredge as desired. Extreme care was exercised to insure that salinity, temperature, and, in particular, the pH and Eh readings were made in the same manner at all recorded stations.

The samples taken for Foraminiferida population analyses were carefully treated in the field to insure that an equal bottom sediment sampling area was studied at each field location. The bottom sediment locations selected in Indian River Bay, as based on a random number procedure, included the carefully controlled taking of a sample of known volume. A washed portion of the sample was kept in 50 percent alcohol at all times. To distinguish the foraminifers that were alive at the time of collection a protoplasmic stain was applied. The Rose Bengal staining technique, as described by Walton (1952), was applied. Protoplasmic staining techniques, as described by Walton and modified by others, did not appear to the authors to be entirely satisfactory, particularly with the agglutinate forms, whose tests are roughly textured and appear opaque in reflected light. Possibly, the use of the stain in an alcohol medium may have had a negative effect on the staining technique. Stained protoplasm is not always visible through the tests. It is problematical whether the color appearing in the test results from the staining of foraminiferal protoplasm or that of other smaller organisms that have invaded the tests after death of the original occupant. Conclusive evidence for the adequacy of staining tests is not available; accordingly, considerable reservation remains about the accuracy of the herein reported results on this account.

PHYSICAL CHARACTERS OF THE WATERMASS AND BOTTOM SEDIMENTS

In addition to the data obtained for the Foraminiferida study, a detailed study was made of variations in the physical characters of the watermass over tidal and diurnal cycles to test the data for variation caused by tidal change and sunlight effect. Particular emphasis was placed on these effects in the mid-east Indian River Bay and at the western limit of tidal intrusion in Indian River approximately 12 miles west of the Atlantic coast. A close association was noted between temperature and salinity variations in the open central lagoonal area of mid-east

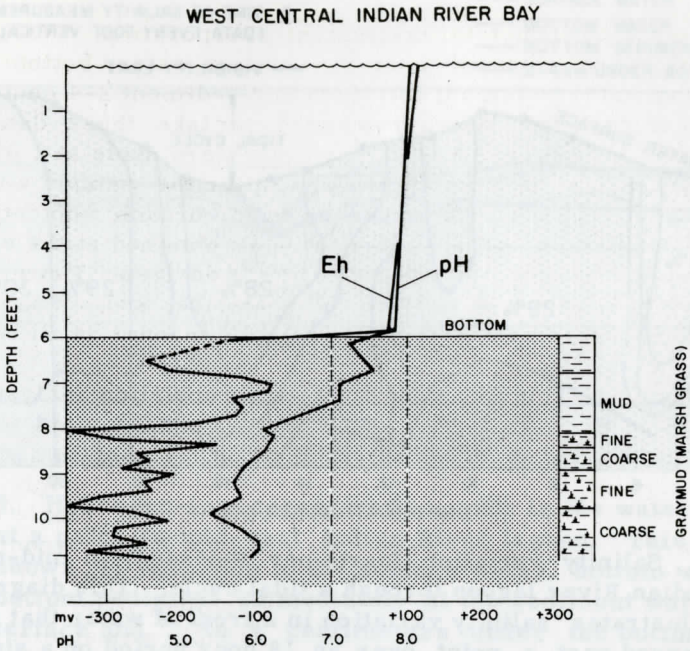

Figure 6. Oxidation reduction potential (Eh in mv) and hydrogen-ion concentration (pH) relationships in the water mass and Holocene sediments underlying west central Indian River lagoon. Note the strongly reducing and acid conditions predominant in the underlying lagoonal muds and Spartina marsh sequences.

Indian River Bay. Figures 7 and 8 show salinity and temperature variation in the water column through a tidal cycle in mid-July 1967. Daily tidal-cycle salinities varied from 30 0/00, which is almost ocean water in the Delaware coastal area, to 28 0/00, which is typical of the central bay area. Closely coinciding, temperature varied from 21 to 26° C with the colder water being associated with the tidal intrusion. It is interesting to note that the watermass circulates through Indian River Bay in vertical fronts rather than slanted wedges of denser water. This is evidenced by the fact that the readings upon which the Figures 7 and 8 are based include readings per foot of depth at each hour shown on the diagram. In other words, Figures 7 and 8 are based on over 100 readings. Figure 9 shows pH variation for surface water, bottom water, bottom sediment at the sediment water interface and the sediment 2-4 centimeters under the bottom for mid-east Indian River Bay. As may be seen from Figure 9, pH does not vary significantly with tidal intrusion or on a daily basis in the deeper (6 to 8 feet) mid-east Indian River Bay area. The anomalous reading at 11 A. M. has not been accounted

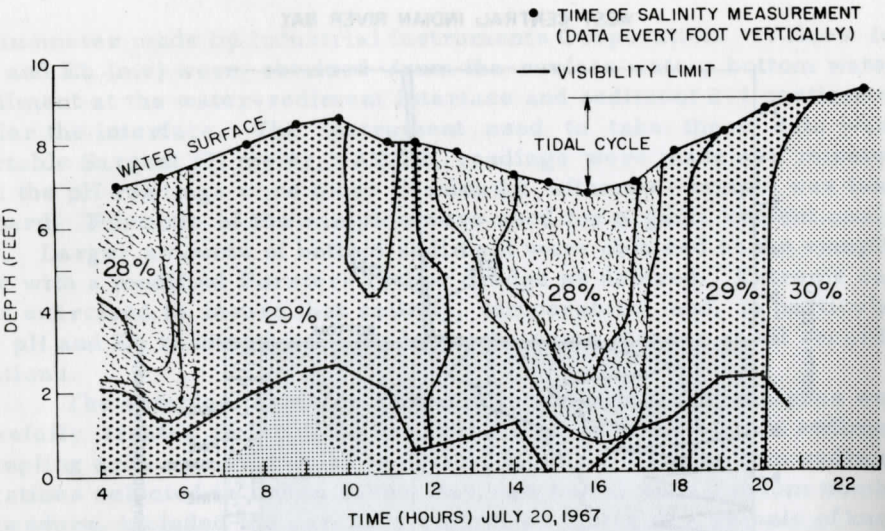

Figure 7. Salinity variation (parts per thousand) in mid-east Indian River lagoon through a tidal cycle. This diagram illustrates salinity variation in a front of water that advanced past a point over an 18 hour period on a single day. Visibility limit determined by use of Secchi disc. Salinity measurements every foot vertically. Depth of water variation through the tidal cycle is referred to the lagoon bottom.

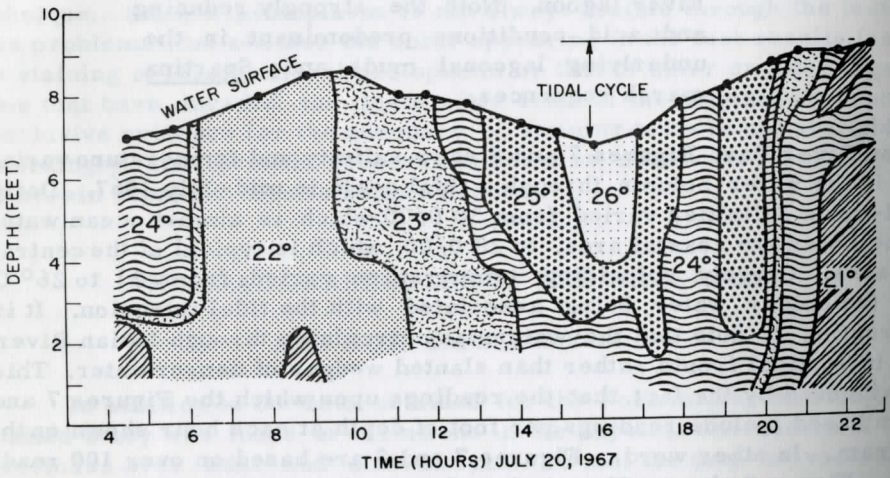

Figure 8. Temperature variation (degrees Centigrade) in the water mass at a location in mid-east Indian River lagoon through a tidal cycle over a period of approximately 18 hours on a single day. This diagram shows movement of the water mass past a single point through an 18 hour period. Note the essentially vertical front of the cooler water of the incoming tide. Depth of water variation through the tidal cycle is referred to the lagoon bottom.

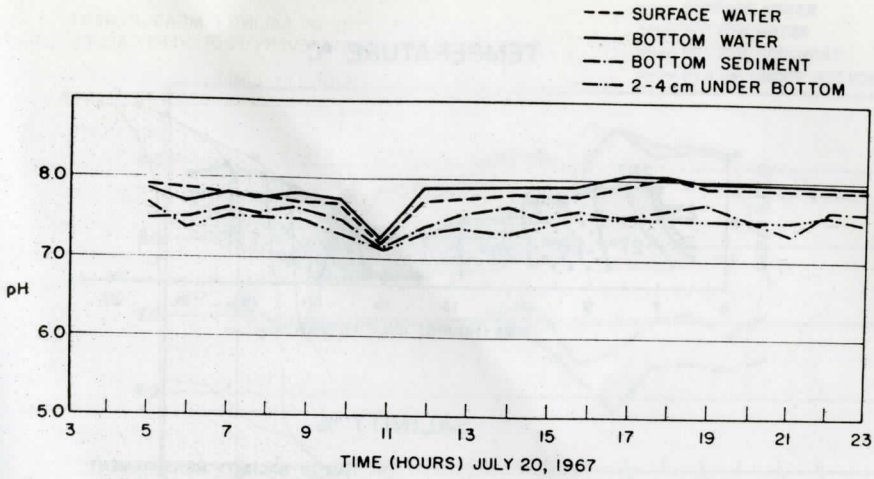

Figure 9. Hydrogen-ion concentration (pH) in the water mass at a point in mid-east Indian River lagoon. This graph shows pH variation in the surface water, bottom water, bottom sediment immediately at the sediment water interface and 2 to 4 centimeters under the bottom at a point in mid-east Indian River Bay over a period of approximately 18 hours on a single day.

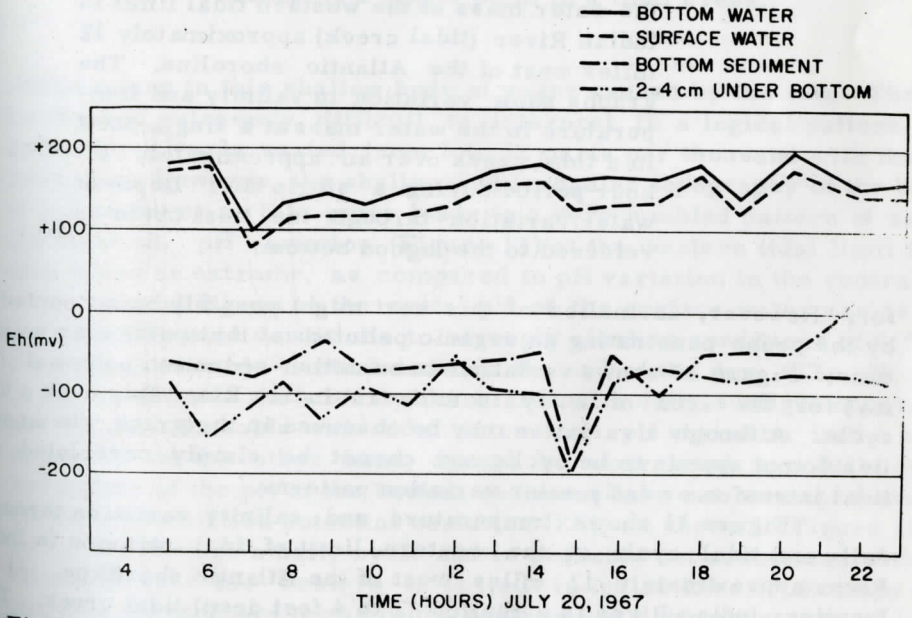

Figure 10. Oxidation reduction potential (Eh in mv) in the water mass at a point in mid-east Indian River lagoon. The graph shows Eh variation over an approximately 18 hour period on a single day.

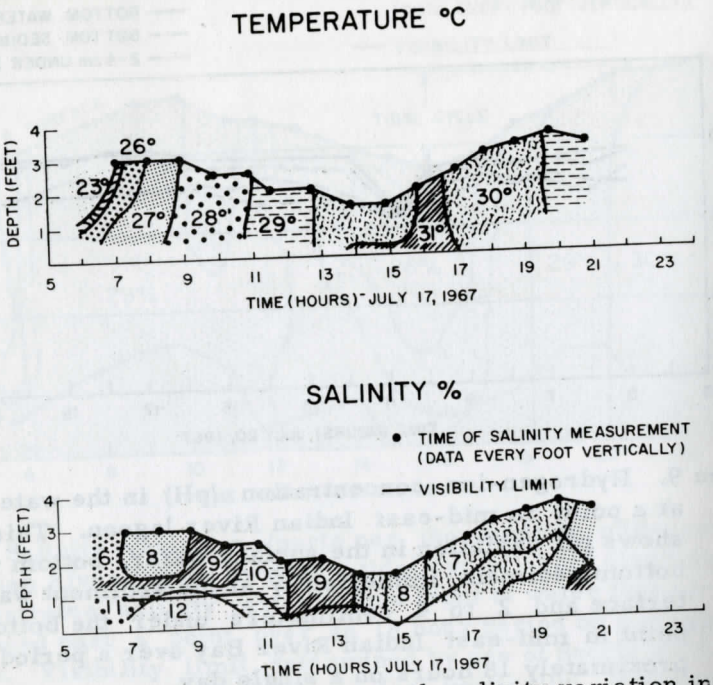

Figure 11. Temperature and salinity variation in the water mass at the western tidal limit in Indian River (tidal creek) approximately 12 miles west of the Atlantic shoreline. The graphs show variation in salinity and temperature in the water mass at a single point in a tidal creek over an approximately 15 hour period during a single day. Depth of water variation through the tidal cycle is referred to the lagoon bottom.

for. However, anomalies of this sort might possibly be accounted for by the probe penetrating an organic pollutant at that particular point in time. Figure 10 shows variation in oxidation-reduction potential (Eh-mv) for the area of study in mid-east Indian River Bay over a daily cycle. Although a variation may be observed in the graph, the anomalies do not appear to be cyclic and cannot be simply correlated with tidal intrusion or daily solar variation patterns.

