SOUTHEASTERN GEOLOGY

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 25, NO.2 OCTOBER, 1984

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

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SOUTHEASTERN GEOLOGY

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THE CARBON CYCLE AND THE RATE OF VERTICAL ACCUMULATION OF PEAT IN THE MISSISSIPPI RIVER DELTAIC PLAIN

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ABSTRACT

A large percentage of the annual plant biomass production in interdistributary basins of the Mississippi River delta plain either remains on the marsh in the form of organic-rich sediment (peat) or is lost to the atmosphere as carbon dioxide and methane. Peat is formed due to vertical accretion as the marsh surface is maintained relative to a mean water level. The rate of accretion determined from ^{137}Cs dating averaged 0.85 cm and 0.95 cm yr^{-1} for the fresh and brackish peat deposits respectively. However, increase in water level, obtained from analysis of tide gauge data, was estimated to be in the order of .60-1.0 cm yr^{-1}. The rapid rate of peat accumulation (254-296 g C m^{-2} yr^{-1}) is attributed to increases in water level due to subsidence resulting from the consolidation of Mississippi River deltaic deposits.

INTRODUCTION

Peat accumulation is a common feature of Louisiana's Mississippi River deltaic plain. The development of the various Mississippi River delta lobes and associated interdistributary peat forming environments is well documented (Fisk, 1954; Coleman and Gagliano, 1964; Coleman and Smith, 1964; Kolb and Van Lopik, 1966; Frazier, 1967; Ho and others, 1976; Kosters and Bailey, 1983). However, it is not clear whether peat accumulation results from high organic production rates, slow decomposition rates or some combination of the two extremes. Net below ground accumulation of peat occurs when organic matter production exceeds the sum of the processes resulting in the loss of organic carbon (for example biological decomposition and/or detrital export).

The American Society for Testing and Materials (ASTM) defines peat as organic material containing more than 75% organic matter by dry weight. Although in Louisiana many peat deposits often do not contain more than the required 75% organic matter (Kosters, 1983), this study on rate of organic carbon accumulation in the Mississippi River Deltaic Plain provides useful information about modern peat-forming environments.

The interest to exploit peat deposits, a potential industrial fuel source, requires that we understand the dynamics of carbon cycling in wetland communities. Little is known about the carbon fluxes which control the rates of organic matter accumulation in Louisiana coastal marshes. The coastal marshes are extensive and represent 41% of total marshes in the United States (Turner and Gosselink, 1975).

Louisiana peat forming environments have been studied from the standpoint of modern day plant community structure and marsh typology (Penfound and Hathaway, 1938; Russell, 1942; Brown, 1948; O'Neill, 1949; Shiflet, 1963; Palmisano, 1970; Palmisano and Chabreck, 1972; Chabreck, 1972) and recently from the standpoint of peat diagenesis (Bailey and Kosters, 1983; Kosters and Bailey, 1983). However to date, studies have not considered the complex relationships that exist among carbon fixation, water level, subsidence, organic transformation and peat accumulation in coastal Louisiana. In this paper we have presented data and discussion showing the role of major carbon fluxes governing peat accumulation in the Mississippi River deltaic plain.

MATERIALS AND METHODS

Study Sites

Two study areas were chosen in the extensive fresh and brackish marshes to

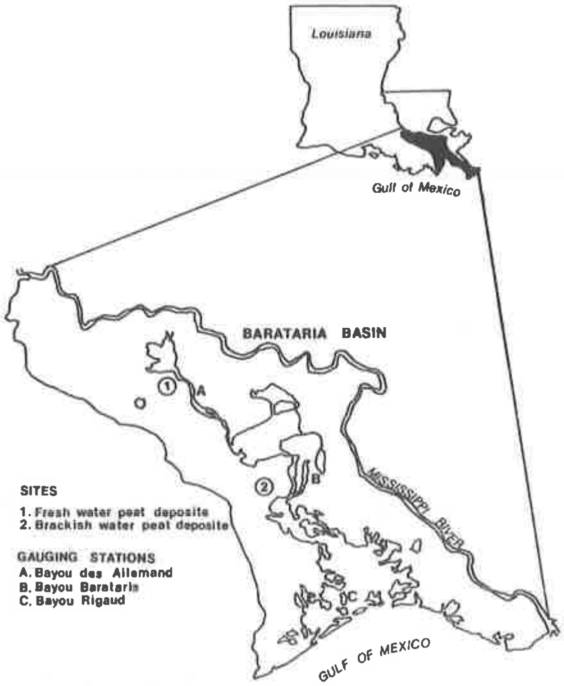

Figure 1. Location map of study area.

investigate the in situ interactions between changing water levels, organic matter production, organic matter accumulation, and decomposition reactions. The predominant vegetation of the fresh and brackish marsh is *Panicum hemitomon* and *Spartina patens* respectively. These marshes are the two predominant peat forming environments in coastal Louisiana and are located within Barataria Basin (Fig. 1). This basin is a well defined sedimentary and hydrologic unit in which accretionary processes rely on in situ plant production and mineral sediment input from inundation (DeLaune and others, 1978). The lower section of the basin is flushed daily by tides, but hydraulic energy rapidly diminishes inland from the coast. Inorganic sediment input during accretionary processes likely accounts for the reported high ash content of Louisiana peats (Kosters, 1983).

Annual Biomass or Organic Matter Production

Portable light and dark chambers were placed over portions of the marsh for determining the relative amount of biomass produced via photosynthesis. Two light chambers were constructed from 3 mm clear plexiglass. The chambers have an internal height of 77 cm, were 0.366 m^2 in cross-sectional area (square) and had an internal volume of 281 liters. The air temperature within the clear chambers did not increase more than 8^oC above ambient during the short measurement period. Another two dark chambers of similar internal dimensions were insulated with styrofoam (2 cm thick) and covered with a reflective space blanket. The air temperature within these chambers remained within 1^oC of ambient during the incubation. Details of the analytical procedure are described by Smith and others (1981).

This chamber method provided a static appraisal of CO_2 flux within the ecosystem. Changes in the carbon concentration within the light and dark chambers during incubation were treated as estimates of net ecosystem productivity and ecosystem respiration (plant + soil) respectively. Gross productivity was estimated from the carbon differential established between the light and dark chambers during incubation (Barko and others, 1977). Gross fixation, soil plus respiration and soil respiration, were estimated by integration of the temporal curves. Annual net production was calculated by subtracting plant respiration from gross fixation.

Peak vegetative standing crop was sampled on one occasion during October to determine annual aboveground production at each site. Live and dead standing material, cut 2 cm from the marsh surface, was collected from four randomly selected 0.25 m^2 quadrats. The samples were placed in burlap sacks and dried at 60°C until constant weight was obtained. The peak clip plot data was converted to annual production estimates using the yearly turnover rate for *Spartina patens* and *Panicum hemitomon* reported by Hopkinson and others (1978).

Emission of Carbon Dioxide and Methane from Peat Deposits

The emission of carbon dioxide and methane from the fresh water and brackish peats were measured by monitoring accumulation of the gases beneath chambers placed over the peat or marsh surface. The chambers were light proof which prevented photosynthesis. A similar diffusion chamber technique has been used by King and Wiebe (1978) to estimate methane emission. Triplicate determinations were from each peat type at approximately six week intervals. The marsh vegetation overlying the peat was clipped and removed prior to making the measurements. The chambers were shaded at all times, thereby minimizing the air temperature increase within the chamber during flux measurements.

The top of the chambers were fitted with an air sampling port constructed from a 0.6 cm Swagelock bulk-head union modified for gas-chromatographic septum penetration and a vent port constructed from a 1.2 cm Swagelock bulk-head union. An open ended tube (0.6 cm O.D. x 100 cm L) was attached to the vent port, thereby allowing pressure equilibration without entry of outside air during sampling. The vent port was capped on the chambers used on the open water bodies. Aliquots (10 cm^3) of the atmosphere within the chamber were withdrawn at 10 min intervals (0-60 min) into glass syringes (Smith and others, 1981).

Methane content of the gas samples was determined on a Varian 3700 gas chromatograph equipped with a flame ionization detector (FID) operated at 340°C. Carbon dioxide content of the gas sample was run on the same instrument using a catalytic conversion unit operated at 400°C (Williams and others, 1972). The catalyst oven was added between the column exit and the FID inlet (Smith and others, 1981). Carbon dioxide and methane fluxes were computed from the concentration increase within the headspace of the chamber (Smith and others, 1981).

Annual Rate of Peat Accumulation

Vertical marsh accretion rates in each peat deposits was determined using ^{137}Cs dating technique (DeLaune and others, 1978). Artificial marker horizons (white feldspar clay) are not suitable for the low density, high organic content (>30%) substrates found in fresh and brackish marshes (DeLaune and others, 1983). Annual average accretion rates were calculated from the depth in the peat profile of the radioactive element ^{137}Cs. It was first introduced into the biosphere as a result of atmospheric nuclear testing with fallout levels first appearing in 1954 and peaking in 1963 (Pennington and others, 1973). Cores (15 cm diameter) taken at each site were sectioned into 3 cm increments, dried, ground, and well mixed. The large diameter cores eliminates any compaction. The ^{137}Cs activity in each section throughout the core was determined by γ counting of the oven-dried sample using a lithium-drift germanium detector and multichannel analyzer. Total carbon content was determined by dry combustion. Density was determined from oven-dry sediment in the known volume of each section. Net carbon accumulation was calculated from the carbon content of the sediment, the accretion rate, and the bulk density.

Water Level

Average increase in water level was calculated by linear regression analysis of tide gauge data obtained from the U.S. Army Corps of Engineers, New Orleans District. The rise in water level included both the effects of true sea level rise and subsidence.

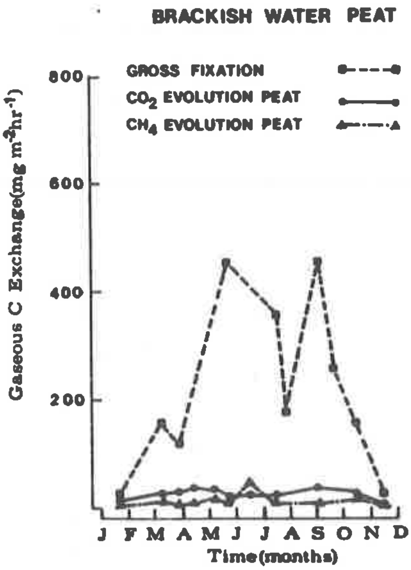

Figure 2. Gaseous carbon exchanges from brackish water peat deposit.

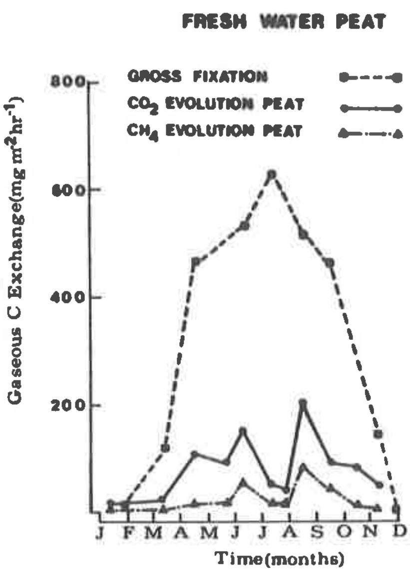

Figure 3. Gaseous carbon exchange from freshwater peat deposit.

RESULTS

Biomass Production

In situ estimates of gross CO_2-C fixation are presented in figures 2 and 3. Annual net carbon fixation was estimated to be 398 and 699 g C m^{-2} for the brackish peat *(Spartina patens)* and fresh peat *(Panicum hemitomon)*, respectively (Fig. 2 and 3). Peak photosynthetic activity occurred during the early summer, and decreased in October. Net carbon fixed was converted into biomass using average carbon content of the marsh macrophyte (40% C), thus the equivalent of 995 and 1750 g of plant biomass per m^2 was produced annually in the brackish and fresh peat respectively. Annual production estimated from clip plot data were 1128 and 1456 g m^{-2} for the brackish and fresh peat respectively. These estimates agree closely with the biomass estimate obtained using gaseous exchange measurements.

Gaseous Carbon Emission to the Atmosphere

The seasonal emissions of carbon dioxide and methane for the two peat deposits are shown in figures 2 and 3. The average daily carbon dioxide emissions from the brackish and fresh peat deposit were 490 and 1690 mg C m^{-2}, respectively. Points plotted in the figures represent the mean of three individual determinations having coefficients of spatial variability averaging 55 and 72 over the sampling dates for the brackish and fresh marsh, respectively. The carbon dioxide emission was equivalent to an average annual loss of 180 and 620 g C m^{-2} for the brackish and fresh peat deposits respectively.