Figure 11 shows temperature and salinity variation through a daily and tidal cycle at the western limit of tidal intrusion in Indian River approximately 12 miles west of the Atlantic shoreline. At this location, Indian River is a shallow (2 to 4 feet deep) tidal creek. Temperature variation there is much greater than that of the central lagoonal area to the east. Variations from 23 to 30°C can be in part associated with tidal intrusion but also are clearly associated with a rise in

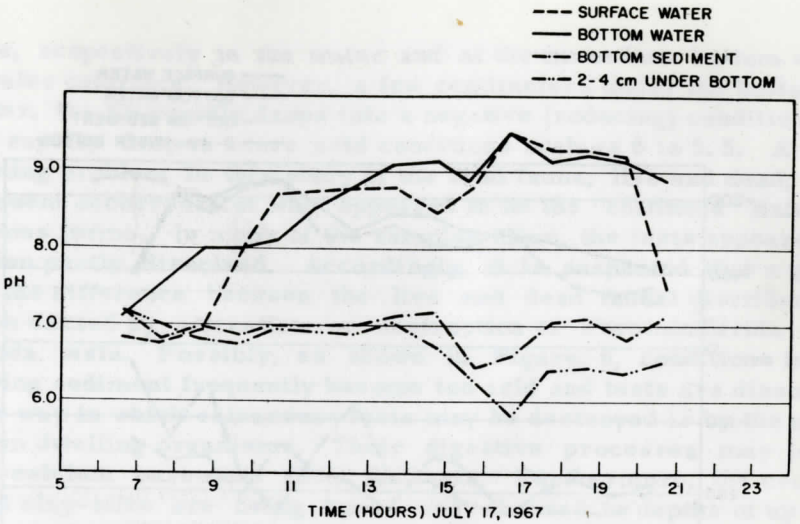

Figure 12. Hydrogen-ion concentration (pH) variation at the western tidal limit in Indian River (tidal creek). The graph shows variation in the pH over an approximately 15 hour period during a single day in surface water, bottom water, bottom water-sediment interface, and 2 to 4 centimeters under the bottom at a single point in a tidal creek.

temperature in this shallow body of water caused by the sun. The salinities are extremely difficult to interpret in a logical pattern. The salinities clearly varied from 6 to 12 parts per thousand with the tidal intrusion. However, the shallow and irregular topography of the bottom of Indian River at this point leads to a very jumbled pattern of salinity distribution. pH variation (Figure 12) at the western tidal limit in Indian River is extreme, as compared to pH variation in the central part of the coastal lagoon to the east. pH of the surface water varies from very slightly acid conditions to strongly alkaline conditions of 9.5 with a dropback towards neutral towards the end of the day. The pH cycles in the watermass are clearly associated with daily sunlight occurrence and not correlateable with tidal intrusion. pH variation in the bottom sediment, on the other hand, cannot be correlated to a daily cycle. Anomalies of the pH of the bottom sediment have not been accounted for. Oxidation-reduction potential variation (Eh), as shown in Figure 13, also varies through a daily cycle and is probably not associated with tidal intrusion. near the western tidal limit in Indian River. Steadily more positive Eh readings occur in the surface and bottom water until late in the afternoon, whereupon a sharp decrease in Eh of both surface and bottom waters occurs. Millivolt readings in the bottom sediment tended to rise through the day but were highly anomalous, erratic and lack a definite pattern.

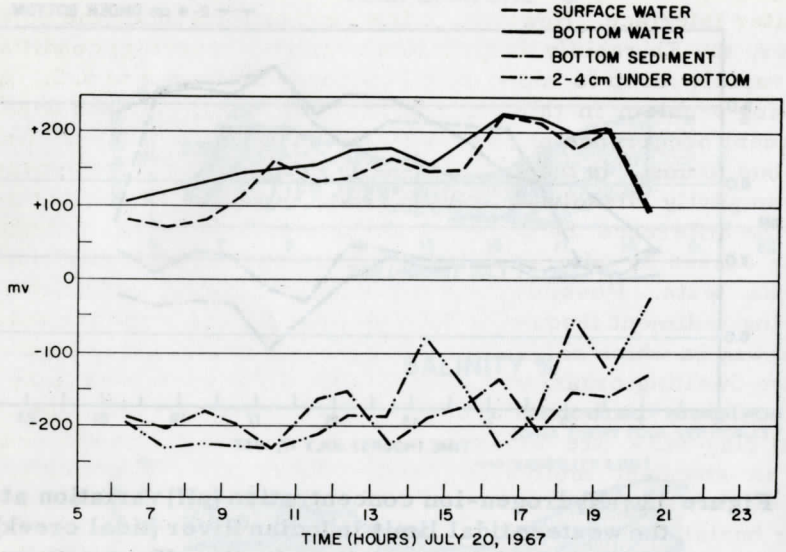

Figure 13. Oxidation-reduction potential (Eh in mv) at the western tidal limit in Indian River (tidal creek). This graph shows variation in Eh over an approximately 15 hour period during a single day in surface water, bottom water, bottom water-sediment interface, and 2 to 4 centimeters under the bottom of the tidal creek.

The physical characteristics of the watermass in the mid-lagoonal area as compared to the extreme limits of tidal intrusion in the tidal creek suggest that in certain areas of the lagoonal environment conditions are uniform throughout the day. In certain other areas of Indian River, highly variable physical conditions of the water mass occur. Accordingly, some of the foraminifers must be highly tolerant of varied physical conditions. It must also be realized that the variations observed in this study were observed only over a short period in the middle of the summer. No statistical analysis has been made of the long term effect. Obviously, critical studies should include much more data obtained over a longer period and covering climatic variation in addition to seasonal weather changes.

Figure 6 illustrates another type of problem encountered in the relationships of physical criteria to the distribution of Foraminiferida in coastal environments. Figure 6 shows the distribution of Eh and pH readings in the water column and in the bottom sediment to a depth of about 12 feet. A sharp change in Eh-pH condition occurs at the bottom sediment-water interface. Eh and pH are positive (oxidizing) and

alkaline, respectively in the water and at the immediate bottom sediment water interface. However, a few centimeters under the bottom of the water, the Eh rapidly drops into a negative (reducing) condition and the pH rapidly drop to more acid conditions such as 6 to 5.5. A most interesting problem in this study of the total fauna, live and dead, was the frequent occurrence of what appeared to be the "chitinous" lining of calcareous forms. In many of the samples taken, the tests appeared to have been partly dissolved. Accordingly, it is suspected that a great deal of the difference between the live and dead faunal distributions might be caused by alteration or destruction of Foraminiferida and Ostracoda tests. Possibly, as shown in Figure 6, conditions in the underlying sediment frequently become too acid and tests are dissolved. Another way in which calcareous tests may be destroyed is by the action of bottom dwelling organisms. Their digestive processes may leach out the calcium carbonate from the tests. Furthermore, the central lagoonal clay-silts are being bored and churned to depths of up to 4 feet by an abundant molluscan fauna. There are, accordingly, many evidences that suggest that foraminifer tests may be selectively destroyed after burial.

In some samples foraminifer specimens were obviously "fossil" or derived. Brown crushed and corroded tests formed small portions of some of the samples. These specimens were not included in this study and were eliminated as transported particles from elsewhere. Other authors have noted contaminant specimens in studies of recent sediment microfaunas (Kilenyi, 1969).

CORRELATIONS OF THE PHYSICAL PARAMETERS

This study limits the investigation of the relationships between the physical parameters and abundances of the benthonic Foraminiferida to the immediate environment (the water-sediment interface) and to a single season, the end of the summer. Furthermore, the majority of data used for direct comparison to the foraminiferal occurrences were taken from the midday portion of the daily cycle without the individual data being adjusted to the tidal variation. The bottom water is described in terms of salinity, temperature, hydrogen-ion concentration, oxidation-reduction potential and depth. The sediment is characterized by the sediment particle-size distribution expressed as the ratio between particles larger than 63 microns (sand) and particles smaller than 63 microns (silt and clay). The other characteristics considered are the hydrogen-ion activity (pH-sediment), oxidation-reduction potential (Eh-sediment) and estimates of the amounts of organic material in the total sediment and in the "finer" fraction of the sediment. The entire lagoonal system studied is rather small, approximately 10 by 2 nautical miles. The ranges of values for the physical parameters are small compared with the lagoonal-estuarine systems that are the subject of other

investigations (Ellison et al., 1965; Buzas, 1965; Phleger and Walton, 1950, and others). The water coming into the bay from the tidal streams is less saline, more alkaline, and warmer than that entering from the ocean. Continuum of values for salinity, temperature and pH-bottom water, is quite regular from west to east across the estuary. Correlations between depth and other physical parameters do not appear to be significant. All stations sampled are quite shallow and the difference between the maximum and minimum depths is only six feet. The midwest area of the bay is the deepest part of the system with the exception of the tidal inlet channels. The shallowest parts of the estuary are the western tidal streams and the tidal delta in the eastern part of the bay.

Both the bottom water and the interface sediment are slightly alkaline. No pH-bottom water value is less than 7.00 and only two stations have pH-sediment values of less than 7.00. The differences between the values for both pH-bottom water and pH-sediment are small. The constancy of the pH values may relate to the fact that observations were made through only the portion of the daily pH cycle in which the pH values tended toward their maximum values. Studies across a daily cycle have shown that the pH-bottom water tends to rise through the daylight hours towards a maximum in the late afternoon and the values drop later in the day. The values for pH-sediments also tend to rise during the day but far less sharply. Both pH-bottom water and pH-sediments tend to approach each other in the darker hours at about the level of the pH-sediment lower values. The differences between pH-bottom water and pH-sediment at any single station are greater in the tidal creeks to the west than in the eastern part of the bay. Since the observations were made between the hours of 5 A. M. and 11 P. M., it is not known if the pH values became even lower during the hours when no observations were made. Values reflecting the pH cycles of the individual stations may be more important than the values between the stations because of the effect of pH on the calcareous Foraminiferida. The pH cycle probably relates to photosynthetic and respiratory activity; however, these parameters were not measured. Oxidizing conditions tend to be prevalent in the eastern part of the bay where salinity is higher and where there is a sand substrate with little organic material. The Eh values also undergo a daily cycle in which Eh-bottom water tends to rise through the day and to drop off in the evening. Eh-sediment values also tend to rise through the day but do not decrease as sharply in the evening. Generally within the cycle, at any single time of observation, Eh-bottom water values tend to be greater when Eh-sediment values are higher.

The estimates of sediment organic content are assumed to relate to the amount of potential nutrient available to the Foraminiferida. The sediment organic content-fines was considered separately from sediment organic content-total on the supposition that very small particles of organic matter can constitute a sort of particulate food, if

particulate food of this nature is taken by the Foraminiferida. Actually, very little is known about foraminiferal nutrition (Loeblich and Tappan, 1964). Several investigators have noted that abundances of the Foraminiferida are greatest where the sediments contain considerable organic material (Phleger and Walton, 1950; Todd and Low, 1961; and others). Both sediment organic content-fines and sediment organic content-total follow about the same pattern of distribution with high values occurring in the upper reaches of the stream and mid-bay muds and low values in the sands of the tidal delta.

DISTRIBUTION OF THE FORAMINIFERIDA

The living Foraminiferida of the Atlantic Coast of the United States have been investigated both north and south of Delaware. Cushman (1944) described the shallow water species off the New England Coast. Phleger and Walton (1950) studied the marsh and bay fauna in Barnstable Bay, Massachusetts, where there is little variation in the salinity and temperature and suggested that the main factors determining distribution are organic productivity and movement of bottom material by tidal action. Said (1951) investigated Narragansett Bay and found that some species varied with salinity while others did not. The studies by Parker (1952) and Phleger (1952) dealt with distributions in the Gulf of Maine. Phleger suggested that there is a close association between the species and the type of sediment in which they are found. Parker and Athearn (1959) working in the Poponesset Bay area, found that species differed in their responses to salinity. Todd and Low (1961) investigated varous "microenvironments" (sea beaches, harbors, ponds, inlets, etc.) and from their findings suggested that abundant foraminifer populations are associated with marine vegetation and "organic rich" sediments. Few foraminifers were found in clean sand. Lidz (1965), working in Narragansett Bay, found maximal populations in fine grain sediments that are associated with high organic content in high reducing conditions. Murray (1968), working in Buzzards Bay, presented the results of a detailed analysis of samples and a review-statement of foraminiferal assemblages of lagoons and estuaries. Murray noted that extreme complications could arise from modifications of the death assemblage, which is essentially a collection of specimens assembled over a long period of time, with potential seasonal and long term variations and the superposition of posthumus alteration of the assemblage.