There was an apparent seasonal trend in the carbon dioxide evolution for each peat deposit which was found to be statistically significant. Carbon dioxide evolution was directly related to the sediment and air temperature ($P \leq 0.05$). Generally the carbon dioxide evolution was greater during the summer period when temperatures were high as compared to the winter months. However the vertical flux of carbon dioxide to the atmosphere was found to have an inverse relationship with depth of water inundating the peat deposits ($P \leq 0.05$).

The average daily methane emissions from the brackish and fresh peat deposits were 200 and 440 mg C m^{-2}, respectively. The methane emission was equivalent to 73 and 160 g C m^{-2} yr^{-1}, respectively for the brackish and fresh peat deposits. Although the average evolution of methane from the fresh peat deposits was greater than the evolution from the brackish peat deposits the difference was not significant ($P \leq 0.05$). There was also an apparent seasonal trend in the methane evolution from the two peat deposits which was found to be statistically significant. Methane emission from the peat deposit to the atmosphere was greater during the summer months as compared to the winter months (Figs. 2 and 3). Other investigators have also reported the flux of methane to be a function of temperature both in situ and in vitro (King and Wiebe, 1977).

Annual Rates of Peat Accumulation

In this study net vertical accretion or rates of organic carbon accumulation was essentially the same in each organic deposit studied. Vertical accretion ranged between 0.95 cm for the brackish peat and 0.85 cm for the fresh peat. The organic carbon content was 35.0 ± 4 and 38.1 ± 3 percent respectively for the brackish and fresh marsh. Bulk density for each site was .08 g cm^3 and .09. Calculations using bulk density, accretion rates, and organic carbon content show that annual carbon accumulation was essentially the same for each peat studied and ranged between 296 and 254 g C m^{-2} yr^{-1} for the brackish and fresh peat deposits respectively.

Water Level

Increase in water level at the Bayou Rigaud gauge was 1.6 cm yr^{-1} between 1963 and 1980. Increase in water level was 1.0 and 0.45 cm yr^{-1} at the Bayou Barataria and Bayou Des Allemands gauge, respectively (Fig. 4). Water level increases in the fresh and brackish peat deposits studied would be in the range of 0.6 cm yr^{-1} to 1.0 cm yr^{-1}. The decrease in rate of water level change in moving inland from the coast reflects lower regional subsidence in the inland areas. The rapid subsidence in the Mississippi River deltaic system is a result of the concomitant processes of crustal downwarping caused by sediment overburden, consolidation of sediments of the Gulf Coast geosyncline, and local consolidation. Coleman and Smith (1964) concluded that differential compaction of organic rich sediment horizons is minimal compared to overall subsidence.

DISCUSSION

In the interdistributary coastal marshes in the Mississippi River deltaic plain, the

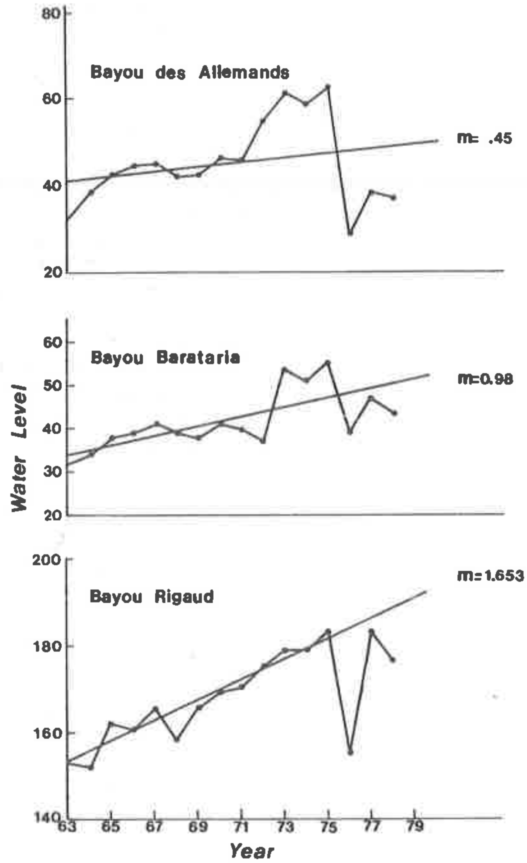

Figure 4. Water level increases in Barataria Basin (M = rate of water level increase in cm yr^{-1}).

marsh surface must maintain its elevation with mean water level rise in order for the macrophytes to survive and remain viable. A diagrametric representation of processes occurring simultaneously in the brackish and fresh peat deposits are shown in Fig. 5. Rates are summarized in Table 1. In sediment deficient environments with rapid subsidence there are large increases in water level. If the peat is to support productive macrophytes, essential for their continued survival, the surface must vertically accrete to keep the macrophytes in the intertidal zone. This occurs through the input of inorganic sediment and accumulation of the organic material produced by the macrophytes which decomposes slowly under anaerobic conditions. Even though the marsh soil is anaerobic a significant portion of the organic material deposited and remaining on the marsh is decomposed and loss to the atmosphere as carbon dioxide and methane.

At present a balance exists in the fresh and brackish marshes between production, accretion and decomposition. Carbon fixed by marsh macrophytes in the fresh and brackish peat deposits counterbalances the amount of carbon being deposited as peat and the carbon evolved to the atmosphere as methane and carbon dioxide. However what the future holds for these large carbon reservoirs is not clear. In the salt marshes in the lower portion of the basin, which borders the Gulf of Mexico, accretionary processes have not kept pace with apparent sea level rise and consequently the area is rapidly deteriorating (Baumann and DeLaune, 1982). In conjunction with subsidence and increasing water level the tidal passes at the lower end of Barataria Basin are widening thereby permitting a general increase in salinity within the entire basin (Levin and others, 1983). The increase in water level and greater salinity can reduce marsh macrophyte production, the organic source for peat formation. Since the marsh

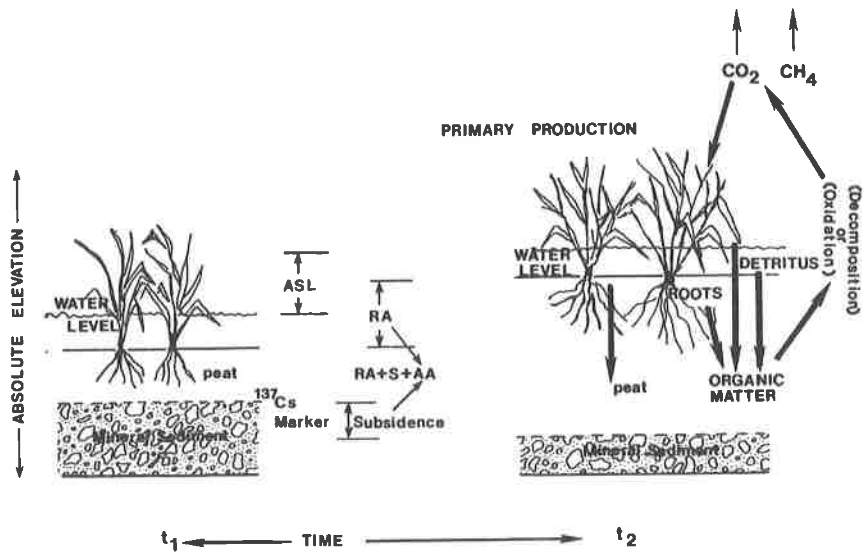

Figure 5. Schematic of processes governing peat accumulation (ASL Absolute Sea Level, S = Subsidence, RA = Rate of Accretion, AA = Absolute Accretion).

Table 1. Major carbon fluxes in fresh and brackish marshes.

	Brackish		Fresh	
	Daily	Annual	Daily	Annual
Net carbon fixation		398 g C m^{-2}		699 g C m^{-2}
Net carbon fixed as biomass		995 g C m^{-2}		1750 g C m^{-2}
Clip plot biomass		1128 g C m^{-2}		1456 g C m^{-2}
CO_2 emission	490 mg-C m^{-2}	180 g C m^{-2}	1690 mg-C m^{-2}	620 g C m^{-2}
CH_4	200 mg-C m^{-2}	73 g C m^{-2}	440 mg-C m^{-2}	160 g C m^{-2}
Net vertical accretion		0.95 cm yr^{-1}		0.85 cm yr^{-1}
Carbon accumulation		296 g C m^{-2}		254 g C m^{-2}

remains saturated most of the year, increase in water level is not likely to increase decomposition of methane and carbon dioxide evolution. Increase in frequency of inundation will enhance the mineral sediment content of peat and contribute to the high ash content reported for Louisiana peats (Kosters, 1983).

Eustatic sea level controlled by global temperature is projected to increase due to increased amounts of carbon dioxide in the atmosphere (Hansen and others, 1981). The likely effect of the projected increase in sea level on coastal peat deposits is uncertain. In rapid subsiding coastal areas such as the Mississippi River Delta Plain where marshes rapidly deteriorate, any increase in eustatic sea level will likely remove this area as a major global carbon sink. However, existing deposits may be preserved in the subsurface and become fossil carbon sinks.

ACKNOWLEDGMENTS

This research was supported by Louisiana State University Center for Energy Studies and the Office of Sea Grant (NOAA).

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IN SITU AND TRANSPORTED INVERTEBRATE ASSEMBLAGES FROM THE UPPER CLIFF COAL INTERVAL, PLATEAU COAL FIELD, NORTHERN ALABAMA

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ABSTRACT

Two fossiliferous horizons of invertebrate macrofossils associated with coal deposits from the Upper Cliff coal interval (Pennsylvanian), Pottsville Formation in northern Alabama are recognized. These two horizons preserve both transported and *in situ* fossil assemblages.

The lower horizon occurs within a shaley-siltstone lithofacies and is dominated by the inarticulate brachiopod *Orbiculoidea*, the bivalve *Pteronites*, and the trace fossil *Planolites*. Invertebrates from this horizon are preserved with articulated valves, in living position, or in masses horizontal to bedding but still articulated.

The upper fossiliferous horizon occurs within a sandstone lithofacies and consists of a basal transported shell bed *(Schizophoria* zone) dominated by the brachiopod *Schizophoria*. The *Schizophoria* zone contains a mixed brachiopod-mollusc assemblage composed of a transported marine fauna and a *in situ* restricted marine fauna. The transported elements consist of open marine epifauna that are commonly disarticulated and indicate hydrodynamic size-sorting. The *in situ* fauna consists of *Pteronites* and *Wilkingia* preserved in living position or oriented horizontal to bedding.

INTRODUCTION

Studies of Early Pennsylvanian marine invertebrate faunas in Alabama are few and lacking in detail. The Black Warrior Basin has served as the locus for most of the existing published investigations. Butts (1926) listed four fossiliferous horizons in the Warrior Coal Field and figured specimens from localities outside Alabama. Metzger (1965) compiled extensive faunal lists of invertebrate forms occurring in each of his seven stratigraphic intervals in the Black Warrior Basin, but presented no taxonomic descriptions. McKee (1975) studied sediment-fossil relationships concluding that the maximum mean grain size of the enclosing sediment was the limiting factor in the occurrence of invertebrate fossils, but again, details concerning taxonomy were lacking. Recently, Henry and others (1981) used the occurrence of the goniatite ammonoid *Belinguites elias* Manger and Sanders to indicate a Middle-Early Pennsylvanian age for rocks of the Parkwood Formation in Franklin County.

To date, there have been no detailed published studies of Pennsylvanian invertebrates within the Plateau Coal Field. Gray (1981) and Gibson (1982, 1983) reported the occurrence of productid brachiopods, pinnaceans, pectinaceans, trilobites, and abundant trace fossils in the Upper Cliff interval of the Pottsville Formation in Blount County. Detailed taxonomic descriptions and stratigraphic distribution of these faunas are currently being prepared (Gibson and Gastaldo, in preparation). The purpose of this investigation is to present evidence for the *in situ* and transported nature of the faunas preserved in the Upper Cliff interval in Blount County. Knowledge of the preservational mode of the fauna is essential for local paleoecological reconstructions and determining geographic extent of the fossiliferous horizons (Gibson and Gastaldo, in preparation).