Investigations of the New York area include a description of species found in New York Harbor by Shupack (1934) and an ecological study by Ronai (1955). Ronai suggested that the controlling factors for the brackish water forms in the New York bight are salinity, sediment, organic content and the currents which effect the distribution of the sediments. Buzas (1965) found no relationship between foraminiferal

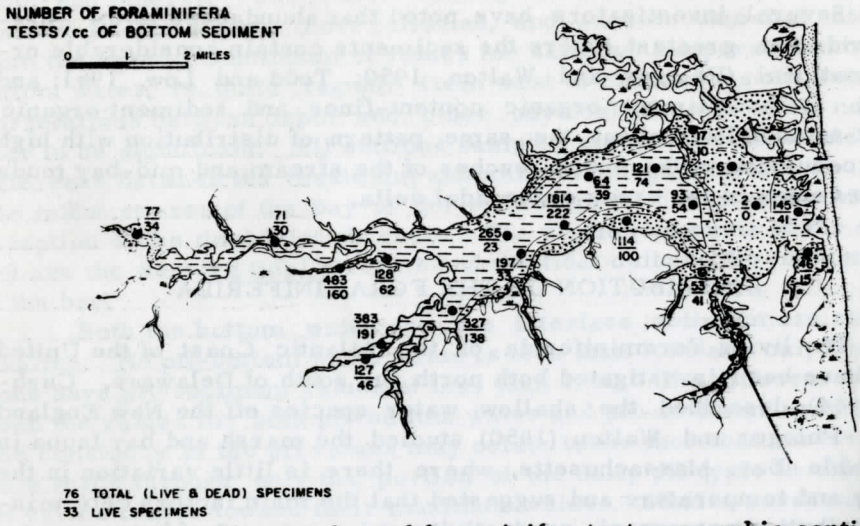

Figure 14. The number of foraminifer tests per cubic centimeter of bottom sediment at sampled locations in Indian River lagoon and associated tidal creeks. Sampling locations were picked by a process of applying a table of random numbers to a cross-sectional grid placed across the map of the lagoon.

species and particle size of the sediments, depth, temperature, salinity, pH and Eh in his study of the foraminiferal distributions of Long Island sound. Ellison, Nichols and Hughes (1965), working in the Rappahannock River Estuary, Virginia, found that foraminiferal populations were distributed according to depth and bottom type. In his study of Mason Inlet, North Carolina, Miller (1953) suggested that the foraminiferal populations were controlled by the substrate, but unlike the other investigators who suggested that sediment-type is the major determinant, he found that maximal populations occur in clean sands and proposed that the anaerobic conditions of organic clays made substrate unsuitable. Howard (1965) suggested that turbulence is significant in the control of the shallow water Foraminiferida at Big Piney Key, Florida, insofar as turbulence relates to turbidity and, hence, to plant production.

The number of foraminifer tests per cubic centimeter of bottom sediment was carefully calculated in the laboratory. Frequency of occurrence of foraminifers in the samples studied are presented in Figure 14. Foraminiferal populations vary from foraminifer "deserts" in the clean sand tidal delta area in the eastern part of Indian River Bay to foraminifer "jungles" in the central lagoonal clay-silts in mid-Indian River lagoonal area. Figure 14 contrasts the total (live and dead) specimens compared with "live" specimens. Live populations varied from 0 to 222 specimens per cubic centimeter in the Indian River Bay

sediments. Buzas (oral communication) noted that the Foraminiferida population varied sharply over a 100 centimeter square bottom sediment area in adjacent northeast Rehoboth Bay. Accordingly, great caution must be used in attempting to form meaningful interpretations from population numbers in coastal lagoons. However, the observations of relative abundance, noted herein, are partially confirmed by limited studies of supporting samples located on Figure 3.

Twenty-three species of Foraminiferida in 12 genera were identified in the Indian River lagoon. All of the species are benthonic shallow water forms. Only one species, Elphidium poeyanum (d'Orbigny) is represented by less than one percent of the fauna (assemblages of Foraminiferida) at all stations. Thirteen species are present in proportions of at least five percent at one or more sampling stations. Only nine sampling points have a "dominant" species, that is, a species that comprises 50 percent or more of the fauna. Of the remaining sampling stations, four have "major" species (comprising 25 percent to 40 percent of the fauna) and the rest are "minor" (five percent to 24 percent), "very minor" (one percent to four percent), or "present" (less than one percent). Five sampling stations have two major species characterizing the fauna with the rest appearing as minor, very minor or present. Two sampling stations have no major species; the fauna is made up of more species in lesser proportions. Species diversity for the Indian River Bay appears to be considerable.

Six Foraminiferida assemblages are recognized for the live and total (live and dead) assemblages. Each "assemblage" is defined as a lateral change in the Foraminiferida faunal character of the lagoon. The Foraminiferida assemblages are distinguished from each other by proportionate composition of the species occurring in them. Figure 15 lists the Foraminiferida assemblages (live and dead per upper five centimeters of bottom sediment) and shows their distribution throughout Indian River Bay. Figure 16 lists the Foraminiferida assemblages (live: bottom sediment-water interface) and shows their distribution throughout Indian River Bay.

Foraminiferida assemblage #1-Western Tidal Streams and Mid-Bay: Ammobaculites salsus. This assemblage occurs in the tidal streams to the west and a portion of the mid-western part of the main lagoonal area of Indian River Bay and is dominantly comprised of Ammobaculites salsus Cushman and Bronnimann. Figure 17 presents the Ammobaculites salsus assemblage in camera lucida drawings and shows relative abundance of each species within the assemblage. Numerous species occur in this assemblage, some are common; whereas, some are rare or only present as a trace.

Assemblage #2-Western Bay-Tidal Stream Mouth: Ammobaculites salsus-Elphidium incertum. The western bay-tidal stream mouth assemblage is comprised mainly of two dominant species; namely, Ammobaculites salsus and Elphidium incertum (Williamson).

Assemblage #3-Southeast Bay-Nearshore: Pseudoclavulina

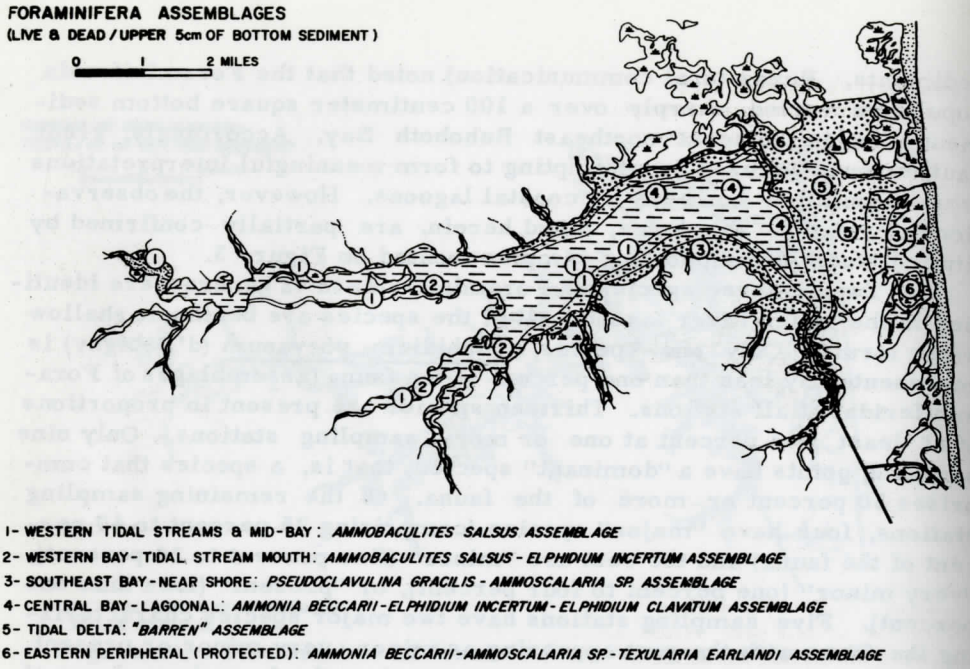

Figure 15. Foraminiferida assemblages of Indian River lagoon. Assemblages listed are based on an analysis of live and dead specimens in the upper 5 centimeters of bottom sediment sampled by a 2 inch diameter bottom sediment coring device. The foraminiferida assemblages are named after dominant species within the fauna.

gracilis-Ammoscalaria sp. The southeast bay-nearshore assemblage is characterized by an abundance of Pseudoclavulina gracilus Cushman and Bronnimann and Ammoscalaria species.

Assemblage #4 is illustrated in Figure 17. The Central Bay-Lagoonal assemblage is dominated by Ammonia beccarii (Linné), Elphidium clavatum Cushman and Elphidium incertum. Distribution of the assemblage #4 is restricted to the central lagoon in the live-dead foraminifer distribution map shown in Figure 15. However, three subassemblages can be identified from live specimens (Figure 16). These subassemblages are: 4(a) Elphidium incertum-Ammonia beccarii subassemblage which occurs in the organic clay-silts at the lower end of the tidal creeks; 4(b) Ammonia beccarii-Elphidium incertum subassemblage which occurs in the west central lagoonal clay-silts and at the mouth of another larger tidal creek (Indian River); and 4(c) Elphidium clavatum-Ammonia beccarii subassemblage which occurs in the east central lagoonal clay silts near the edge of the intruding tidal delta. Serious discrepancies may occur in the areal distribution of assemblage #4 because of postdepositional solution of the dominantly calcareous forms in this fauna.

Assemblage #5-Tidal Delta (Based on six specimens). The tidal delta "barren" assemblage is essentially a foraminifer desert. Here in the clean sands of the Indian River tidal delta inlet fan very few

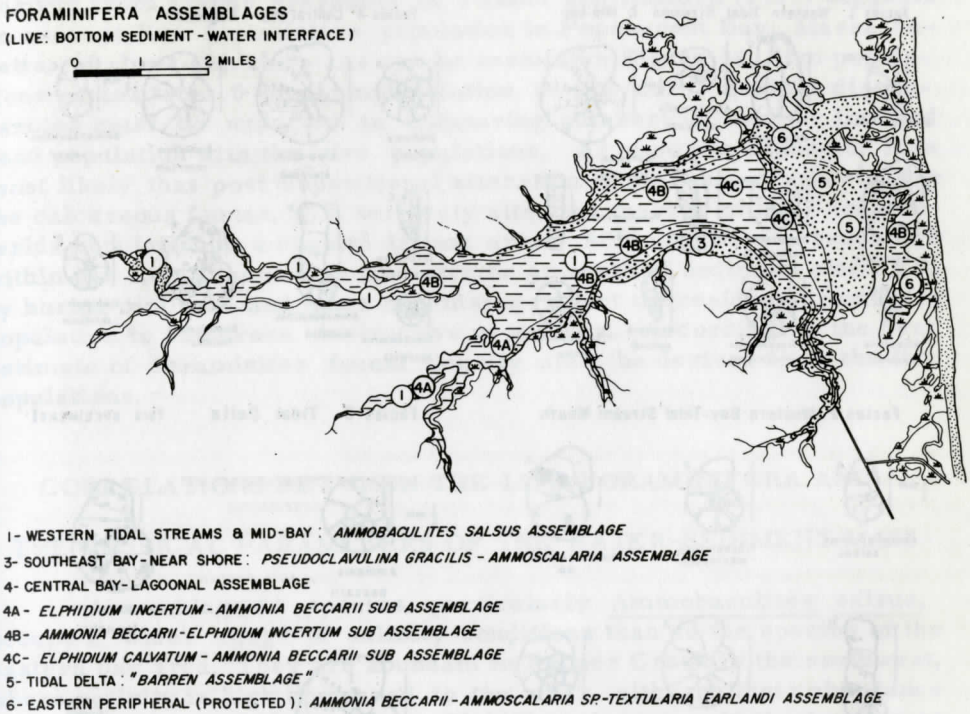

Figure 16. Foraminiferida assemblages of Indian River lagoon based on live specimens obtained from a 2 inch diameter bottom sampling device. Determination of living forams was by a standard rose Bengal staining technique.

foraminifers are found. This lack of foraminifers was noted in other sediment studies of the tidal delta by M. Field (oral communication). Accordingly, a percentage distribution of one species relative to the other is useless. In fact, because of the high tidal current energy in the area, the few foraminifer specimens found may have been washed into the environment from elsewhere. The "barren" assemblage is not, therefore, a true faunal assemblage; rather, it is listed here to complete coverage of the geographic setting of the lagoon.

Assemblage #6-Eastern Peripheral (protected). The *Ammonia beccarii*-*Ammoscalaria sp.* - *Textularia earlandi* assemblage is confined to protected shoreline areas of muddy sand in the eastern periphery of Indian River Bay.