METHODS AND MATERIALS

The study area is located within the Plateau Coal Field approximately 10 kilometers northwest of Oneonta, Alabama (Figure 1). Contour mining practices have uncovered extensive exposures of the Upper Cliff interval.

The Upper Cliff coal interval (Coulter, 1947) consists of shales, siltstones,

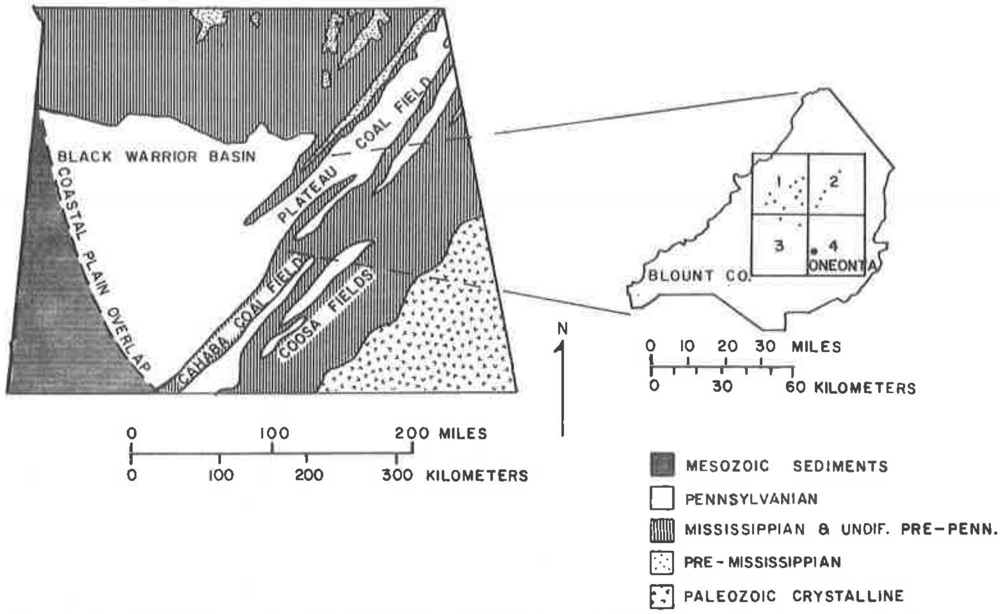

Figure 1. Geologic map of northern Alabama showing the distribution of the four principal Pennsylvanian coal fields. Inset shows the Blount County study area. The dots represent measured section localities. 1- Blountsville 7.5 minute quadrangle, 2- Clarence 7.5 minute quadrangle, 3- Cleveland 7.5 minute quadrangle, 4- Oneonta 7.5 minute quadrangle. (Modified after Ferm and others, 1967).

sandstones, and coal lithologies. Gibson (1983) reported invertebrate zones which occur above the Upper Cliff #1 coal in coarsening upwards, thinly-bedded dark gray shales and shaley-siltstones (Figure 2). Siderite ($FeCO_3$) nodules and bands are common in the basal portion of the lithofacies. This unit is overlain by fine to medium grained, buff-colored sandstones that are highly fossiliferous in the basal part.

Suites of specimens collected for taxonomic study were investigated to determine whether or not they represent an *in situ* or transported assemblage. In this study, the term *in situ* also includes fossils that may have underwent reworking but did not undergo much lateral transport (see below). Essentially, they still remain in the area in which they lived. Eight criteria were used to determine the nature of the occurrence.

Fossils oriented in living position based upon published studies of inferred paleoautecology provide conclusive evidence of an *in situ* occurrence (Fagerstrom, 1964; Dodd and Stanton, 1981). Articulation of valves, especially in the bivalves, indicates that fluid-dynamics was minimal or non-existent and, thus, the fossil assemblage is *in situ* or at best, only locally reworked (Johnson, 1960; Dodd and Stanton, 1981). Delicate forms of ornamentation such as spines tend to become broken during transport or during destructive biologic activity. Therefore, well preserved delicate ornamentation is generally indicative of *in situ* conditions (Johnson, 1960; Fagerstorm, 1964; Dodd and Stanton, 1981).

Differential abrasion often occurs during conditions of transport, whether over great distances or locally (Johnson, 1960; Fagerstrom, 1964; Dodd and Stanton, 1981). For example, raised features on a shell tend to abrade faster than grooved features. The amount of abrasion is a function of the shell structure and flow regime.

Fragmentation of shell material is considered strongly indicative of transport conditions (Johnson, 1960; Fagerstrom; 1964; Dodd and Stanton, 1981). Predation is another factor affecting the amount of fragmentation of fossil shells.

Many fossil organisms, for example, crinoids, are composed of numerous parts that disarticulate rapidly upon death. The various parts may behave differently depending on the prevailing current or wave activity. The result of higher energy

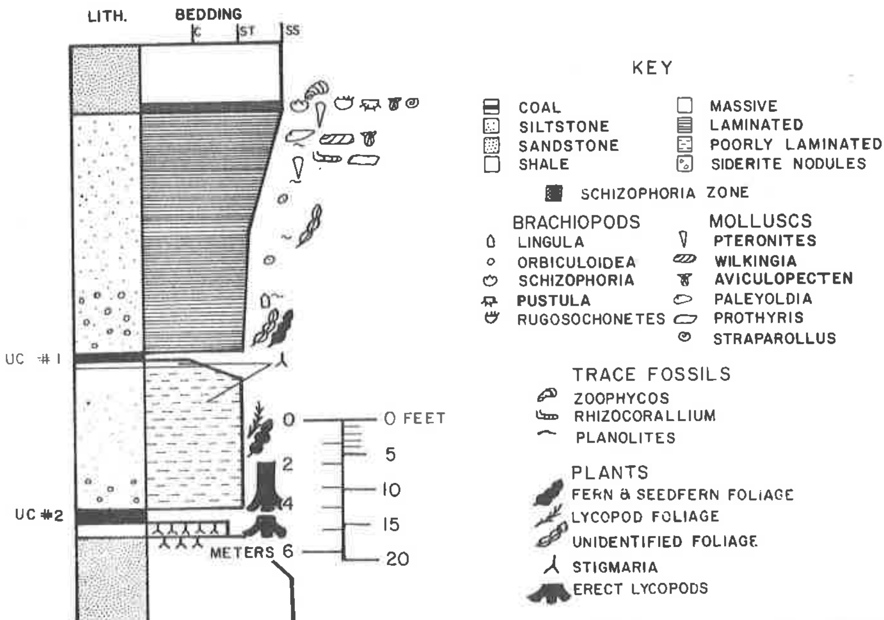

Figure 2. Generalized measured section of the Upper Cliff coal interval. Lithology is shown on the left. Bedding, texture, and fossil content is shown on the right.

influence is a general scattering of the disarticulated parts. These sorted skeletal parts are usually of similar density, size, and construction and tend to concentrate in one place. Beds composed primarily of broken crinoid columnals are good examples. Therefore, completeness of a multi-parted articulated organism can be of some use in determining the degree of reworking or transport of a fossil.

Size sorting is evidence for transport either due to actual transport of the shell material or winnowing of surrounding material leaving behind a shell lag (Johnson, 1960; Fagerstrom, 1964; Dodd and Stanton, 1981). Finally, preferred orientation can sometimes be used. Fluid dynamics can orient elongate fossils so that their long axis is parallel to current movement. The observer must, however, discriminate between orientation due to transport and orientation as part of the organisms paleoautecology.

With the exception of life position, each of the above criterion alone does not indicate conclusively whether a fossil is preserved *in situ* or has undergone reworking or transport. Some criteria are better indicators than others. For this reason, it is desirable to use as many of the criteria as possible to determine degree of transport.

In addition, transport can be achieved via two methods. First, fossils can be transported a certain distance laterally in which case the fossil is exotic to its surroundings. This may be accomplished by current activity or severe storm activity. Second, transport can take the form of back-and-forth motion generated by daily tidal or wave activity. If this is the case, the fossil is essentially in the same general location as it lived (*in situ*), but may have undergone appreciable wear (Boucot, 1981).

RESULTS

Two invertebrate fossiliferous horizons are recognized above the Upper Cliff #1 coal within the study area (Figure 2). The lowermost fossiliferous horizon occurs within the thin-bedded gray shale and siltstone lithofacies. Sedimentary structures include flasher to lenticular bedding load structures and small scale ripples. A faunal list is presented in Table 1.

The dominant faunal elements of the shaley-siltstone lithofacies are the bivalve genus *Pteronites*, the trace fossil *Planolites* and the inarticulate brachiopod *Orbiculoidea*. *Orbiculoidea* are most common in the basal portions of the lithofacies and

Table 1. *Faunal Content and Mode of Occurrence in the Shaley-Siltstone Lithofacies*

FAUNA	MODE OF OCCURRENCE	
BRACHIOPODA		
Orbiculoidea	IS	(LP,AV-DV, UF-F, AC)
Lingula	IS	(LP,AV,AC)
MOLLUSCA		
Pteronites	IS	(LP,AV, NDA, UF, APO)
Wilkingia	IS	(LP, AV, NDA, UF, APO)
Aviculopecten	IS-T	(AV-DV, DA-NDA)
Paleyoldia	IS	(AV, UF)
Nucula	T	(DV)
Prothyris	IS	(AV, NDA, APO, UF)
Edmondia	IS-T	(AV-DV, DA)
Phestia	IS	(AV, NDA, UF)
Dunbarella	IS-T	(AV-DV, NDA)
Euphemites	?	(NDA)
TRACE FOSSILS		
Rhizocorallium	IS	
Zoophycos	IS	
Scalarituba	IS	
Chondrites	IS	
Kouphichnium	IS	
Asterophycos	IS	
Planolites	IS	

T - Transported
IS - *In Situ*
LP - Living Position
AV - Articulated Valves
DV - Disarticulated Valves
DON-Disarticulated Valves Not Preserved
DA - Differential Abrasion
NDA-No Differential Abrasion
F - Fragmented
UF - Unfragmented
SS - Size Sorting
APO -Abiotically Preferred Orientation
AC - Authigenically Cemented

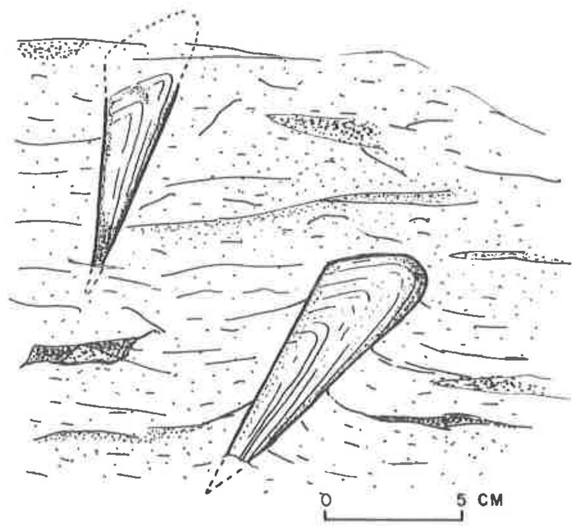

Figure 3. *Sketch of* Pteronites *in living position. The specimen to the right shows inclination due to sediment compaction.*

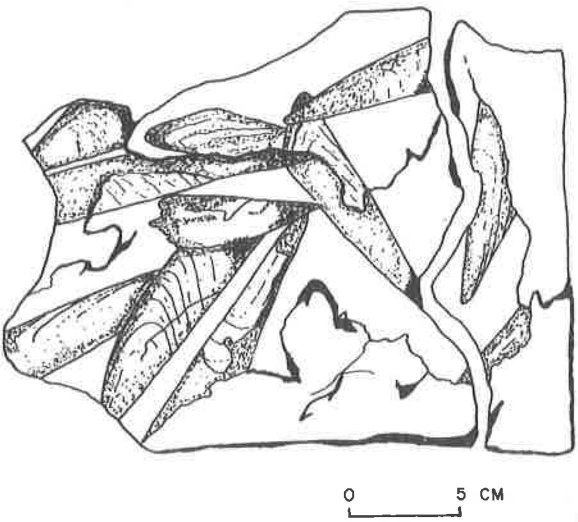

Figure 4. *Bedding surface showing* Pteronites *thanatocoenosis. Note the apparent alignment of long axis.*

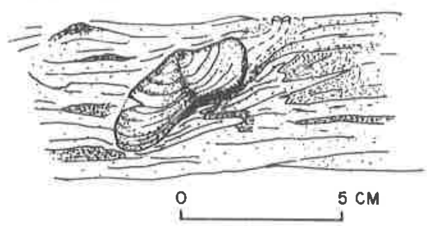

Figure 5. Wilkingia *in living position showing distortion of bedding.*

are commonly within siderite nodules for which they serve as the nuclei for authigenic cementation. *Orbiculoidea* preserved in this manner are often articulated although crushed or distorted.