The average number of live foraminifers per square centimeter in Indian River lagoon is rather high compared with other similar areas on the Atlantic Coast. Buzas (1965) reported an average of 17.7/cm^2 for the inshore areas of Long Island sound. Ellison *et al.* (1965) reported a maximum of 16.7/cm^2 in the Rappahannock. An estimate

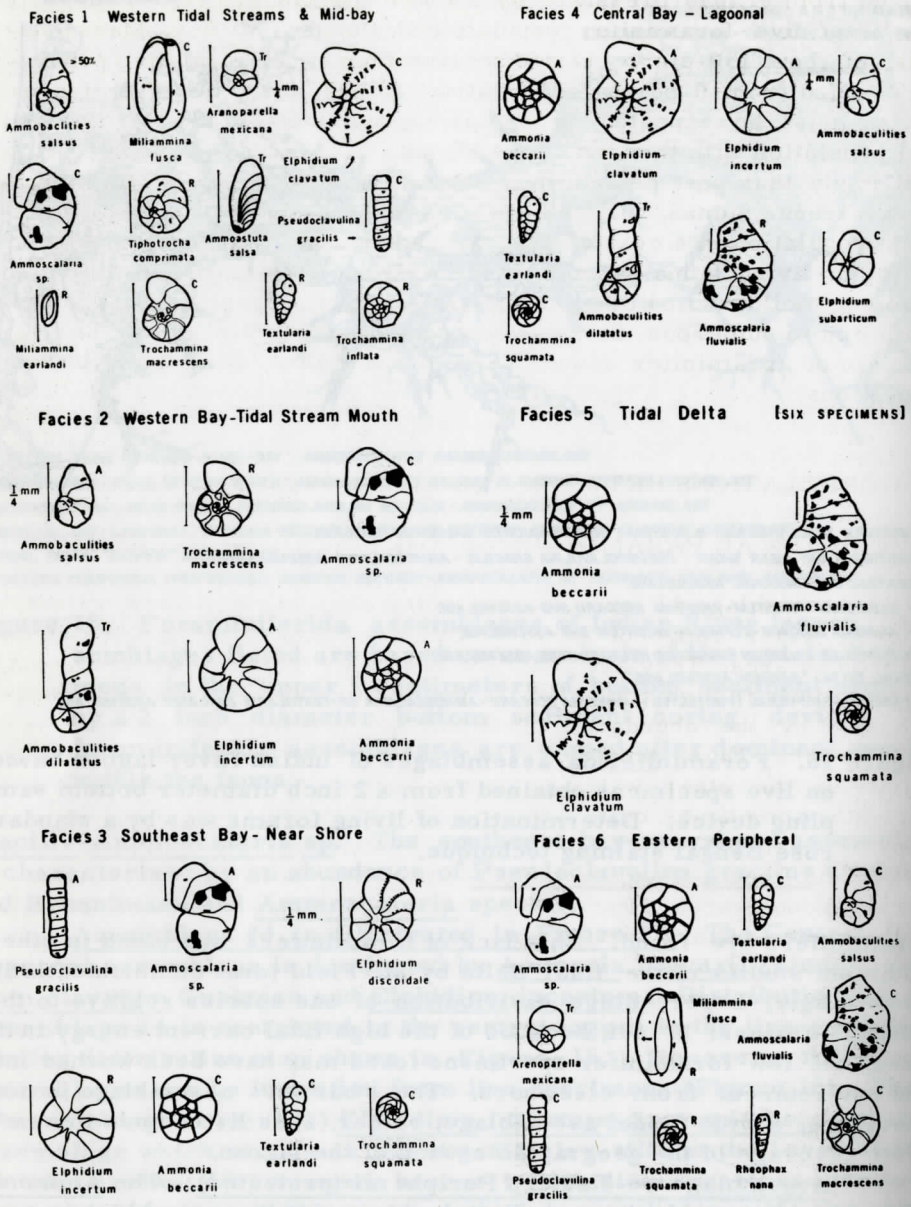

Figure 17. Camera lucida drawings illustrating Foraminiferida species that occur in the assemblages listed in Figures 15 and 16. The scale bar represents 0.25 millimeter. Relative abundance of specimens is indicated by a-abundant equals greater than 25 percent of the fauna, c-common equals 5 to 24 percent, r-rare equals 1 to 4 percent, Tr-trace equals less than 1 percent.

derived from graphs presented by Parker and Athearn (1959) suggests an average live foraminifer population in Poponesset Bay, Massachusetts, of about 150/cm^2. As can be seen from Figure 14, live populations varied from 0 per sample station to 222/cm^2. Serious discrepancies must be expected in comparing numbers of total or live and dead population with the live populations. As shown in Figure 6, it is most likely that post depositional alteration of the faunas, in particular the calcareous faunas, will seriously alter the makeup of the Foraminiferida population in a coastal lagoon area. Acid conditions, predators within the living fauna and the extreme amounts of sediment alteration by boring mollusks and other organisms subject the residual foraminifer population to numerous destructive processes. Accordingly, the best estimate of foraminifer faunal density must be derived from the live populations.

CORRELATIONS BETWEEN THE LIVE FORAMINIFERA AND THE PHYSICAL PARAMETERS OF THE WATER-SEDIMENT MASS

Assemblage #1 species, particularly Ammobaculites salsus, occupy a wider range of salinity conditions than do the species of the eastern bay area. They are abundant in Pepper Creek in the southwest, where salinity is high compared to the river, although their abundance is masked by the abundance of Elphidium incertum in the lower part of the creek. Thus, Ammobaculites salsus is probably not limited by salinity or the other water pattern parameters. Abundance of single species is sometimes quite erratic. The abundance of Ammobaculites drops off just past the tidal stream entrances. Textularia earlandi Parker and Pseudoclavulina gracilis are at a peak in the eastern peripheral lagoonal area. Perhaps Ammobaculites salsus cannot compete for the higher salinity environments and is thereby confined to the more brackish waters of the tidal streams and central bay area. Possibly, competition among the species of the Foraminiferida themselves may be severe. Very little work has been done to test the idea of competitive pressures on a single environment by varied species of the Foraminiferida.

Elphidium, particularly E. clavatum and E. incertum, also appear to have a similar relationship that may be competitive. Elphidium incertum has its largest numbers in tidal Pepper Creek to the west and spreads to the central lagoonal area. The depth range shown for Elphidium incertum is broader than for Elphidium clavatum. Elphidium clavatum has its greatest abundances in the deeper east bay area. For Elphidium clavatum, depth may be a limiting factor. Perhaps Elphidium incertum may also "prefer" the deeper environment but cannot compete with Elphidium clavatum for it. The Elphidium species are taxonomically closer to each other than are the agglutinate forms and

may be more likely to have similar niches if they are in competition. Whether or not this is a possibility for testing will have to await experimentation with physiological limits to determine restrictions on the environmental limits of tolerance for the Foraminiferida, and on the determination of techniques for accurate identification of partly dissolved tests of calcareous foraminifers in order to correctly estimate their distributions in the field.

None of the species appear to be limited in distribution by the sediment pattern. None correlate with the sand-mud ratio. On the other hand, minimal populations of all foraminifers have been noted to occur in the sandy, tidal delta area in the eastern part of the lagoon. Other investigators, Buzas (1965), Ronai (1955), Todd and Low (1961) and others have also found minimal populations in the sand substrates of the areas they investigated. Buzas also found no relationship between the particle size of the substrate and the numbers of foraminifers other than the barrenness of the sand areas. It may be that sediment particle size limits all species in sand areas but not elsewhere. Large sand grains may abrade the tests or bury them deeply when sand is shifted rapidly by tidal currents; thus, the animals may be killed before they can colonize a sandy area. No doubt, live foraminifers are transported into the tidal delta of Indian River lagoon along with other particles. The shallow depths may be another inimical factor where the sands are exposed by very low tides to desication and extreme atmospheric temperatures. Depth, in Long Island sound, in conjunction with sand substrate, may not be a limiting factor. The sand areas where Buzas worked are under 20 meters of water.

The clay particles of the sediment sampling stations to the west of the tidal delta in Indian River Bay tend to remain suspended longer than do the sand particles of the tidal delta and the clay may not be as difficult for the foraminifers to cope with. Thus an area that does not contain much sand or sand in rapid motion may have an adequate substrate for any of the species. Another inimical aspect of the sand areas may be lack of organic material which is often associated with such areas. If organic particles are taken as nutrient by the foraminifers then there may be too little food to sustain them.

CONCLUSIONS

Six Foraminiferida "assemblages", based on proportionate representation of various species, can be distinguished in the Indian River lagoon-estuary system. The distribution of the total foraminifer population (both live and dead specimens) is similar to the distribution of the live population but the areas occupied are not entirely similar. Generally, none of the physical parameters or combinations of parameters investigated in this study appear to control the distribution and abundance of the species found in Indian River Bay. Note that this

correlation has been based on only 20 samples selected by a random number technique. Obviously, further testing must be made using a larger number of samples. Furthermore, refinement of the characters of the water mass and the types of sediment character used must be made in an attempt to determine whether or not foraminifer species can, in fact, be correlated with physical parameters. On the other hand, there may, in fact, be no significant correlation between species occurrence and physical characters of the water mass and sediment.

Shallow, narrow, irregularly shaped coastal lagoons such as Indian River Bay, are strong mixing bowls. Possibly, the species of Foraminiferida which can live in these environments are highly tolerant of extremely varied conditions. If this is so, they may, accordingly, never be restricted to the narrow geographic zones of occurrances as delinated in this study. In Indian River lagoon-estuary, the water mass is well mixed except for the river stations and those near the inlet. Assemblages, while they do differ at the two ends of the estuary, also differ where the water mass is relatively homogeneous. Salinity, temperature and pH of bottom waters were not highly varied at the times that observations were made in this study and these parameters were observed in only a small segment of their daily and seasonal cycles. Their observed values probably do not nearly approach the limits of tolerance for the foraminiferal species encountered. Since the physiological limits are largely unknown, experimental investigation along these lines may be valuable. The apparent lack of correlation between the physical parameters and the estuarine species found in this study suggests that these species can tolerate wide ranges of conditions. Paleontologically, they may be indicators only of estuarine conditions rather than of the narrower ranges of parameters of the subcoastal lagoonal environments. Buzas (1965) attributed differences in the abundance of species to different nutritional requirements and suggested that they are "selective feeders". Said (1951) speculated that variations of abundance is due to predation. The results of this study strongly suggest that biotic environmental conditions such as competition and predation may assume some role in the distribution and abundance of foraminiferal species. These factors remain to be studied.

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APPENDIX

The Foraminiferida Species of Indian River Lagoon

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Ammobaculites dilatatus Cushman and Bronnimann
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Ammobaculites exiguus Cushman and Bronnimann
Cushman, J. A., and Bronnimann, P., 1948, Cushman Lab. Foram. Res. Contr., v. 24, p. 37-42.
Ammobaculites salsus Cushman and Bronnimann
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Ammoscalaria sp.
Arenoparella mexicana (Kornfeld) emend. Anderson, 1951.
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Pseudoclavulina gracilis Cushman and Bronnimann

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ORIGIN OF COLORS AND IRONSTONE BANDS IN THE COLUMBIA FORMATION, MIDDLETOWN-ODESSA AREA, DELAWARE

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ABSTRACT

Colors in the Columbia Formation (Pleistocene) in the study area are dependent on the amount of iron present. The source of iron found in these sediments is the underlying greensand of the Rancocas Formation (Paleocene-Eocene age). Iron seems to have been transported from its source by ground water and precipitated in the capillary fringe zone and zone of watertable fluctuations.

Ironstone bands, however, are thought to have been formed in the zone of aeration by the precipitation of iron from ground water. In general there is no relationship between the shape and size of the ironstone bands and the textures and structures of the sediments that contain them. The shapes of the bands reflect the type of the ground-water flow at the time of formation.

INTRODUCTION

Columbia (Pleistocene) sediments, which cover most of the Atlantic Coastal Plain portion of Delaware, consist of boulders, gravels, sands, silts, and clays. Sands are predominant; in general, they are poorly sorted and their mineralogy is strongly dominated by quartz (more than 80 percent). The amount of feldspars averages 18 percent. Rock fragments comprise about 1 percent and the rest is heavy minerals (about 80 percent are opaque minerals) (Jordan, 1964). Since the time of McGee (1887), and before, the Columbia sediments have been studied and classified, but many problems which they pose, such as origin and age, still baffle present day geologists (Jordan, 1962, 1964). At present, most agree on the fluvial origin of these deposits in northern Delaware (Marine and Rasmussen, 1955; Ward and Groot, 1957; Rasmussen et al., 1960, Jordan, 1964; Spoljaric, 1967, 1970).

In the area under study the Columbia sediments are unconformably underlain by the greensands of the Rancocas Formation (Paleocene-Eocene age). The greensands are unconsolidated and contain on the

average about 60 percent glauconite.

The purpose of this study is to investigate the origin of sediment colors; in addition, features produced by ground-water are discussed with special attention to the formation of ironstone bands, which are quite common in this area.

Acknowledgments

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METHODS

Field Methods

The study was conducted in two major phases: investigation of the Columbia sediment colors in outcrops, and the application of the results of this surface investigation to the interpretation of the distribution and origin of colors in the subsurface (Figure 1). Colors were described in the field and matched against those of the rock-color chart (Goddard et al., 1963) and given appropriate symbols.

Subsurface investigation was done in the area shown in Figure 1. Forty holes were drilled and samples were taken at 5-foot intervals and at lithologic and color changes in the sediments. The surface elevations of the holes were measured with an engineering level and an altimeter. For mapping purposes it was necessary to reduce the number of colors of drilling samples to five by grouping similar ones together: dark brown (10 R 3/4), brown (10 R 4/5), light brown (5 YR 4/4), yellow (10 YR 7/4), and light gray (10 YR 8/2).

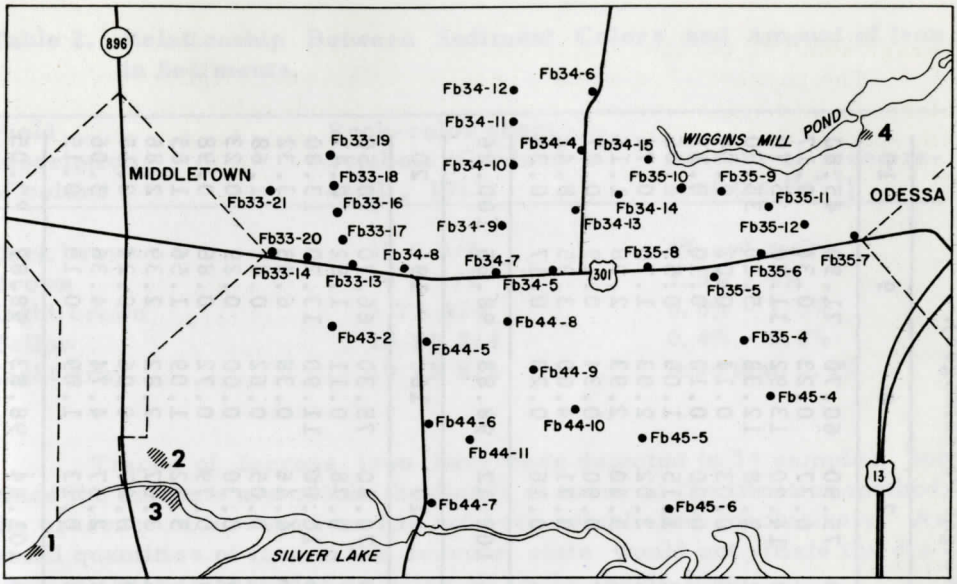


Figure 1. Location of the study area. The well numbering system is that used by the Delaware Geological Survey. The numbered shaded areas (1-4) refer to the gravel pits.

Laboratory Methods

Chemical analyses of the sediments were done by wet chemistry, atomic absorption spectrophotometry, and visible spectrophotometry. Analyses were performed as described by Shapiro and Bannock (1962) and Shapiro (1967). The totals of oxide percentages for individual samples usually add up to less than 100 percent probably indicating incomplete solution of SiO_2 and Al_2O_3. Sample solutions prepared by means of lithium tetraborate fusion and sodium hydroxide fusion reveal the same type of error. The difference in the individual oxide percentages obtained by the analyses of these two sets of the sample solution is negligible.

RESULTS

Color and Chemical Composition of Sediments

Twenty outcrop samples of the Columbia sediments were studied for color and chemical composition. The chemical compositions are presented in Table 1. Chemical analyses reveal that the color of the sediments is strongly dependent on the amount of iron present (Table 2).

Table 1. Chemical Composition of Columbia Sediments (in Weight Percentages).

Samples	1	2	3	4	5	6	7	8	9	10
SiO_2	65.20	57.20	64.20	70.80	68.80	59.00	75.90	60.70	71.90	45.80
TiO_2	1.10	0.92	0.21	0.28	2.17	0.51	0.37	0.23	0.34	0.17
Al_2O_3	13.57	13.66	9.40	11.24	12.90	10.70	14.90	13.92	11.50	5.70
Fe_2O_3	9.26	15.41	5.80	2.27	7.78	17.52	2.38	12.38	5.20	35.80
MnO	0.18	0.19	0.07	0.02	0.07	0.98	0.03	0.14	0.02	0.04
MgO	0.26	0.28	0.13	0.25	0.28	0.11	0.20	0.10	0.10	0.17
CaO	0.65	0.94	0.32	0.94	0.79	0.32	1.15	1.08	0.55	1.06
Na_2O	1.88	1.54	1.23	1.53	1.62	1.34	2.02	2.03	1.75	0.74
K_2O	2.65	2.46	2.30	2.59	2.70	2.64	3.40	2.93	2.98	1.16
P_2O_5	0.27	0.49	0.20	1.14	0.31	0.67	0.90	0.34	0.19	0.87
H_2O_T	4.02	6.20	4.77	2.19	3.75	6.21	0.71	4.80	3.95	8.42
CO_2	0.59	0.43	0.22	0.02	0.31	0.15	0.16	0.23	0.17	0.33
	99.63	99.73	98.85	93.27	101.48	100.14	102.12	98.88	98.65	100.26

Samples	11	12	13	14	15	16	17	18	19	20
SiO_2	66.80	77.80	44.60	77.50	73.50	31.40	70.30	75.30	66.70	70.50
TiO_3	0.63	1.19	0.04	0.33	0.32	0.17	0.78	0.11	1.82	0.47
Al_2O_3	13.20	11.00	6.30	12.90	13.90	5.30	16.00	11.90	13.20	13.80
Fe_2O_3	2.76	3.00	36.40	1.77	2.14	42.80	4.46	0.38	6.43	3.32
MnO	0.02	0.80	0.02	0.02	0.02	0.04	0.05	0.62	0.81	1.68
MgO	0.36	0.20	0.16	0.15	0.74	0.20	0.30	0.00	0.29	0.23
CaO	1.06	1.50	0.43	0.46	0.94	0.79	0.79	0.72	0.88	0.98
Na_2O	2.07	1.39	0.65	1.50	1.86	0.44	3.32	1.09	1.56	1.85
K_2O	3.02	2.15	1.35	2.35	3.46	0.75	3.07	2.82	2.39	2.86
P_2O_5	0.16	0.22	0.65	0.07	0.17	0.90	0.27	0.06	0.24	0.15
H_2O_T	2.52	2.27	8.80	1.48	2.17	0.78	2.37	4.74	4.36	3.06
CO_2	0.16	0.02	0.03	0.14	0.14	0.03	0.13	1.09	0.19	0.15
	92.76	101.54	99.43	98.67	99.36	93.60	101.84	98.83	98.87	99.05

Table 2. Relationship Between Sediment Colors and Amount of Iron in Sediments.

Field description of colors	Rock-color chart symbols (Goddard et al., 1963)	Amount of iron present (as Fe^{+3})
Dark brown	10 R 3/4	4% and more
Brown	10 R 4/6	1.5% - 4%
Light brown	5 YR 4/4	0.8% - 1.5%
Yellow	10 YR 7/4	0.4% - 0.8%
Light gray	10 YR 8/2	less than 0.4%

Traces of ferrous iron have been detected in 14 samples, but since the amounts are within the limits of error of the titration method, the presence of ferrous iron has not been established conclusively. As small quantities of iron in the ferrous state would not vitiate the discussion, iron is regarded as being all in the ferric form.

In addition to iron, manganese may sometimes determine the color of the sediment. When abnormally large amounts of manganese are present (more than 0.60 percent, MnO), the color becomes dark purple-black (5 R 2/2). Such large manganese content, usually in thin bands of quite limited extent, is exceptional and does not influence the overall interpretation of sediment color and amount of iron.

Other cations present show irregular quantitative variations from sample to sample and no consistent color relationship has been established.

The amount of water (H_2O total) in analyzed samples is directly proportional to their iron content (Figure 2). This suggests that iron is present in these samples in the form of hydroxides or hydrous colloids or gels. Several X-ray diffraction analyses using CuKα radiation failed to reveal peaks that could be used for identification of the iron compounds; absorption of CuKα radiation and fluorescence of FeKα radiation within the iron-rich samples may explain a lack of a clean diffraction pattern. However, as no iron minerals have been detected by microscopic investigation, it is concluded that these compounds are either amorphous or poorly crystallized.

Behavior of Iron

The amount of iron in natural waters is usually extremely small except under abnormally acidic conditions of pH less than five (Hem, 1960). The iron content of ground water is commonly less than one part per million (ppm) (James, 1966). However, in the glauconitic greensands (Rancocas Formation) in the study area it ranges between five and ten ppm, while in the Columbia samples the iron content is less

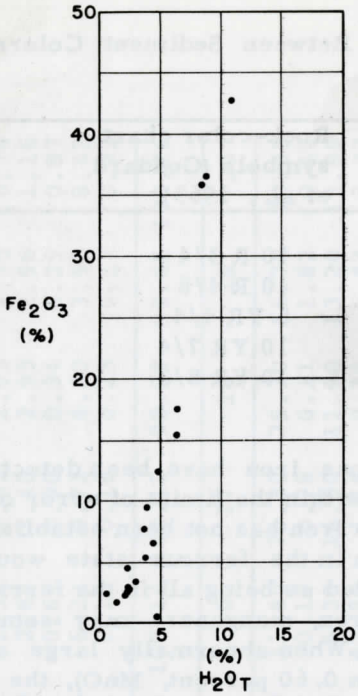


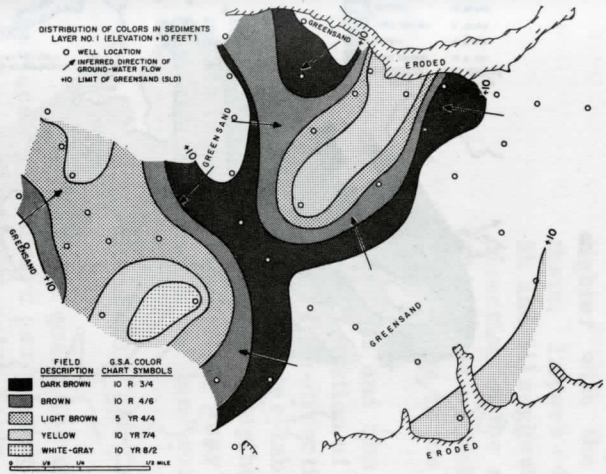
Figure 2. The relationship between the amount of iron and total water of the sediments.

than one ppm (Delaware Geological Survey file). High values found in the water from the greensands are due to large concentrations of ferrous iron in solution. This is in agreement with the data presented by Back and Barnes (1965) who found that almost all iron in solution in ground water is in the ferrous form. White, Hem, and Waring (1963) have also reported high iron concentration in acidic and reducing ground water. Hem and Cooper (1959) state that concentrations as much as fifty ppm are attained in natural waters under suitable chemical conditions.

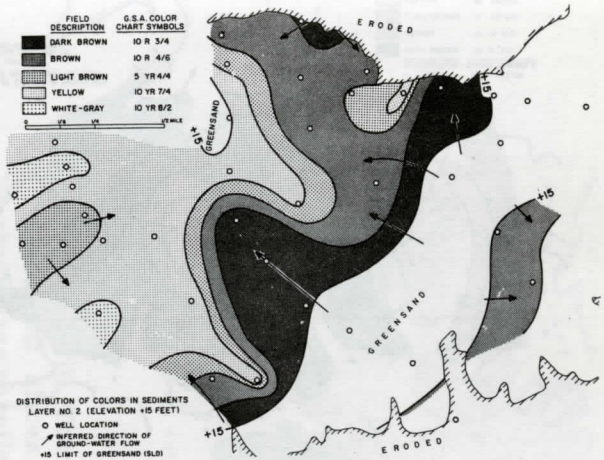
The only process now producing significant iron concentrations in solution is shallow subsurface leaching of iron and its subsequent transport in ground water of low pH and Eh (James, 1966).

Distribution of Colors and Transport of Iron

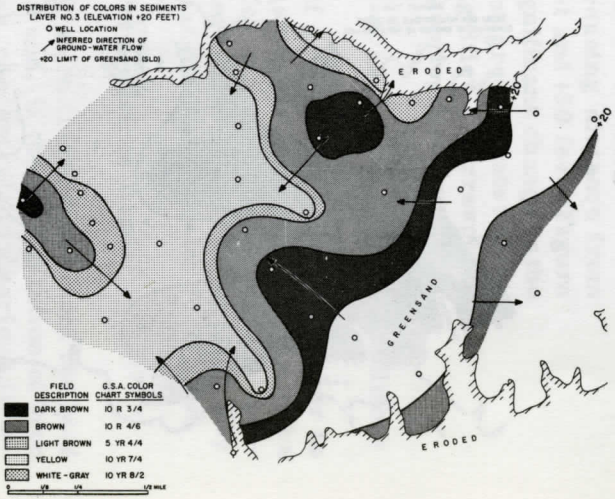
Distribution of colors in the Columbia sediments (Figure 3a to i) suggests that the transporting agent of nearly all iron was ground water of the saturated zone where the conditions were favorable for the ferrous iron to remain in solution. Oxidation of ferrous into ferric iron and its subsequent precipitation apparently occurred in the capillary fringe zone and zone of water-table fluctuation. Here the interface


DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 1 (ELEVATION +10 FEET)
O WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
+10 LIMIT OF GREENSAND (SLD)
FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE
GREENSAND
ERODED
+10

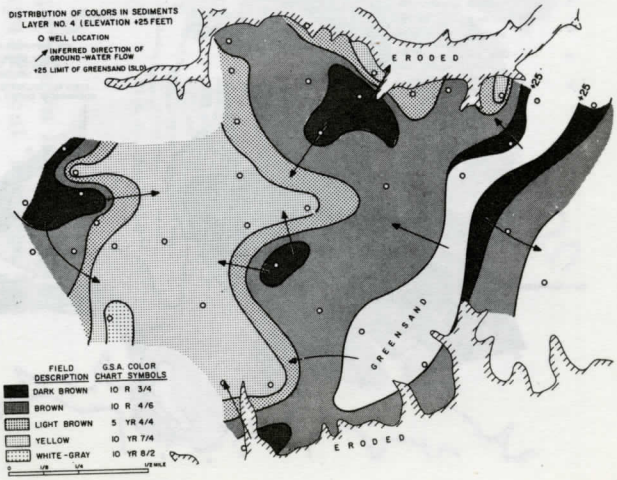
a


FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE
DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 2 (ELEVATION +15 FEET)
O WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
+15 LIMIT OF GREENSAND (SLD)
GREENSAND
ERODED
+15