The most common genus of body fossil in the shaley-siltstone lithofacies is *Pteronites*. It is usually preserved in one of two orientations. *Pteronites* preserved essentially in living position are very common (Figure 3). Most individuals are preserved inclined at an angle of approximately 30° to bedding. Hoare and Sturgeon (1972) note similar occurrences from the Pennsylvaniaan of Ohio. They conclude this inclination to be a result of compaction of the surrounding sediment. Distorted bedding around inclined *Pteronites* within the study area are consistent with this interpretation.

The second preserved orientation for *Pteronites* is in masses oriented parallel to bedding (Figure 4). Although no measurements of orientation were taken, field observation suggested that *Pteronites* preserved in this manner are aligned with respect to their longest axis. Hoare, Sturgeon, and Kindt (1979) describe similar accumulations of *Pteronites* from the Brush Creek Shale (Pennsylvanian, Ohio) and conclude that such accumulations represent a thanatocenosis. *Pteronites* were probably a delicate form. The lack of breakage and differential abrasion of the shells, and the oriented, articulated nature of the fossils indicate that the *Pteronites* death assemblage did not undergo appreciable lateral transport and occur essentially *in situ*.

Articulation of valves provides the next best indication of additional *in situ* fossils in the Upper Cliff interval (Table 1). For example, the genus *Wilkingia* is almost always found articulated either with its valves closed (Figure 5) preserving its posterior gaepor with its valves open indicating exposure or excavation after death (Figure 6). *Wilkingia* is often found in living position as well. Other articulated bivalve genera are commonly preserved with either the valves open or closed including *Aviculoptecten*, *Paleyoldia*, and *Porthyris*. All of the identified trace fossils (Table 1) represent *in situ* occurrences.

Table 2. Faunal Content and Mode of Occurrence in the Sandstone Lithofacies

FAUNA	MODE OF OCCURRENCE	
BRACHIOPODA		
Orbiculoidea	IS-T	(AV-DV, UF-F)
Schizophoria	T	(DV, DA, SS, APO)
Pustula	T	(DV, DON, DA, APO)
Rugosochonetes	T	(DV, DON)
Linoproductus	T	(DV, DON)
Derbyia	T	(DV)
MOLLUSCA		
Pteronites	IS-T	(AV-DV)
Aviculopecten	T	(DV, DA, APO)
Wilkingia	IS-T	(LP, AV-DV, APO)
Dunbarella	T	(DV)
Astartella	T	(DV)
Septimyalina	T	(DV)
Edmondia	T	(DV, APO)
Nuculopsis	T	(DV)
Phestia	T	(DV)
Euphemites	?	
Straparollus	T	(SS, APO)
Palaeostylus	?	
Bellerophon	?	
TRACE FOSSILS		
Zoophycos	IS	
Rhizocorallium	IS	

T - Transported
IS - *In Situ*
LP - Living Position
AV - Articulated Valves
DV - Disarticulated Valves
DON-Disarticulated Valves Not Preserved
DA - Differential Abrasion
NDA-No Differential Abrasion
F - Fragmented
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SS - Size Sorting
APO-Abiotically Preferred Orientation
AC - Authigenically Cemented

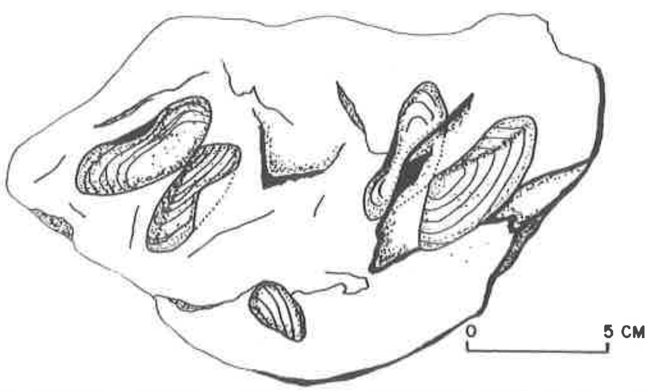

Figure 6. Bedding surface showing articulated Wilkingia *with open valves indicating excavation after death without appreciable transport.*

The second fossiliferous horizon occurs within the basal portions of the overlying sandstone lithofacies. This horizon is 6-18 centimeters thick, thins to the northeast, and is capped by a ubiquitous occurrence of the trace fossil *Zoophycos*. Invertebrate taxa and *in situ* and transported faunal characters preserved in this horizon are presented in Table 2.

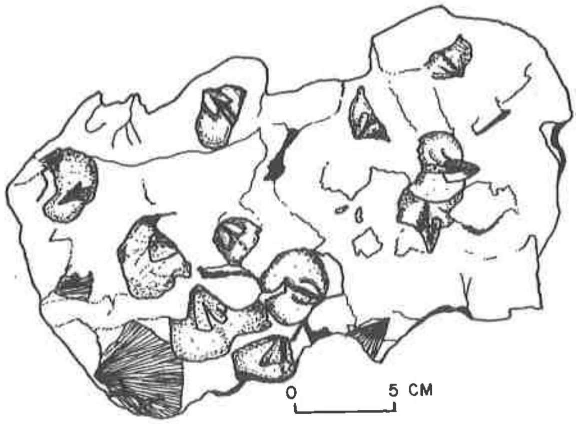

Figure 7. Bedding surface of Schizophoria *zone showing internal molds of disarticulated* Schizophoria *and fragmented bivalves.*

The dominant invertebrate preserved within this basal fossiliferous shell-bed is the orthid brachiopod *Schizophoria*. *Schizophoria* occurs at all of the studied localities at the same stratigraphic position. This zone is named the *Schizophora* zone and serves as a marker bed locally (Gibson, 1983).

Brachiopods and bivalves are preserved together in great abundance within the *Schizophoria* zone (Figure 7). Hoare, Sturgeon, and Kindt (1979) note similar accumulations from the Pennsylvanian of Ohio concluding that they represent a thanatocoenosis. Disarticulated *Schizophoria*, size sorting, alignment of elongate gastropods, and the lack of infaunal components within the *Schizophoria* zone also support this interpretation (Table 2).

Schizophoria are preserved in variable conditions. Most commonly, they occur disarticulated with one of the valves more dominant in any particular place. Since brachiopod musculature tends to keep brachiopod valves closed upon burial (Ager, 1967), the numerous disarticulated valves suggest some degree of reworking or transport. The apparent sorting of valves lends additional support to this conclusion.

Pustula, Rugosochonetes, Linoproductus and *Derbyia* are also preserved with their valves disarticulated. With the exception of *Derbyia*, these brachiopods are spinose forms. Only on one specimen of *Pustula* is there any evidence of preservation of spines. Only the basal portions of these spines are preserved near the hingeline.

The molluscan fauna also suggests reworking and transport. *Pteronites* occurs within the *Schizophoria* zone with valves disarticulated and articulated. *Aviculopecten* and *Wilkingia* also occur with both articulated and disarticulated valves. A few *Wilkingia* are preserved in living position but most are oriented horizontal to bedding. All of the other bivalves are generally preserved disarticulated.

Among the gastropods, only the genus *Straparollus* offers evidence of the *in situ* or transported nature of the *Schizophoria* zone. *Straparollus* is commonly found in masses horizontal to bedding. All of the individuals are roughly equal in size suggesting size sorting. A noticeable orientation with respect to a slightly elongate axis indicates some abiotic orientation.

DISCUSSION

The occurrence and preservational nature of invertebrates associated with the Upper Cliff coal interval is significant. First, the basic taxonomic descriptions can be undertaken (Gibson and Gastaldo, in preparation). To date no such published taxonomy exists for Pennsylvanian invertebrates or Alabama. Second, the *in situ* nature of the shaley-siltstone fauna will allow paleoecological studies to be undertaken and the environment of deposition to be assessed accurately (Gibson, 1983).

The *Schizophoria* zone is interpreted to represent a transported assemblage that contains some *in situ* elements such as *Pteronites, Wilkingia* and *Orbiculoidea* (Table 2). The abrupt sedimentological change from the underlying shaley-siltstone to sandstone with scouring supports the transported nature of the *Schizophoria* zone. Attributes of the transported fauna can offer information concerning the ecological habitat of the *Schizophoria* zone fauna. This fauna appears to represent an offshore-marine environment of deposition as the source area for the *Schizophoria* zone fauna (Gibson, 1983).

The *Schizophoria* zone can be traced at all of the measured localities. Within the study area, this zone serves as a good marker bed. At present, the total geographic extent of the *Schizophoria* zone is unknown. Further investigation is planned to determine if this zone can be traced throughout other areas.

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LITHOSTRATIGRAPHY, DEPOSITIONAL ENVIRONMENT, AND SEQUENCE FRAMEWORK OF THE MIDDLE EOCENE SANTEE LIMESTONE, SOUTH CAROLINA COASTAL PLAIN

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ABSTRACT

The middle Eocene Santee Limestone of South Carolina consists of three laterally equivalent lithofacies in its area of outcrop and subcrop: a bryozoan biomicrudite/biosparrudite, a foraminiferal biomicrite (Chapel Branch Member, new name), and a molluscan-mold terrigenous mudstone (Caw Caw Member). Lateral facies relationships of these lithostratigraphic units reflect a carbonate ramp depositional setting composed of the following facies belts: deep subtidal (35-50 m, open marine, middle shelf) bryozoan biomicrudites; shallow subtidal to intertidal (0-35 m, open marine, inner shelf) shoaling upward bryozoan biosparrudites; shallow lagoonal (0-10 m, semi-restricted) foraminiferal biomicrites (Chapel Branch Member); and shallow lagoonal (0-10 m, semi-restricted) molluscan terrigenous mudstones (Caw Caw Member).

The Santee Limestone represents a carbonate highstand supersequence bounded by Type 1 unconformities which are also supercycle boundaries. This supersequence is furthermore separated into three sequences by a Type 2 unconformity at the base of the *Cubitostrea sellaeformis* zone which resulted from a minor fall in sea level, and by a major non-depositional marine hiatus within the *C. sellaeformis* zone associated with a rapid increase in the rate of relative sea level rise during the latter part of the eustatic middle Eocene highstand (Tb supercycle).

INTRODUCTION

Renewed interest in the Eocene formations of the Atlantic Coastal Plain has led to a series of recent articles concerning the stratigraphy of the middle Eocene Santee Limestone of South Carolina (Banks, 1977; Ward and others, 1979; Baum and others, 1980; Kier, 1980; Powell and Baum, 1981; 1982). These studies have generally focused on stratigraphic or paleontologic relationships between several large quarries which lie in the downdip subcrop of the Santee Limestone and/or the overlying Cross and Cooper formations. Detailed petrologic study of the Santee has been limited to the works of Banks (1977), which presented the lithostratigraphy of a partial section of the Santee exposed at the Martin Marietta Berkeley quarry, and Powell and Baum (1981) which presented the lithostratigraphy of a quarry site core at the Martin Marietta Georgetown quarry.

The lateral distribution of these units exposed in quarries and/or cores, and that of equivalent updip units which have limited exposure in outcrop, has received little attention since the earlier interpretations of Sloan (1908), Cooke (1936), and Pooser (1965). In light of recent stratigraphic revisions, a more detailed investigation of the lateral distribution and lithostratigraphic relationships of these units was undertaken.

This study presents the results of a comprehensive lithostratigraphic analysis of the middle Eocene (Claibornian) Santee Limestone. The report is an assimilation of surface and shallow subsurface mapping utilizing lithostratigraphy, biostratigraphy, and petrography to delineate the vertical and lateral occurrence and distribution of internal lithostratigraphic units throughout the area of outcrop and subcrop of the Santee Limestone. Particular emphasis has been given to the physical and chronostratigraphic relationship between these lithostratigraphic units in an attempt to present a comprehensive depositional model and sequence framework for the Santee Limestone.