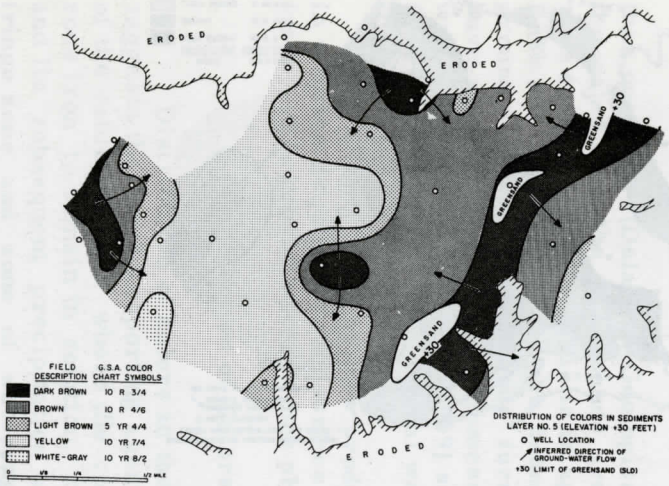
b


DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 3 (ELEVATION +20 FEET)
O WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
+20 LIMIT OF GREENSAND (SLD)
FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE
GREENSAND
ERODED
+20

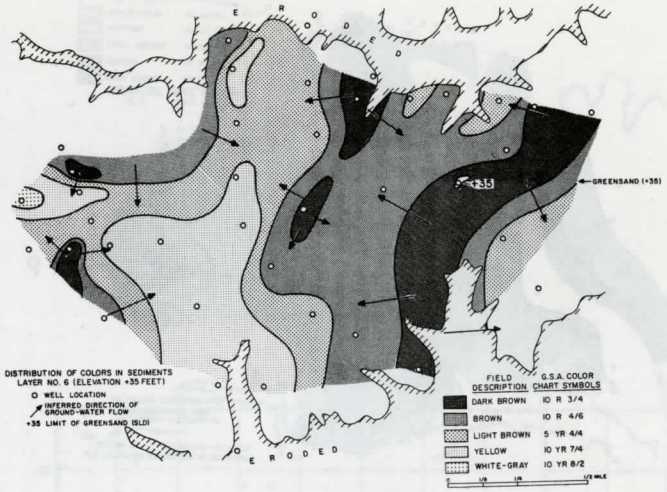
c


DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 4 (ELEVATION +25 FEET)
O WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
+25 LIMIT OF GREENSAND (SLD)
FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE
GREENSAND
ERODED
+25

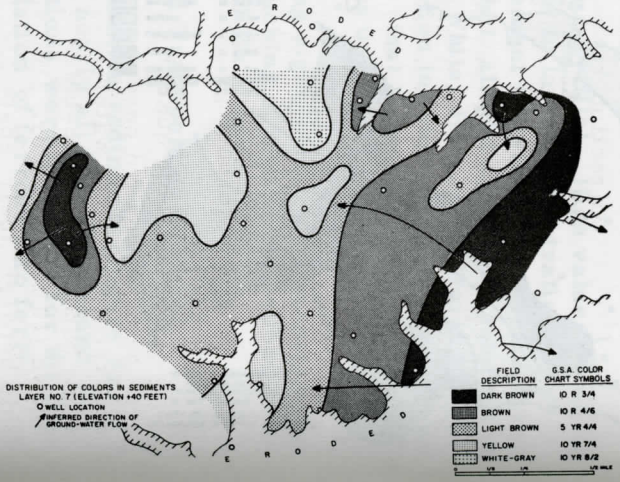
d


ERODED
ERODED
GREENSAND +30
GREENSAND +30
GREENSAND +30
ERODED
FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE
DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 5 (ELEVATION +30 FEET)
WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
+30 LIMIT OF GREENSAND (SLD)

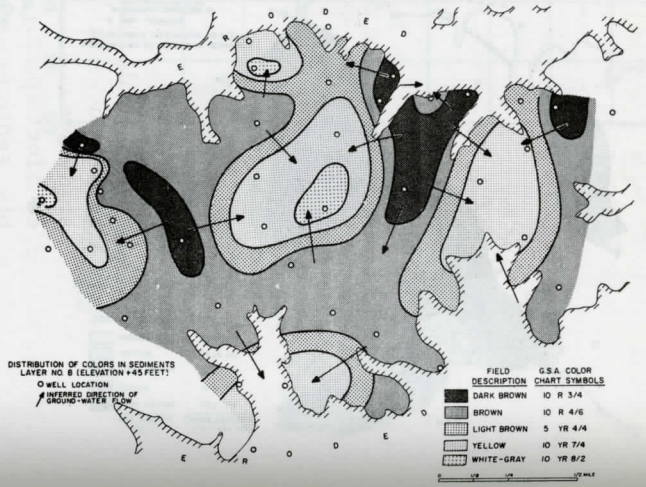
e


ERODED
+35
GREENSAND (+35)
ERODED
DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 6 (ELEVATION +35 FEET)
WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
+35 LIMIT OF GREENSAND (SLD)
FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE

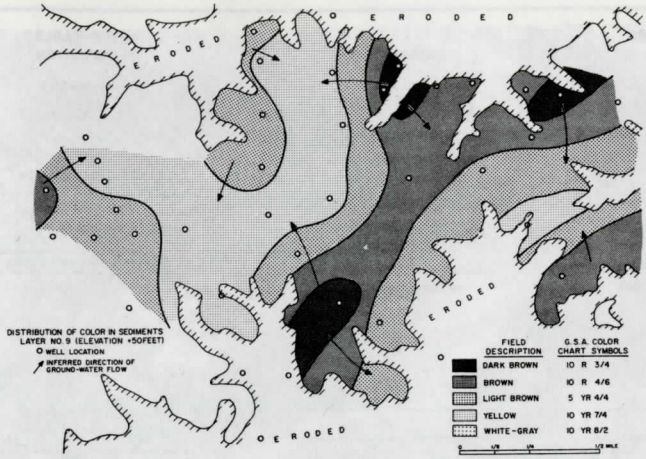
f


ERODED
ERODED
DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 7 (ELEVATION +40 FEET)
WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE

g


ERODED
ERODED
DISTRIBUTION OF COLORS IN SEDIMENTS
LAYER NO. 8 (ELEVATION +45 FEET)
WELL LOCATION
INFERRED DIRECTION OF GROUND-WATER FLOW
FIELD DESCRIPTION
G.S.A. COLOR CHART SYMBOLS
DARK BROWN 10 R 3/4
BROWN 10 R 4/6
LIGHT BROWN 5 YR 4/4
YELLOW 10 YR 7/4
WHITE-GRAY 10 YR 8/2
0 1/8 1/4 1/2 MILE

h



i

Figure 3 (a to i). The maps shown in this figure represent slices (layers) cut through the study area parallel to the sea level and spaced 5 feet apart. The dependence of the Columbia sediment colors on the location of the greensands is apparent; the colors are darker close to the greensands and they become lighter with the increasing distance from them. Layers +45 feet and +50 feet (Figure 3h and 3i) suggest a significant change in the direction of the ancient ground-water flow. This most probably signifies a corresponding change in the surface topography.

between the water-table and the zone of aeration produced conditions favorable for the precipitation of iron.

The inferred flow-direction of ground water is from higher to lower position of the water-table, as shown by arrows on the maps (Figure 3a to i); thus the distribution of colors in the Columbia sediments in the area of study may be used to determine topographically high and low areas of the land surface at the time of color formation.

Morphology and Origin of Ironstone Bands

Considerable quantities of iron were transported from the land surface by downward percolating water and, if the precipitation of iron occurred above the water-table, it was concentrated in various ironstone bands of purple-black (5 R 2/2), dark-brown (10 R 3/4), and brown (10 R 4/6) color. The study of these bands reveals a relationship between their shape and the type of ground-water flow operating at the time of their formation. The three mechanisms which are thought

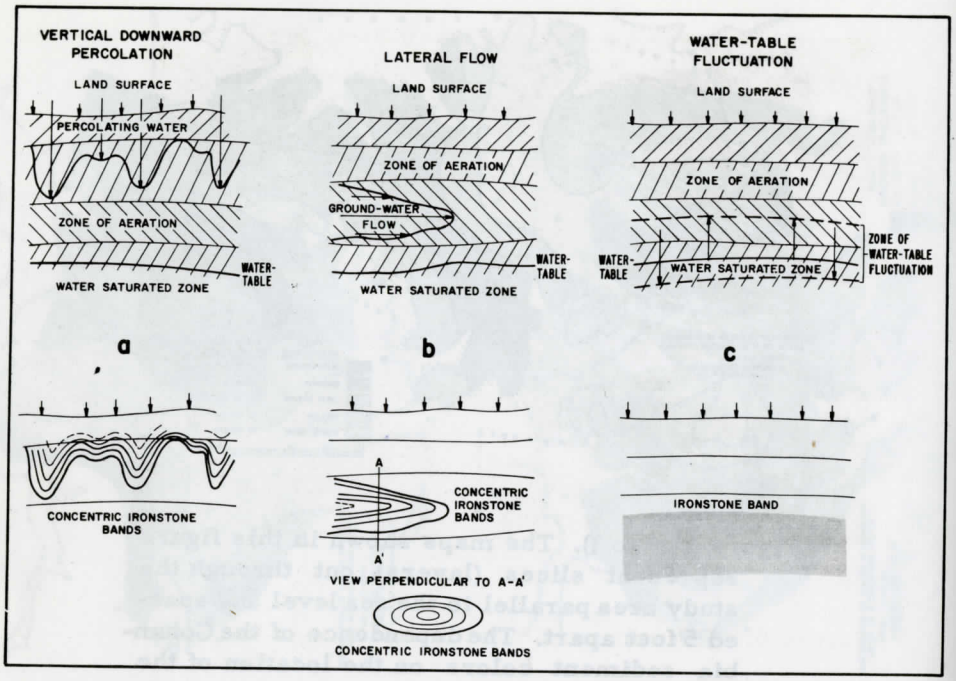


Figure 4 (a to c). Mechanisms which are thought to be responsible for the formation of various ironstone bands. In addition, the water-table fluctuations (Figure 4c) are thought to have produced most of the sediment colors in the Columbia Formation in the area of present study.

to be responsible for the formation of various types of bands are shown in Figure 4.

Bands pointed downward are either compound or single and they are probably formed by precipitation of iron from downward percolating water. Since the rate of such ground-water flow varied from place to place the bands produced by this mechanism are irregular (Figure 4a) and variable in size.

Lateral flow of ground water above the water-table is, evidently, caused by the deflection of downward percolating water into a horizontal (or nearly horizontal) flow; the cause of such a deflection is unknown. The ironstone bands formed by such ground-water movement appear in outcrops as either elliptical or sidewise pointed bands (Figure 4b and Figure 5) depending on whether the cross-sectional plane is vertical and perpendicular to the lateral ground-water flow, or vertical and parallel to it.

Fluctuation of water-table is a relatively slow process (compared with the percolation of ground water from the land surface downward) and thus the bands produced by this mechanism are somewhat

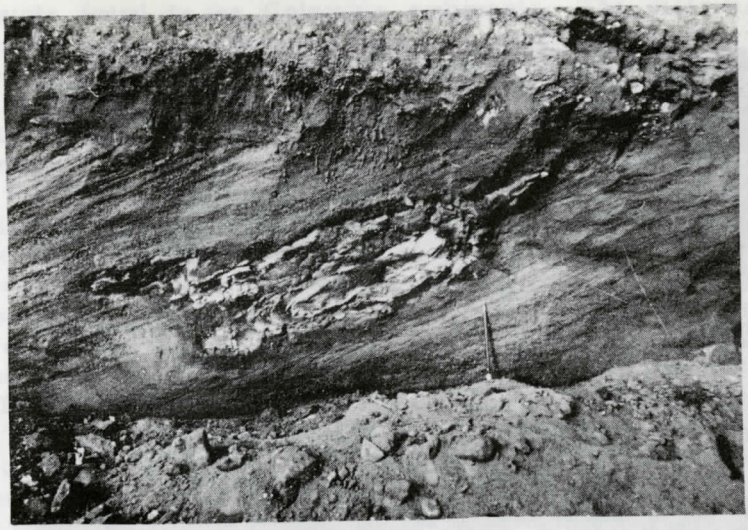


Figure 5. Elliptical and concentric ironstone bands.

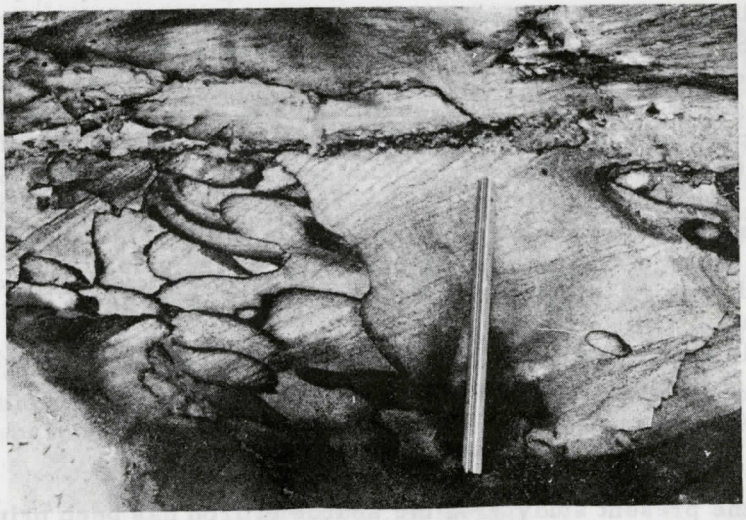


Figure 6. Mosaic-like network of complex ironstone bands.

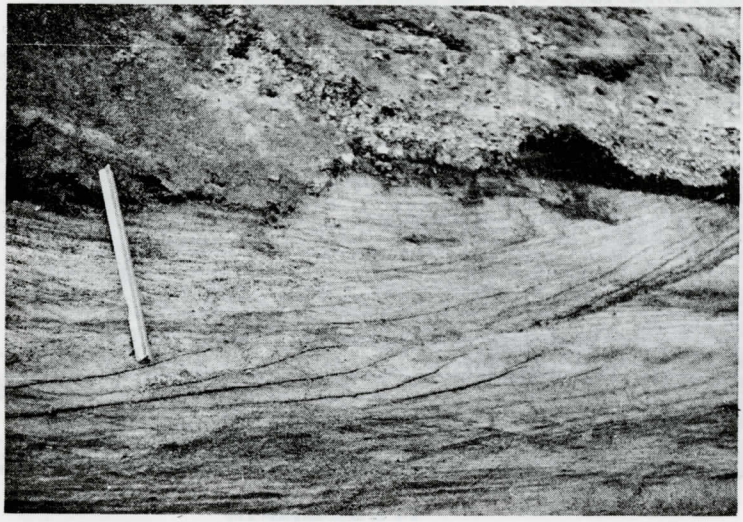


Figure 7. Ironstone bands formed along the foresets of cross-bedded sands.

different from the ones described previously. These bands are nearly horizontal and laterally persistent (Figure 4c) and usually they can be followed along whole exposures (at places more than 30 feet). Precipitation of iron takes place in the capillary fringe zone and in the zone of water-table fluctuation. This mechanism is thought to be responsible for the origin of most of the sediment colors in the area of study.

In case of a shallow water-table there is a possibility of interactions of two or even all three mechanisms described above. Such interactions are indeed believed to be responsible for the formation of complex bands that make up a mosaic-like network (Figure 6). Sometimes the order of formation (interaction) of individual bands can be determined, but in most cases the complexity is so great that this is impossible.

Only exceptionally is the development of ironstone bands controlled by sedimentary structures, as shown in Figure 7.

Sources of Iron

In the present study area the source of iron has been primarily sought in the underlying glauconitic greensands of the Rancocas Formation. The amount of iron in such sediments in New Jersey (Mansfield, 1920) and in Delaware (Delaware Geological Survey file) often exceeds 20 percent (total iron). Thus ample quantities are available to supply

the amounts found in the Columbia sediments. The distribution of colors in the subsurface deposits shows a correlation with the distance from the greensands: the sediment colors, in almost all cases, are darker near the greensands and lighter with increasing distance from them (Figure 3a-i). This is a strong evidence that the glauconitic greensands of the Rancocas Formation are indeed the source of most of the iron found in the Columbia Formation in this area. Because of the darker colors close to the greensands, it is postulated that the bulk of iron has been transported in solution and in the ferrous form (pH of ground water taken from greensands is always above 7; this would not allow ferric iron to remain in solution) from the greensand source to the site of deposition (precipitation). The color distribution is so regular and consistent that only this conclusion seems acceptable.

The highest elevation of the greensand surface in the study area is about +35 feet. If the greensands were the source of almost all iron in the Columbia sediments (and there is little doubt that this is not so), then it becomes difficult to explain a considerable accumulation of iron in these sediments at elevations above +35 feet (Figure 3g, h, i). It is, therefore, concluded that the original greensand topography must have exceeded at least +50 feet and that a relatively large amount of greensand was eroded during deposition of the Columbia sediments; the highest elevation known at present is about +47 feet, located just south of the area of study.

CONCLUSIONS

The close association of the Pleistocene Columbia sediments with the glauconitic greensands of the Rancocas Formation makes this, perhaps, a unique area to study the distribution and origin of sediment colors. The method employed in this study has shown useful in locating the source of iron which forms the pigment of the Columbia sediments and also in determination of ancient ground-water movement. Indirectly, this method has also enabled to recognize topographically high and low areas of the ancient land surface. However, because of the apparent uniqueness of the present study area, the application of this technique to other areas is probably somewhat limited.

The origin of ironstone bands is, no doubt, more complex than suggested here. One of the main reasons for not attempting to grasp the problem more deeply, is a meager knowledge of the ground-water movement in unsaturated zones (Rasmussen and Andreasen, 1959).

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PETROLOGY AND MICROPALEONTOLOGY OF ORDOVICIAN ROCKS IN CENTRAL ALABAMA

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ABSTRACT

Variations in constituent grain types, degrees of grain packing, microfossil assemblages and primary sedimentary structures provide criteria for recognizing the depositional environments of three Ordovician units which were selected for detailed local study in central Alabama. A Lower Ordovician (Beekmantown) Newala Limestone sequence in the Cahaba Valley comprises primarily unfossiliferous carbonate mudstones and intraclastic-pelletal packstones and grainstones interpreted as tidal flat deposits. A Middle Ordovician (lower Mohawkian) Little Oak Limestone sequence--also located in the Cahaba Valley--is a sparsely fossiliferous carbonate wackestone that bears an ostracode assemblage similar to that of the Edinburg Formation (lower Mohawkian) in Virginia. The Little Oak is interpreted as a below wave base, open-shelf "bioturbite." The other sequence that was studied is the Unit III portion of the Chickamauga Limestone (Rogers, 1961a) at a Birmingham Valley locality. This sequence consists of skeletal packstones and grainstones which are characterized micropaleontologically by ubiquitous octactinellid sponge spicules and a cyclostome bryozoan Corynotrypa. Unit III of the Chickamauga Limestone is interpreted as an above wave base, open-shelf deposit. Upper Ordovician rocks are absent within the study area.

INTRODUCTION

Friedman (1969b) stated that by using recently established criteria, the geologist is able to assemble a detailed picture of depositional environments for carbonate sequences on a local as well as regional scale. Lithologic, petrographic and paleontologic criteria given by Friedman (1969a), Laporte (1969) and Tyrrell (1969) for the environmental interpretation of carbonate rock sequences are applied by the author to three sequences of Ordovician age in central Alabama.

Financial support for this study was given by an Auburn University Research Grant-in-Aid, no. 68-65. Credit should be particularly

acknowledged for Thomas J. Carrington's generous cooperation in acquainting me with the local stratigraphy. Further appreciation is expressed to T. J. Carrington and G. N. DeRatmiroff for their helpful suggestions toward improving the manuscript.

STRATIGRAPHY

The stratigraphic section of the Lower Ordovician of central Alabama comprises mainly carbonate mudstones and pelletal-intraclastic packstones with fossils generally absent to sparse; Middle Ordovician rocks are generally skeletal lime wackestones with varying amounts of dolomite, chert and clay. The Lower Ordovician rocks are separated from the Middle Ordovician rocks by an unconformity, with either the Lenoir or Chickamauga Limestone (Unit I of Rogers, 1961a) overlying Newala or older Ordovician rocks. In ascending order the stratigraphic sequence pertinent to this study consists of the Newala, Lenoir, Little Oak and Chickamauga (Unit III of Rogers, 1961a) limestones (Figure 1). Upper Ordovician rocks are absent in central Alabama.

The Ordovician rocks of the Birmingham and Cahaba valleys strike northeastwardly outcropping infrequently as linear bands paralleling the deeply eroded centers of anticlines. Stratigraphic sections were described and samples were taken from the Newala Limestone and Little Oak Limestone at Cahaba Valley localities and from the Chickamauga Limestone sequence (Unit III) at the Birmingham Valley locality (Figure 2).

A comprehensive analysis of Ordovician stratigraphy in central Alabama can be found in Adams et al. (1926). The most recent account of Middle Ordovician stratigraphy and paleontology in central Alabama is that of Rogers (1961a, 1961b).

METHODS

Three localities (Figure 2), each selected on the basis of its continuity of section and lithologic "representativeness" of an Ordovician unit, provided the material of this study.

Samples were taken at 5-foot intervals from a nearly continuous 140' cored sequence of the Newala Limestone, a continuous 88' quarry exposure of the Little Oak Limestone and a continuous 148' roadcut sequence of the Chickamauga Limestone.

Stained rock slabs, acetate peels and insoluble residues were prepared from the 28 samples of the Newala section, 18 samples of the Little Oak section and 30 samples of the Chickamauga section. Each sample was trimmed, slabbed, polished, and etched with dilute hydrochloric acid. Acetate peels, 2 x 2 inches, were prepared from the

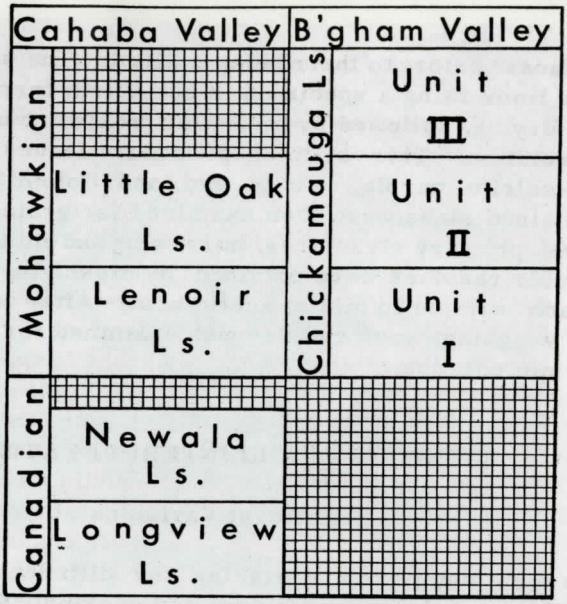

Figure 1. Correlation of Ordovician stratigraphic section in central Alabama with North American stages. (After Rogers, 1961a).

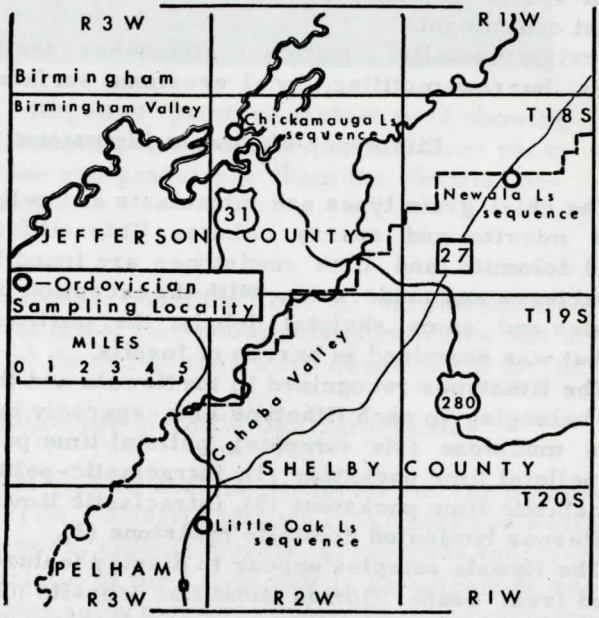

Figure 2. Index map showing the three Alabama localities where samples were collected for this study.

etched surfaces prior to their being stained. The staining procedure consisted of immersing a specimen in potassium ferricyanide solution, rinsing and drying, followed by a similar treatment in an alizarin red-S staining solution. This staining procedure leaves ferroan-dolomite blue, ferro-calcite purple, calcite red and dolomite unstained. The peels and stained slabs were then examined for grain type, fossil content, texture, primary structures, burrowing and mineralogic composition. Insoluble residues were obtained by dissolving approximately 50 grams of each sample in dilute acetic acid. After washing, filtering, drying and weighing, each residue was examined for microfossils with a binocular microscope.

LITHOTYPES AND INTERPRETATION

Significant Variables

The most significant variables for differentiating the primary lithotypes of this study are grain type and packing. Grain types recognized are skeletal debris, pellets and intraclasts; the degree of packing ranges from mud-supported rock with less than 10 percent grains to grain-supported rock.

Micrite is the dominant matrix material in the rocks of this study with sparite seldom a dominant matrix component and clay never a dominant constituent.

Environmentally significant structures are laminations, algal structures, burrow-mottling, local erosional surfaces and mudcracks.

Lithotypes of Newala Limestone

The chief grain types are intraclasts and pellets and the matrix types are micrite and sparry calcite. Unfossiliferous, frequently laminated dolomite and lime mudstones are found in association with algal structures and mudcracks. With the exception of a few snail-bearing samples and some skeletal debris, the portion of the Newala sequence that was examined is barren of fossils.

The lithotypes recognized in the Newala and the number of study samples belonging to each lithotype are: sparsely burrowed stromatolitic lime mudstone (six samples), pelletal lime packstone (7), snail-bearing pelletal lime packstone (1), intraclastic-pelletal lime packstone (4), intraclastic lime packstone (5), intraclastic lime grainstone (2) and unfossiliferous laminated dolomite mudstone (3).