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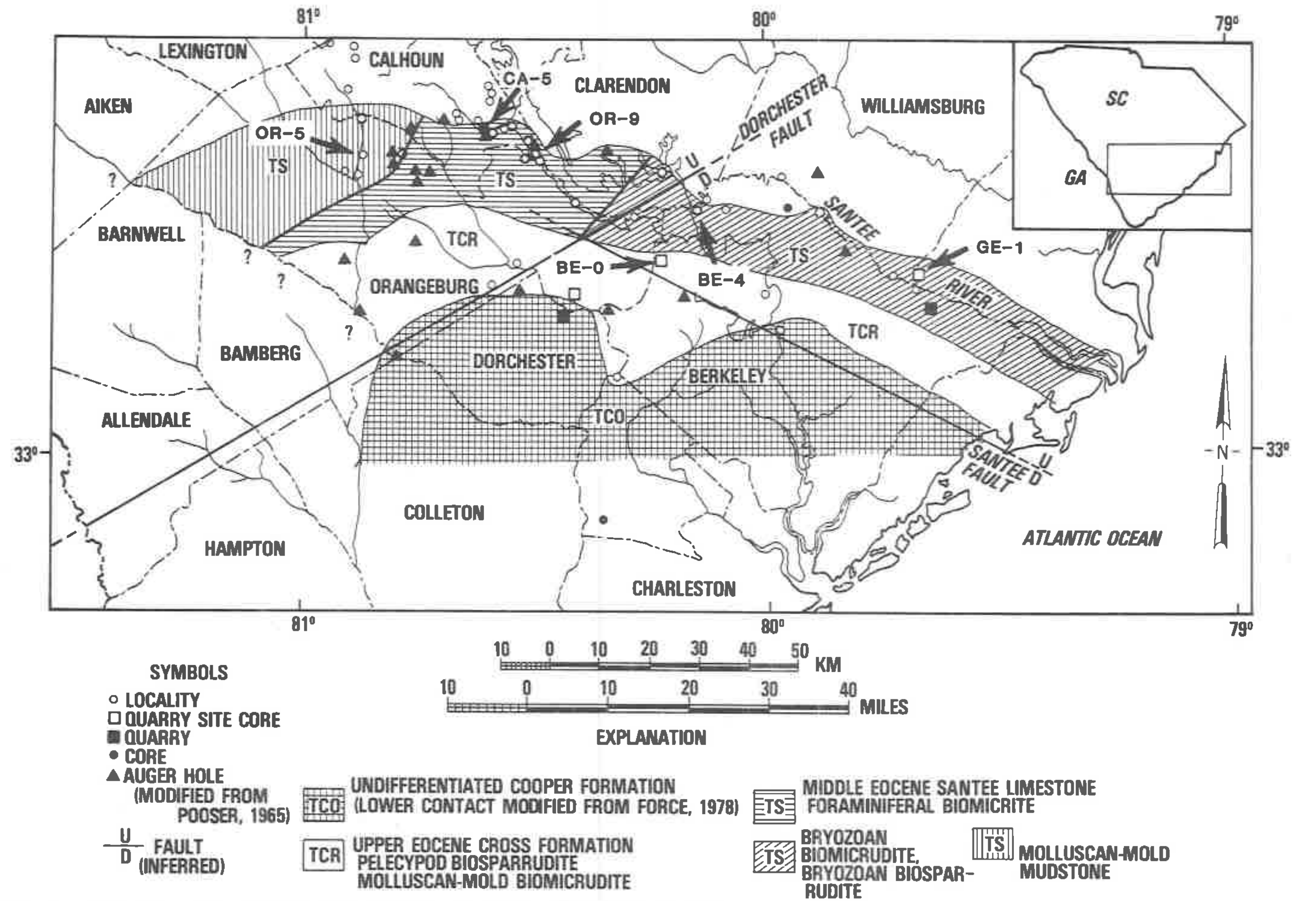


Figure 1. Outcrop and subcrop distribution of units discussed in text. All contacts are approximate. Geology after Baum and Powell (1979), Powell and others (1981), Powell (1981), and Powell and Baum (1982).

Previous Studies

Charles Lyell (1845) was the first to apply the name "Santee" to limestones exposed along the Santee River in South Carolina. Later, Toumey (1848) studied the distribution of the various "marls" of the South Carolina Coastal Plain, primarily from a paleontologic standpoint.

Sloan (1908), in the first comprehensive report on the geology of the South Carolina Coastal Plain, named many of the Cretaceous-Tertiary formations, including the Santee and Mt. Hope "marls". Cooke (1936) formalized nomenclature for these units and combined the Santee and Mt. Hope into one formation, the Santee Limestone. Cooke and MacNeil (1952) later revised this nomenclature and framework, suggesting that most of the Eocene units were separated by unconformities. In this paper, Cooke and MacNeil recognized the equivalence of updip terrigenous units containing *Pteropsis lapidosa* (referred to as the "McBean Formation") to downdip units of the Santee Limestone characterized by *Cubitostrea sellaeformis* (Conrad). This equivalence had earlier been recognized by Sloan (1908, p. 457-458) who referred to the updip unit characterized by *Pteropsis lapidosa* (Conrad) as the "Caw Caw Shale".

Pooser (1965) and Colquhoun and others (1969) presented modifications to Cooke and MacNeil's stratigraphic framework, and suggested that the Santee Limestone and several associated updip units (Congaree Formation, Warley Hill Formation, "McBean Formation", Barnwell Formation) were facies equivalents rather than being discrete rock units bounded by unconformities.

The Warley Hill Marl (Cooke and MacNeil, 1952), or Formation (Pooser, 1965), was erected by Cooke and MacNeil (1952) for strata characterized by *Cubitostrea lisbonensis* (Harris). However, subsequent studies (Banks, 1977; Baum and others, 1980) have shown that the carbonates containing *C. lisbonensis* are the lithostratigraphic equivalent of the Santee Limestone. Thus, Baum and others (1980) and Powell and Baum (1982) referred these beds to the *C. lisbonensis* zone of the Santee Limestone.

Banks (1977) studied the stratigraphy of three downdip quarries and recognized two major lithologic units in the Santee Limestone: the lower bryozoan limestones (= type Santee Limestone of Cooke, 1936) and the upper molluscan-foraminiferal limestones. Based on subsurface information, Baum and Powell (1979) and Ward and others (1979) recognized that these two units were separated by an unconformity. Thus, Ward and others (1979) proposed the name "Cross Member" for the upper limestone unit and "Moultrie Member" for part of the lower bryozoan unit (*Cubitostrea sellaeformis* zone). Baum and others (1980) subsequently presented evidence for a distinct lithostratigraphic and chronostratigraphic break between the Cross Member and the lower bryozoan limestones. Baum and others (1980) thus proposed the name "Cross Formation" for the upper unit and retained the name "Santee Limestone" for the bryozoan limestones exposed at the type section originally described by Cooke (1936).

Kier (1980), in a study of echinoids of the Santee Limestone, recognized two distinct echinoid zones: the lower *Protoscutella mississippiensis* - *Santeelampas oviformis* zone (= *C. lisbonensis* zone, = Warley Hill Formation of Cooke and MacNeil, 1952; Kier, 1980) and the upper *Linthia harmatuki* zone (= *C. sellaeformis* zone).

The Santee Limestone as defined by Baum and others (1980) is middle Eocene (Claibornian) in age based on the occurrence of the oysters *Cubitostrea lisbonensis* (Harris) and *Cubitostrea sellaeformis* (Conrad), well-known guide fossils of the Gulf Coastal Plain (Stenzel, 1949). The Santee is equivalent to the Lisbon Formation of Alabama and to the Castle Hayne Limestone of North Carolina as defined by Baum and others (1978), or to the Comfort and New Hanover members of the Castle Hayne Limestone as defined by Ward and others (1978) (see Powell and Baum, 1982).

LITHOSTRATIGRAPHY AND BIOSTRATIGRAPHY

Stratigraphic terminology for this report follows that of Baum and others (1980) and Powell and Baum (1982). The middle Eocene (Lutetian-Bartonian) Santee Limestone is generally considered to consist of bryozoan-dominated carbonates (biosparrudites, biomicrudites) (Fig. 1; Baum and others 1980, Banks, 1977; Cooke, 1936; Cooke and MacNeil, 1952). The typical bryozoan lithologies of the Santee have been found in this study to grade updip into a foraminiferal biomicrite and a molluscan mudstone. These

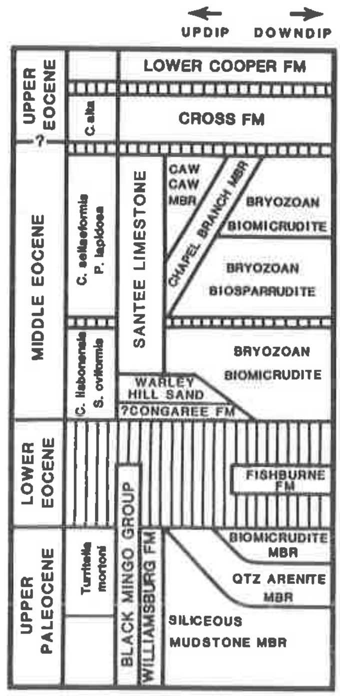


Figure 2. Stratigraphic framework of the upper Paleocene-Eocene formations along the western flank of the Cape Fear arch in South Carolina.

two units are mappable on a scale of several counties and are biostratigraphically equivalent to bryozoan-dominated units of the upper Santee Limestone (Fig. 2). The molluscan mudstone has recently been given formal status of Caw Caw Member of the Santee Limestone (Powell and Baum, 1982). The foraminiferal biomicrite is introduced in this report as the Chapel Branch Member of the Santee Limestone.

The Santee Limestone was deposited during one supercycle of marine onlap (Tb of Vail and others, 1977) which can be subdivided into two distinct biostratigraphic zones in its area of outcrop and subcrop: the lower *Cubitostrea lisbonensis - Santeelampas oviformis* zone, and the upper *Cubitostrea sellaeformis - Pteropsis lapidosa* zone (Fig. 2) (Powell and Baum, 1982; Banks, 1977; Cooke and MacNeil, 1952). These zones can be recognized by the presence of other faunal elements as discussed by Powell and Baum (1982). For example, *Protoscutella mississippiensis* (Twitchell), a species restricted to the *C. lisbonensis* zone and the older *Cubitostrea perplicata* zone of the Gulf Coastal Plain (Toulmin, 1977), is restricted to the *C. lisbonensis* zone of the Santee Limestone (see Kier, 1980; Powell and Baum, 1982). Similarly, *Protoscutella conradi* (Cotteau) is apparently restricted to the *C. sellaeformis* zone (Powell and Baum, 1982; see Kier, 1980). Both the *C. lisbonensis - S. oviformis* zone and the *C. sellaeformis - P. lapidosa* zone can be recognized in the downdip bryozoan facies of the Santee; whereas the foraminiferal biomicrite and molluscan mudstone apparently represent only the *C. sellaeformis - P. lapidosa* zone.

The bryozoan facies of the Santee Limestone can be subdivided into three respective lithofacies: a lower bryozoan biomicrudite, a middle bryozoan biosparrudite, and an upper bryozoan biomicrudite. The lower biomicrudite represents the *C. lisbonensis - S. oviformis* zone, whereas the latter two lithofacies represent the *C. sellaeformis - P. lapidosa* zone. The lowermost *C. sellaeformis* zone beds unconformably overly *C. lisbonensis* zone beds, and beds representing the intermediate zone, *C. smithvillensis* (Stenzel, 1949), are apparently absent in the study area (Banks, 1977; Powell and Baum, 1982).

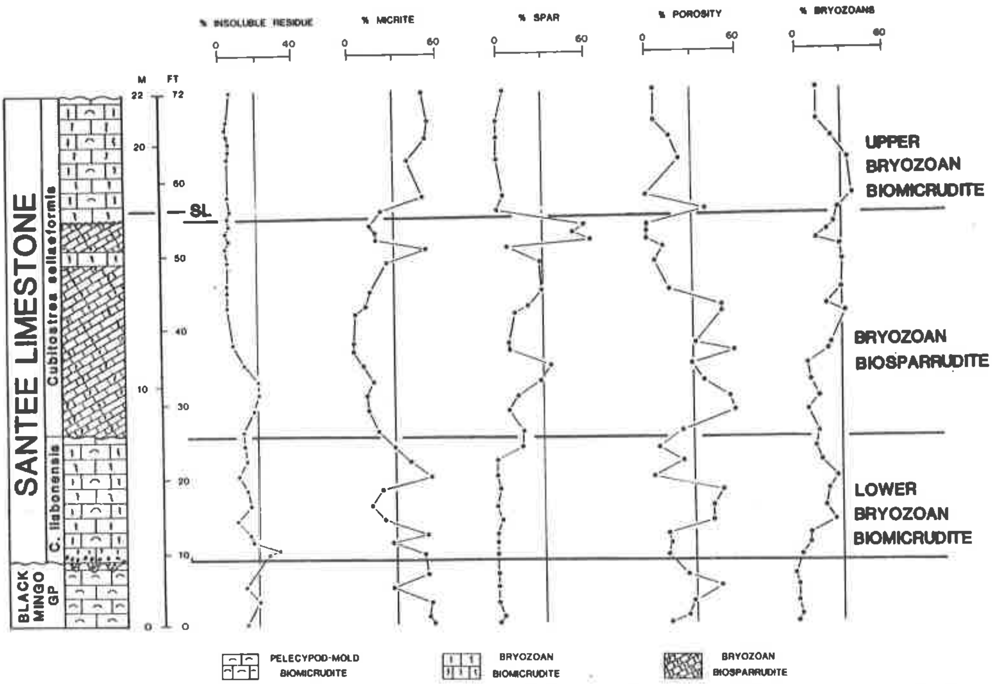


Figure 3. Summary of point count and insoluble residue data for bryozoan facies of the Santee Limestone taken from core GE-1.