The Newala samples appear to display features similar to those described from South Florida tidal flat deposits (Ginsburg, 1964) and Ordovician limestones of the Central Appalachians (Friedman, 1969a). The criteria for recognizing tidal flat environments are summarized by Friedman (1969a) as mudcracks, stromatolitic and oncolitic structures,

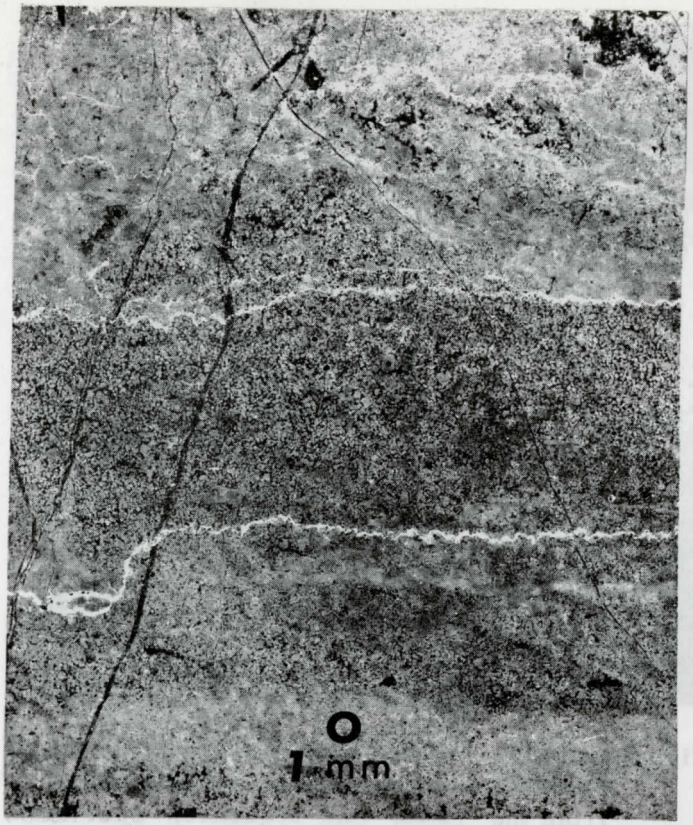

Figure 3. Negative print of acetate peel showing alternating intraclastic-pelletal lime packstone and grainstone from the Newala sequence.

finely laminated dolostones, dearth of fossils, burrow mottles, "birdseye" structures, intraclasts, flat-pebble conglomerates and scour-and-fill structures. All of these features were observed in the Newala.

Lithotype of Little Oak Limestone

The chief grain type is skeletal material and the dominant matrix component is micrite. Fossils include fragmented as well as whole specimens of brachiopods, colonial corals, gastropods and microfossils; particulary characteristic of the Little Oak is an ostracode assemblage consisting of Aparchites suborbicularis Kraft, Bairdiacypris incurvatus Kraft, Primitiella anterorotunda Kraft, Macrocyproides, and Shenandoia acuminulate Kraft. The ostracode assemblage of the Little Oak Limestone in Alabama resembles that described by Kraft (1962) from the Edinburg Formation of Virginia.

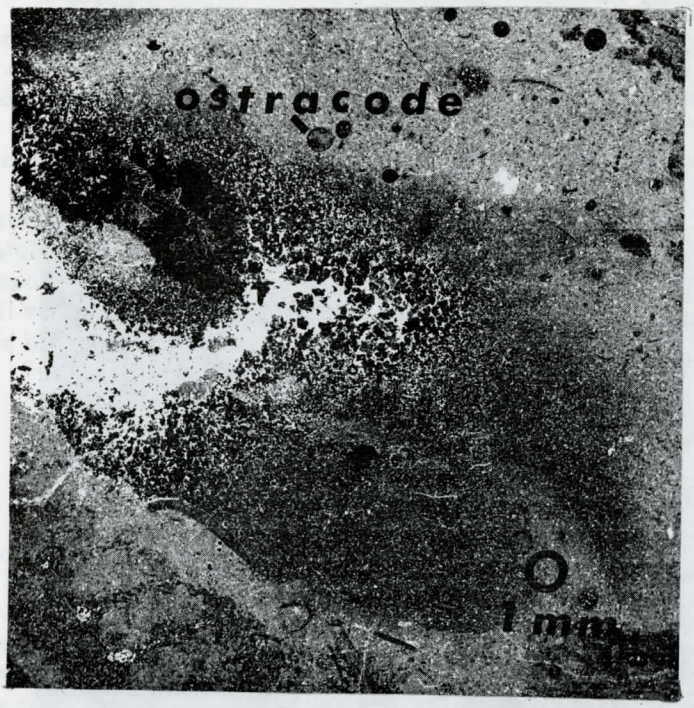

Figure 4. Negative print of acetate peel showing skeletal, burrow-mottled lime wackestone from the Little Oak sequence. Spar-rich burrow is partly replaced by dolomite (white area).

Skeletal, burrow-mottled lime wackestone (18 samples) is the sole lithotype in the Little Oak sequence. Abundant burrow-mottling of micrite suggests a shelf depositional environment below the zone of regular wave reworking (Laporte, 1969). A further indicator of a below wave base environment is the thickness of the wackestone sequence, uninterrupted by grainstone interbeds. Skeletal fragmentation is attributed to the churning and detritus-feeding activities of organisms.

The insoluble residues (8 and 21 percent) of the two least pure limestone samples consisted primarily of replacement chert.

Lithotypes of Chickamauga Limestone

The chief grain type is skeletal debris and the matrix materials are micrite and sparry calcite. The sediments of the Chickamauga were apparently reworked by currents resulting in sheet-stratified skeletal deposits that were relatively free of mud.

The disarticulated and fragmented remains of a diverse variety

Figure 5. Negative print of acetate peel showing skeletal lime grainstone with brachiopods, pelmatozoans and bryozoans from the Chickamauga (Unit III) sequence. Large white areas are replacement dolomite.

of invertebrate skeletal types, including sponges, corals, bryozoans, pelmatozoans and brachiopods, were identified in the Chickamauga samples. Particularly diagnostic of the Chickamauga are microscopic octactinellid sponge spicules and a microscopic encrusting bryozoan of linear series, Corynotrypa.

The Chickamauga lithotypes are skeletal lime packstone (12 samples), skeletal lime grainstone (11) and fossil-bearing lime wackestone (7). The textures and primary structures of the Chickamauga rocks are similar to those mentioned by Tyrrell (1969) as indicative of carbonates deposited in a shallow open-shelf environment. Textural types typical of this environment are skeletal grainstone and packstone. The beds of the Chickamauga commonly contain a sheet stratification indicative of a strong current control (Laporte, 1969).

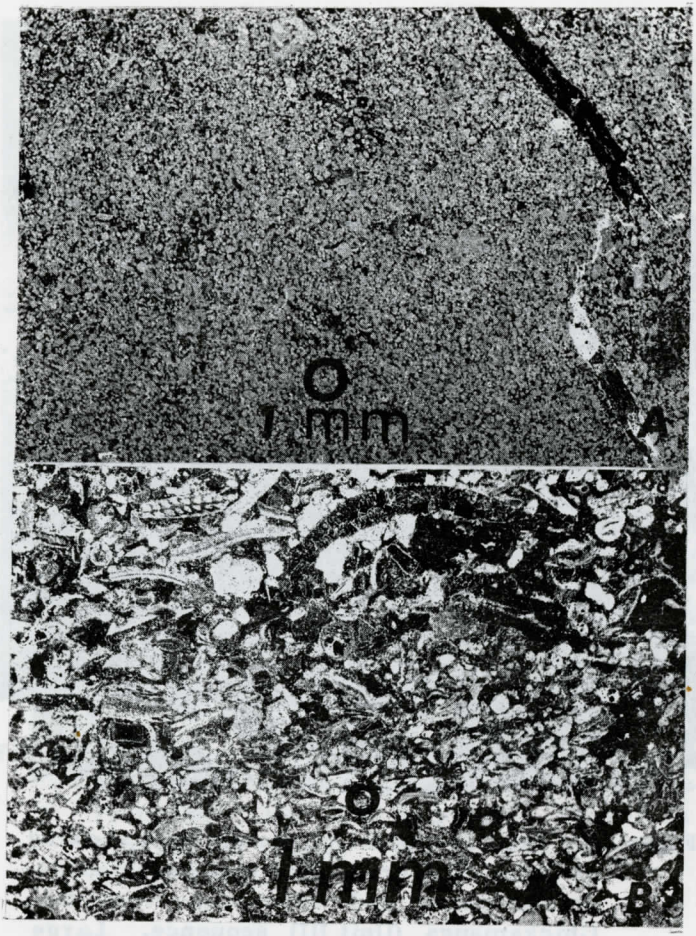

Figure 6. (A) Negative print of acetate peel showing intraclastic grainstone from the Newala sequence. (B) Negative print of acetate peel showing skeletal lime grainstone with sheet stratification from the Chickamauga (Unit III) sequence.

SUMMARY OF DEPOSITIONAL ENVIRONMENTS

The associated occurrences of erosional intraclasts, pellets, mudcracks, laminated deposits, algal structures and sparsity of skeletal grains indicate that the Newala sequence originated in a carbonate tidal flat environment.

The monotonous thickness of burrow-mottled, skeletal wackestone that constitutes the Little Oak Limestone sequence is interpreted as representing a below wave base, open-shelf environment in which carbonate mud accumulated.

The sediment of the Chickamauga (Unit III) is characterized by

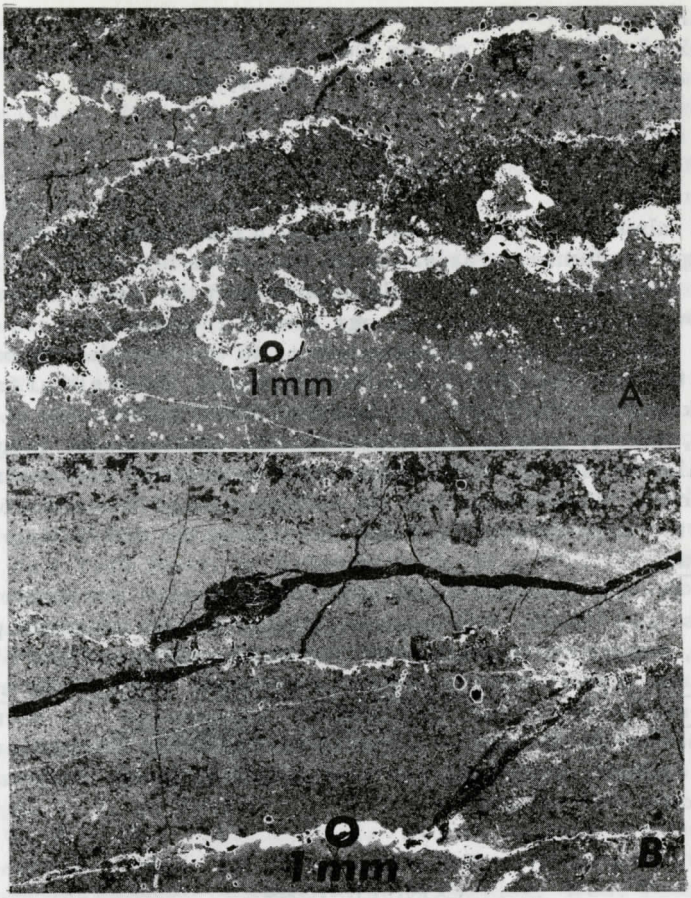

Figure 7. (A) Negative print of acetate peel showing alternating laminae of dense and "spongy" pelletal lime packstone from the Newala sequence. Spongy appearance results from spar-filled molds of algal filaments. White areas are dolomite. (B) Negative print of acetate peel showing alternating dense and "spongy" lime mudstone from the Newala sequence. The white band is dolomite.

broken skeletal material, a wide diversity of skeletal types, sheet stratification and sparite matrix that are interpreted to have resulted from abrasion, normal marine salinity, current transportation and deposition, and winnowing. These high-energy conditions are suggestive of an above wave base, open-shelf environment.

The depositional setting for the Southern Valley-and-Ridge during Ordovician time is described by Clark and Stearn (1968) as a broad

epeiric sea.

Typical lithotypes of the Newala, Little Oak and Chickamauga limestones are illustrated in Figures 3, 4, 5, 6 and 7.

MICROPALEONTOLOGY

The abundance of skeletal grains in the Newala, Little Oak and Chickamauga sequences varies from sparse in the tidal flat layers of the Newala sequence to moderately abundant in the Little Oak sequence to highly abundant in the Chickamauga (Unit III) sequence.

The dominant skeletal contributors to the Newala are snails. In contrast to the low diversity of the Newala fauna, the skeletal fragments of the Chickamauga (Unit III) are derived from a diversity of invertebrate types, namely brachiopods, pelecypods, gastropods, large ostracodes, corals, trilobites, stromatoporoids, bryozoans and echinoderms. The microfossil assemblage of the Chickamauga consists of sponge spicules of the octactinellid type and zooecia of the encrusting bryozoan Corynotrypa.

The fossils of the Little Oak wackestones are predominantly brachiopods and snails with accessory occurrences of small pelmatozoan plates, corals and ostracodes. The ostracode assemblage of Aparchites suborbicularis, Bairdiacypris incurvatus, Primitiella anterorotunda, Macrocyproides and Shenandoia acuminulata is identical to genera reported by Kraft (1962) from the Edinburg Formation of Virginia. The Edinburg Formation is also a dark gray, skeletal wackestone of Middle Ordovician age.

CONCLUSIONS

(1) The sediments of three limestone sections in the Ordovician of Alabama are interpreted as tidal flat (Newala), below wave base open-shelf (Little Oak), and above wave base open-shelf (Chickamauga, Unit III) deposits.

(2) Micropaleontologic evidence such as the diversity of ostracodes and the presence of abundant sponge spicules are aids in recognizing Ordovician depositional environments.

(3) The major lithotypes are categorized as unfossiliferous carbonate mudstones, intraclastic-pelletal packstones, fossil-bearing burrowed wackestones, and sheet-stratified skeletal lime packstones and grainstones.

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