Bryozoan Facies

The lithofacies relationships within the bryozoan-dominated lithologies of the Santee Limestone have been studied in detail using petrography of cores BE-O and GE-1, quarry exposures, and natural exposures (Figs. 1, 3). The type section of the Santee Limestone of Cooke (1936) and the neostratotype section of the Santee Limestone of Ward and others (1979) (see Appendix) lie within the outcrop and subcrop belt of the bryozoan facies.

Lower Bryozoan Biomicrudite: This lithofacies consists of pale yellow (10 YR 8/2) to pale gray (N7) matrix-supported bryozoans, oysters, echinoids, and foraminifera (Table 1). The unit disconformably overlies the pelecypod-mold biomicrudite member of the Williamsburg Formation of Van Nieuwenhuise and Colquhoun (1982) or "Thanetian Black Mingo Formation" of Powell and Baum (1981) (loc BE-O, GE-1; Fig. 4).
The contact is a mesokarstic bored surface commonly coated with phosphate and/or glauconite. The lower 2 feet (0.6 m) of the unit consists of a phosphate pebble biomicrudite, which appears to be a basal lag deposit associated with the unconformity. Thin beds of biosparrudite occur within the unit but are not common and grade vertically and laterally into biomicrudite.

The lower bryozoan biomicrudite represents the *Cubitostrea lisbonensis - Santeelampas oviformis* zone. The overall faunal elements of this unit are quite similar to the other bryozoan carbonate lithologies of the Santee, excluding several diagnostic forms of the *C. lisbonensis - S. oviformis* zone. In outcrop the unit is dominated by *Chlamys*, *Eburneopecten scintillatus* (Conrad), *Ostrea*, *Solena*, Cardiidae, *Protoscutella mississippiensis* (Twitchell), *C. lisbonensis* (Harris), and *Santeelampas oviformis* Cooke.

The unit is known from Berkeley County (core BE-0; loc. BE-4) and Georgetown County (core GE-1). The glauconitic limestone of Sloan's (1908) Warley Hill "Marl" probably represents a lithostratigraphic equivalent in Calhoun County (Sloan, 1908; sur. no. 699; Cave Hall; see Cooke and MacNeil, 1952). The unit is well-exposed in outcrop at loc. BE-4 (see Appendix), and is thickest in core GE-1, where it is 17 ft (5.2 m).

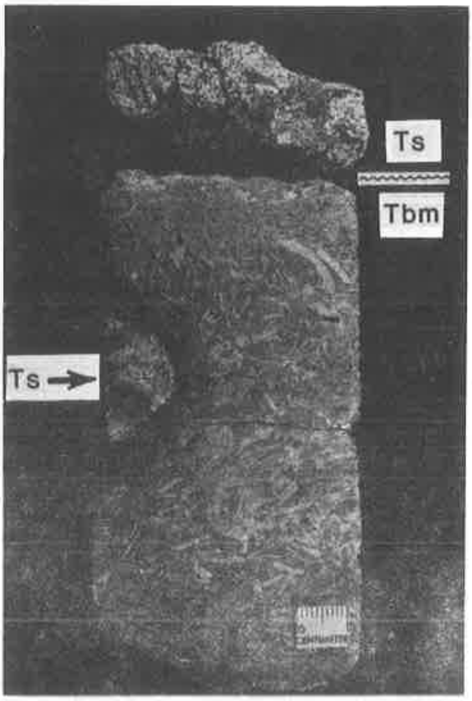


Figure 4. Disconformity between upper Paleocene Black Mingo Group-Williamsburg Formation (Tbm, biomicrudite member) and middle Eocene Santee Limestone (Ts). Note mesokarstic nature of contact. Core BE-O.

Stratigraphic relationships suggest the Warley Hill Sand (= Warley Hill at the type section of Cooke and MacNeil, 1952; Pooser, 1965) is an updip equivalent of the lower bryozoan biomicrudite; the Congaree Formation in its type area may also be an updip equivalent of the lower Santee (see Pooser, 1965, Colquhoun and others, 1969).

Petrography. The framework of this lithofacies is primarily bryozoans (Table 1; Fig. 3; 5B). Other common allochems include pelecypods, echinoids, and foraminifera. Most of the pelecypods are oysters, *Cubitostrea lisbonensis* (Harris). Red algal fragments are present in small quantities throughout the unit (Fig. 5D). There is a strong inverse relationship between micrite content and porosity (Fig. 3). Aragonitic molluscs within this lithofacies are present as molds which are commonly solution enlarged. Glauconite is present as peloids and within intraparticle areas of allochems, particularly echinoids and foraminifers.

The terrigenous content of this lithofacies is low except at the base of the unit where there is an increase in fine quartz sand associated with the underlying Santee/Black Mingo unconformity (Fig. 3, 5A). Also associated with the unconformity is an increase in glauconite pellets and phosphatized pebbles of the underlying Black Mingo Group. Various stages of phosphatization and glauconitization can be seen in the pebbles. The majority of these pebbles range from 3 to 5 mm in diameter. Along the disconformity pyrite is present as fine disseminated flakes.

Vuggy porosity is the dominant pore type in this lithofacies (Fig. 5C). Interparticle porosity, when present, is usually solution enlarged, as is the small amount of moldic porosity. Within this facies interparticle porosity is commonly seen grading to solution-enlarged interparticle porosity and finally grading to vuggy porosity (Powell and Baum, 1981). Cementation is low in the unit, leaving the mean porosity value for the lithofacies relatively high (Fig. 3).

Middle Bryozoan Biosparrudite: This lithofacies is composed of pale gray (N7) to

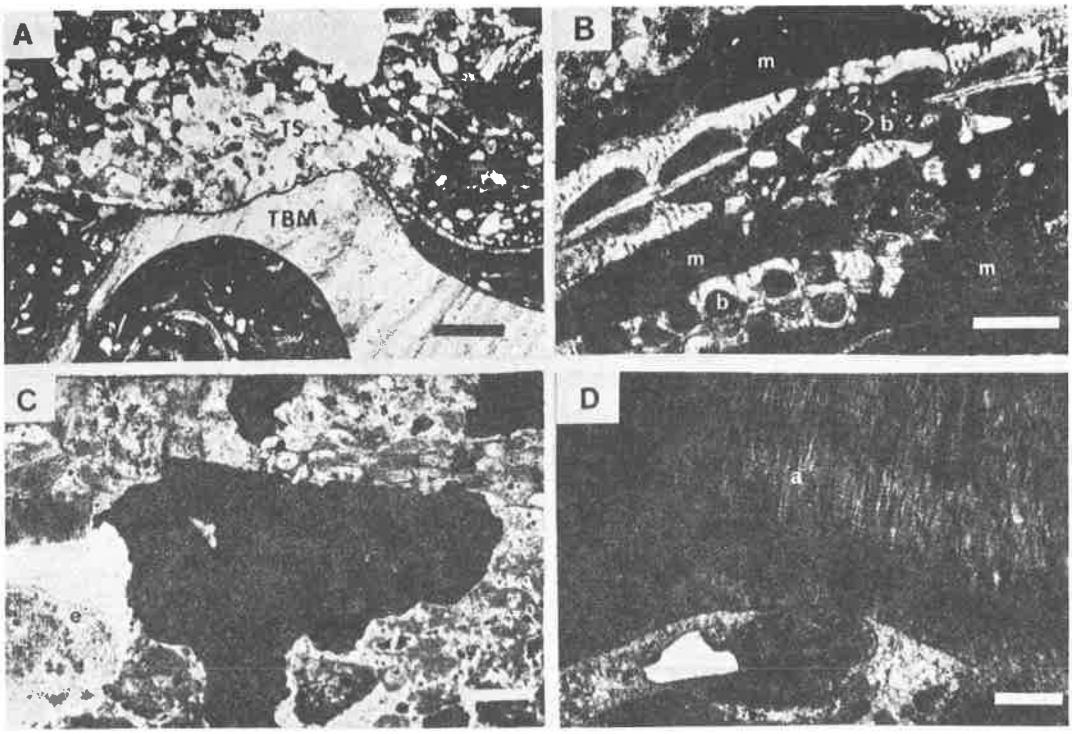


Figure 5. Photomicrographs of the lower bryozoan biomicrudite lithofacies of the Santee Limestone (scale of A,C, and D = 1.0 mm; scale of B = 0.5 mm; all plain light except C).
A. Photomicrograph of disconformity between Santee Limestone (Ts) and Black Mingo Group (Tbm). Note pholad borings into oyster shell.
B. Bryozoans (b) in a micrite matrix (m).
C. Non-fabric selective vuggy porosity. Note solution of bryozoan along top of vug and echinoid (e) with syntaxial calcite overgrowth.
D. Large red algal fragment (a).

yellowish gray (5Y 8/1) grain-supported bryozoans, echinoids, and pelecypods. At localities GE-1 and BE-0 this lithofacies is separated from the lower bryozoan biomicrudite by a distinct unconformity. Insoluble residue increases at the base of the unit and apparently is associated with the unconformity (Fig. 3). Solution pits along the unconformity are coated by isopachous phosphate which clearly originated as a direct precipitate (Fig. 6D). The unconformity apparently represents a hiatus spanning the *Cubitostrea smithvillensis* zone (Banks, 1977; Powell and Baum, 1982).

Excluding the many species of bryozoans, common faunal elements include *Eburneopecten scintillatus* (Conrad), *Chlamys*, *Protoscutella conradi* (Cotteau), *Democrinus*, *Recurvaster*, *C. sellaeformis* (Conrad), and echinoid spines. *Democrinus*, *Recurvaster*, and echinoid spines are often abraded, suggesting transport. Most of the faunal elements, along with the bryozoans, are small.

The unit commonly displays low-angle and moderate-angle planar cross-bedding (Fig. 6A). It varies from tightly cemented to weakly cemented and locally may be friable. At locality GE-1 the basal and upper portions of this lithofacies are dense, whereas the middle portions are unconsolidated to weakly cemented. Large intraclasts resembling mud polygons are present sporadically in the unit (Fig. 6B).

There is a paucity of outcrop of the bryozoan biosparrudite. It is well developed in Georgetown County but has been observed only in core and quarry sections (loc. GE-1). In Berkeley County (loc. BE-0), the lithofacies becomes more micritic and is 17.5 ft (5.3 m) thick. In Georgetown County (loc. GE-1), the unit is 28 ft (8.5 m) thick and more pure, containing fewer beds of biomicrudite.

Table 1. Mean point count and insoluble residue analysis for the carbonate facies of the Santee Limestone.

INSOLUBLE RESIDUE ANALYSIS (WEIGHT %)	LOWER BRYOZOAN BIOMICRUDITE Mean (%)	LOWER BRYOZOAN BIOMICRUDITE Standard Deviation (%)	BRYOZOAN BIOSPARRUDITE Mean (%)	BRYOZOAN BIOSPARRUDITE Standard Deviation (%)	UPPER BRYOZOAN BIOMICRUDITE Mean (%)	UPPER BRYOZOAN BIOMICRUDITE Standard Deviation (%)	FORAMINIFERAL BIOMICRITE Mean (%)	FORAMINIFERAL BIOMICRITE Standard Deviation (%)
Carbonate	85.4	5.3	92.1	6.9	95.7	1.4	80.9	2.3
Sand Size	8.6	3.0	5.4	5.5	0.4	0.5	7.9	2.1
Silt and Clay Size	6.0	2.9	2.5	1.4	3.9	1.6	11.2	1.2

POINT COUNT DATA (N=43)

	LOWER BRYOZOAN BIOMICRUDITE subtotal	LOWER BRYOZOAN BIOMICRUDITE total	BRYOZOAN BIOSPARRUDITE subtotal	BRYOZOAN BIOSPARRUDITE total	UPPER BRYOZOAN BIOMICRUDITE subtotal	UPPER BRYOZOAN BIOMICRUDITE total	FORAMINIFERAL BIOMICRITE subtotal	FORAMINIFERAL BIOMICRITE total
Mean Total Rock Porosity		25.85%		31.71%		14.2 %		8.1 %
ALLOCHEMS								
Bryozoans		18.5		20.4		28.7		T
Shell	13.1		13.4		12.7			
Intraparticle micrite	3.3		3.4		11.1			
Intraparticle porosity	1.2		0.9		1.2			
Glauconite	0.7		0.9		0.9			
Intraparticle cement	0.2		1.5		1.1			
Intraparticle microspar	—		0.2		—			
Phosphate	—		0.1		1.7			
Moldic porosity	T		—		T			
Echinoids		4.3		5.2		2.5		4.2
Foraminifers		0.7		0.4		2.4		16.7
Shell	0.5		0.2		T		6.4	
Intraparticle micrite	0.2		0.1		1.5		3.3	
Intraparticle porosity	—		0.1		0.5		5.3	
Intraparticle cement	—		—		0.3		1.7	
Glauconite	—		—		0.1		T	
Red Algae		1.0		T		1.3		T
Corals		T		0.4		T		—
Intraparticle micrite			0.1					
Intraparticle cement			0.2					
Shell			0.1					
Serpulids		0.2		0.3		T		—
Shell	0.2		0.2					
Intraparticle porosity	—		0.1					
Gastropods		T		0.1		T		T
Moldic porosity			0.1					
Pelecypods		4.5		2.9		3.6		3.7
Shell	3.7		1.6		2.1		3.0	
Moldic porosity	0.4		0.2		1.0		0.7	
Moldic cement	0.2		0.9		0.3		—	
Intraparticle micrite	0.2		T		0.1		T	
Ostracods		T		—		0.3		3.5
Pellets		T		T		T		3.9
Unknown		0.3		0.4		0.3		T
Intraparticle porosity	0.1		—		0.2			
Glauconite	0.2		—		0.1			
Shell	—		0.3		—			
Intraparticle cement	—		0.1		—			
Intraparticle micrite	—		—		—			
Brachiopods		T		T		T		T
Oolites		T		T		T		—
Intraclasts		T		T		0.3		—
Crinoids		0.1		T		T		—
Diatoms		—		—		—		1.0
TERRIGENOUS		8.5		3.5		0.6		6.0
Quartz	3.6		1.9		0.3		6.0	
Detrital phosphate	0.06		1.3		0.3		—	
Heavy minerals	0.2		0.3		0.2		T	
Phosphate pebbles	4.1		T		—		—	
MATRIX		34.6		11.3		48.0		48.3
Micrite	34.5		8.3		47.7		48.3	
Microspar	0.1		3.0		0.3		—	
POROSITY		23.2		30.1		8.5		6.9
Vuggy	22.5		17.4		7.5		—	
Interparticle	0.6		11.7		0.8		6.9	
Intercrystalline	0.1		1.0		0.2		—	
ORTHOCHEMICAL		4.3		25.1		3.5		3.3
Glauconite	1.8		0.7		0.1		1.4	
Interparticle spar cement	2.5		24.4		3.4		1.9	
TOTAL		100.2		100.1		100.6		99.5

Petrography. The biosparrudite is dominated by bryozoans; locally echinoids and pelecypods may be abundant (Table 1, Fig. 3 and 7A). The majority of the allochems are relatively small and commonly are rounded or abraded. The majority of the pelecypods are calcitic forms, such as *Chlamys*, *Eburneopecten*, and *Plicatula*. The unit contains varying amounts of micrite, but in general the content is low. Much of the micrite present is internal sediment (see James and others, 1976; Banks, 1977). The secondary micrite is recognized when allochems are surrounded by a thin isopachous rim cement and interparticle areas are filled with micrite (Fig. 7B, 8B). This rim cement is similar to that described from beachrock (Bricker, 1971). Another type of secondary micrite is found in the upper portion of this lithofacies (Fig. 8A). This type postdates the precipitation of thick, medium to coarse calcite rim cement (Fig. 8A). Glauconite is present in peloidal form. Micrite envelopes are common in this unit, particularly surrounding dissolved aragonitic forms. Intraclasts, oolites, and thin layers of pelsparite can be seen sporadically within the unit.

The basal portions of the unit show a marked increase in insoluble residue as well as a fine monocrystalline quartz associated with the basal diastemic surface (Fig. 3). Partially phosphatized and glauconitized pebbles are present, but are not nearly as large or as phosphatized as those at the base of the lower bryozoan biomicrudite. The unit contains various pore types (Fig. 7B, C). Vuggy, solution-enlarged interparticle, and interparticle porosity dominate. Most of the vuggy porosity is due to solution of interparticle porosity as the two often intergrade. In general, the porosity of the unit

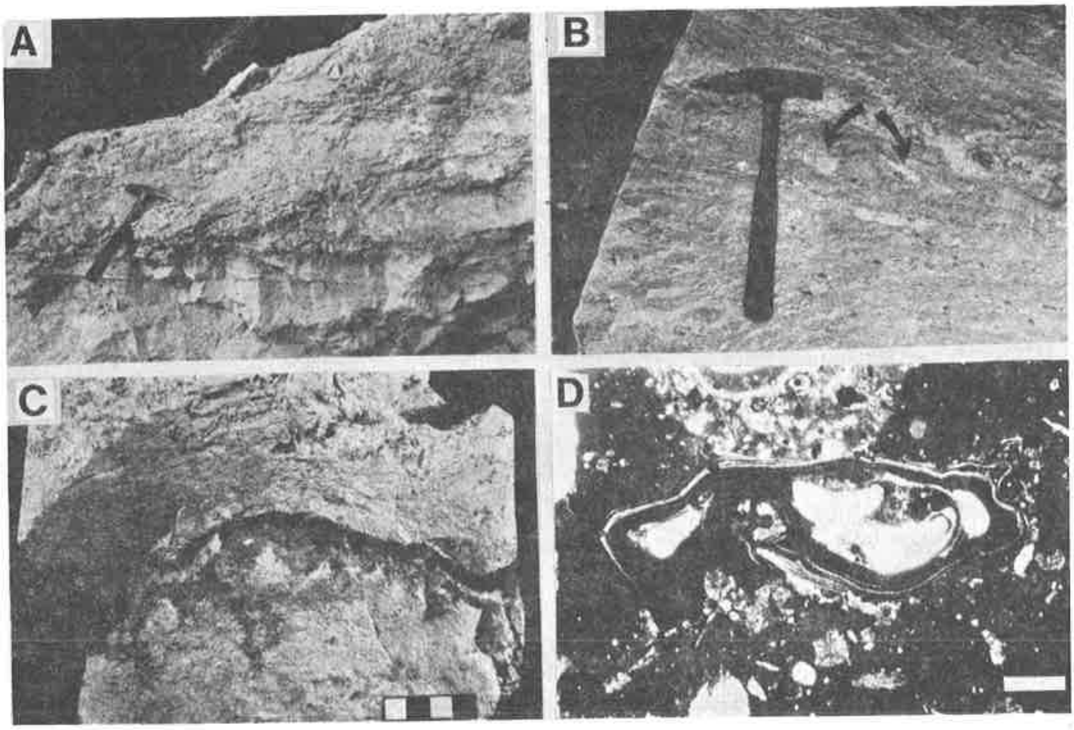


Figure 6. Photographs (A,B,C) and photomicrograph (D) of bryozoan biosparrudite lithofacies of the Santee Limestone.
A. Moderate angle planar crossbedding (hammer is scale).
B. Large intraclasts in bryozoan biosparrudite (hammer is scale).
C. Diastem separating upper bryozoan biomicrudite from bryozoan biosparrudite (loc. GE-1). Note mesokarstic nature of contact (scale graduated in cm).
D. Void-filling phosphate precipitate in solution pit along diastemic surface at base of biosparrudite (plain light, scale = 1.0 mm).

is governed by the amount of void filling which has taken place in any given bed. All stages of cementation can be observed in the unit. Acicular isopachous submarine cement (see Bathurst, 1976; Longman, 1980) commonly predates clear mosaic meteoric cement (Fig. 7D).

At locality GE-1 the biosparrudite is in sharp diastemic contact with the overlying upper bryozoan biomicrudite (Fig. 6C). Beneath the diastem, the biosparrudite is cemented by coarse, isopachous, and somewhat "dirty" spar (Fig. 8A). Spar-reduced porosity has been filled by micrite (Fig. 8A) which is probably associated with the overlying bryozoan biomicrudite. The isopachous nature of this spar clearly indicates a phreatic origin, and the presence of internal micrite sediment suggests cementation occurred within the depositional environment. Moreover, the "dirty" nature of the spar suggests this is a submarine cement. Paleoecologic data indicates the overlying biomicrudite is a deeper water lithosome. Thus, the diastem separating the biosparrudite from the upper biomicrudite is considered to be a non-depositional marine hiatus, perhaps associated with a cessation of, or tremendously reduced, sedimentation rate (condensed or starved interval).

Upper Bryozoan Biomicrudite: This unit is a dense to poorly consolidated, pale gray (N7) to buff yellow (10 YR 8/6), matrix-supported biomicrudite which diastemically overlies the bryozoan biosparrudite lithofacies (Fig. 6C).

The fauna of this unit is extremely diverse and has been studied by Harbison (1944) as well as by C. Wythe Cooke, F. Stearns MacNeil, Gilbert D. Harris, and

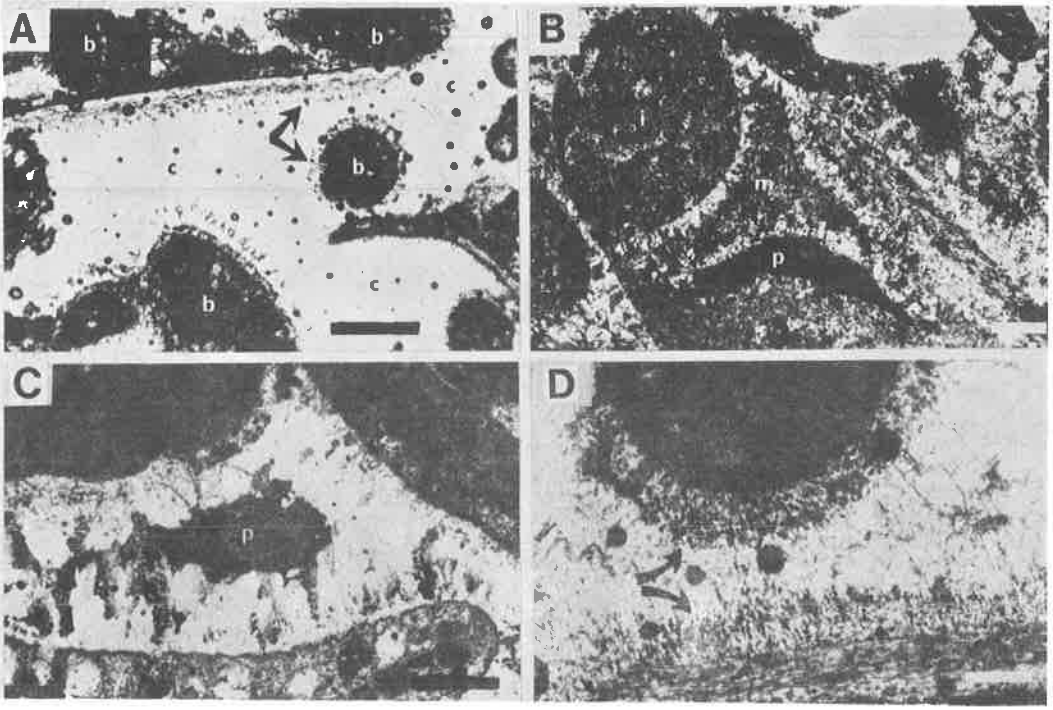


Figure 7. Photomicrographs of bryozoan biosparrudite lithofacies (scale of A and B = 0.5 mm, C = 0.1 mm, D = 0.05 mm; B and C--crossed nicols, A and D--plain light).
A. Bryozoans cemented by clear, coarse meteoric calcite spar (c).
B. Isopachous acicular submarine cement with internal micrite sediment (m) filling porosity; note moldic porosity (p) and intraclast (i).
C. Interparticle porosity (p) reduced by coarse isopachous meteoric calcite spar.
D. Closeup of A. Spar-filled porosity. Note early acicular submarine cement (arrows).

Charles Lyell in their many early collections from Eutaw Springs, Orangeburg County. The overwhelming majority of the fauna studied in this report was collected from the upper bryozoan biomicrudite (for detailed report see Powell and Baum, 1982). In addition to the many bryozoa, the unit is dominated by *Terebratulina, Probolarina, Crassatella cf. C. texalta* (Harris), *Crassatellites, Solena, Semele,* Cardiidae and a both abundant and diverse echinoid fauna. At the Martin Marietta Georgetown Quarry (GE-1), very large specimens of *Aturia alabamensis* Morton are common; however, at the Martin Marietta Berkeley Quarry (BE-0) these large cephalopods are rare. Similarly, *Cubitostrea sellaeformis* (Conrad), very abundant at BE-0, is not present at locality GE-1.

This unit is widely distributed throughout the outcrop and subcrop of the bryozoan carbonate facies of the Santee Limestone (Fig. 1). It is approximately 20 feet (6 m) thick and does not appear to thicken or thin appreciably other than towards its updip pinchout. The contact between this unit and its updip equivalent, the foraminiferal biomicrite, cannot be seen in outcrop; however, the facies transition occurs abruptly between localities in Eutawville and Vance, South Carolina.

Petrography. The upper biomicrudite is dominated by micrite matrix and bryozoans (Fig. 3; Table 1). Other common allochems include echinoids, pelecypods, and foraminifera. Pelecypods in the lithofacies were usually aragonitic and have been dissolved leaving moldic porosity (Fig. 8D). Glauconite and pyrite are orthochemical components, commonly occuring in intraparticle chambers of echinoids, bryozoans, and forams. Red algae are present sporadically within the unit (Fig. 9A). Locally, globorotalids may be abundant (Fig. 9A). The terrigenous content of the upper bryozoan biomicrudite is very low (Fig. 3) generally consisting of fine to very fine

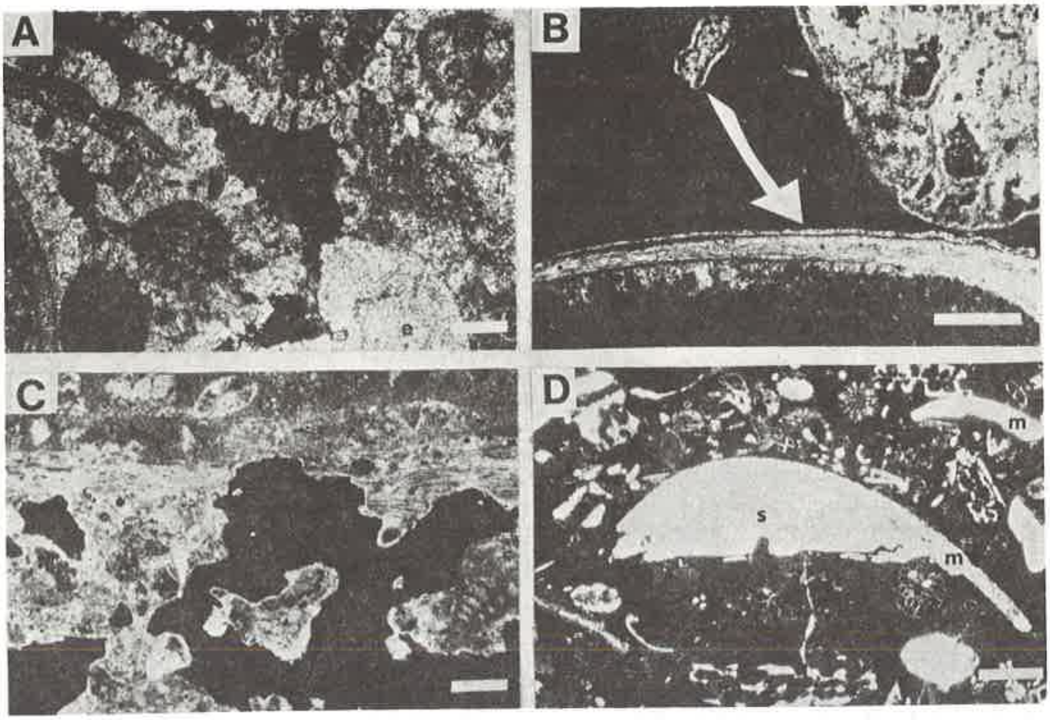


Figure 8. Photomicrographs of bryozoan biosparrudite lithofacies (A,B) and bryozoan biomicrudite lithofacies (C,D) (scale of A and C = 0.2 mm, B = 0.5 mm, D = 1.0 mm; all crossed nicols except D).
A. Internal micrite sediment filling precursor pore space after coarse, "dirty", isopachous fringe cement.
B. Internal micrite sediment after isopachous fringe cement.
C. Vuggy porosity.
D. Bryozoans and moldic pelecypods in micrite, note moldic porosity (m) and shelter porosity (s).

monocrystalline quartz and heavy minerals. Phosphate is present in minor amounts as fine pellets.

At locality GE-1, this unit is dense and varies little lithologically above the diastem discussed earlier. The only major difference is that planktonic foraminifera *(Globorotalia)* content is much higher in the upper part of the section, which would appear to indicate a deepening of the basin. Winnowed zones are common throughout the unit.

Vuggy porosity is the most common pore type (Fig. 8C); locally, however, moldic and shelter porosity may contribute significantly to the total pore content (Fig. 8D). The amount of void-filling by sparry calcite is very low throughout the unit (Fig. 3).

Chapel Branch Member

Stratigraphic Revision: Just north of Eutawville, Orangeburg County, the Santee Limestone changes abruptly from the typical bryozoan carbonates of the Santee to foraminiferal biomicrites (Fig. 1) which contain only trace amounts of bryozoans (Table 1). The equivalence of this biomicrite to typical Santee lithologies is demonstrated by the occurrence of the index fossils *Cubitostrea sellaeformis* and *Protoscutella conradi* in both units (see Powell and Baum, 1982). Thus, it is proposed that this biomicrite be given formal member status in the Santee Limestone, and be referred to as the Chapel Branch Member. The name is derived from Chapel Branch, a stream near locality OR-9. This locality is proposed as a type section for the member. At this

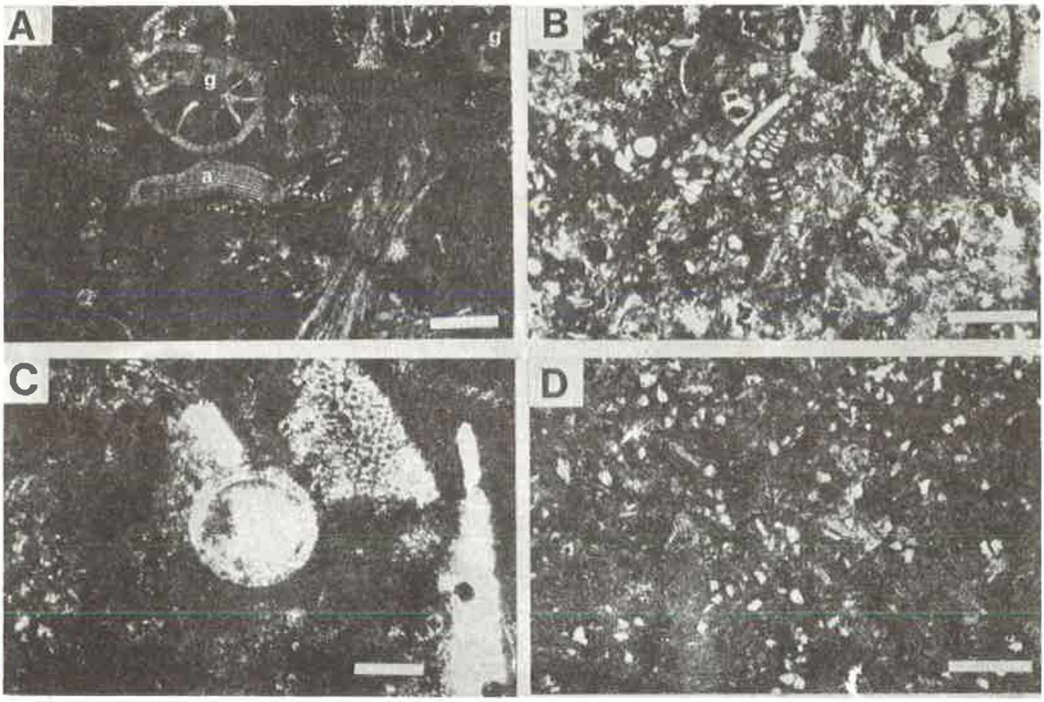


Figure 9. Photomicrographs of upper bryozoan biomicrudite (A), foraminiferal biomicrite (B,C,D) (scale of A = 1.0 mm, scale of B,D = 0.5 mm, C = 0.1 mm; all plain light except A,D).
A. Bryozoans, red algae (a), and globorotalids (g) from uppermost portion of lithofacies.
B. Benthonic (biserial) foraminifers in a micrite matrix.
C. Possible charophyte oogonium from foraminiferal biomicrite.
D. Arenaceous foraminiferal biomicrite.

section, *Pteropsis lapidosa* is abundant, and *C. sellaeformis* can occasionally be found. An auxiliary reference section is locality CA-5, where *C. sellaeformis* occurs in abundance (see Appendix for measured sections).

The Chapel Branch Member of the Santee Limestone basically conforms to Sloan's (1908) description of the "Santee Marl" (Sloan, 1908, sur. nos. 349, 696, 697, 699, 701, 705, 706, 707); whereas, his description of the "Mt. Hope Marl" generally conforms to the typical bryozoan carbonates of the Santee Limestone (Sloan, 1908, sur. nos. 713, 715, 720) (see Cooke, 1936, p. 75).

Distribution and Occurrence: The foraminiferal biomicrite lithofacies, or Chapel Branch Member, consists of yellowish gray (5Y 7/2) matrix-supported foraminifera, ostracods, echinoids, and molluscs. Excluding its abundant microfauna it is relatively unfossiliferous, but locally contains molluscs of genera *Lucina*, *Chlamys*, *Venericardia*, *Ostrea*, *Cubitostrea*, and *Pteropsis*. The sparsity of megainvertebrates, combined with the virtual absence of bryozoans, distinguish this unit from the very fossiliferous bryozoan facies of the Santee Limestone, and allow megascopic differentiation of the two units in the field. Echinoid spines, often *Cidaris pratti* (Clark), are the most common megascopic allochem. Many of the allochems are abraded near the units southernmost outcrop near Vance in Orangeburg County along the south shore of Lake Marion (Fig. 1).

The unit is extremely homogenous and varies only in its fine quartz silt content and its fauna. In eastern Orangeburg County the quartz content reaches 15%; however, in Calhoun County the detrital content is less than 5%. Similarly, in eastern

Orangeburg County the unit is characterized by *Pteropsis lapidosa*, *Lucina*, and *Venericardia*. In Calhoun County *Cubitostrea sellaeformis* dominates the fauna.

The Chapel Branch Member crops out along the southern shore of Lake Marion in Orangeburg and Calhoun counties. From this area the outcrop belt swings westward into Calhoun and northern Orangeburg counties where it grades updip into the Caw Caw Member. The greatest exposed thickness is 20 ft (6 m) at locality CA-5. The unit stratigraphically underlies the Caw Caw Member in auger hole sections near its updip limit (see Appendix). The lithologic transition to both the Caw Caw and the bryozoan facies of the Santee is extremely rapid and mimics the trend of the inferred Dorchester fault.

Petrography: The dominant allochem of this lithofacies is foraminifera (Table 1). The only other common allochems are ostracods and echinoderms. These foraminifera are primarily benthonic in nature; the only planktonic forms represented are rare globigerinids (Fig. 9B). Benthonic species include uniserial and biserial forms as well as nummulitids, miliolids, rotalids, quinqueloculines, and peneroplids. The majority of the foraminifera have been micritized (Fig. 9B). Diatoms and possible charophyte oogonia (Fig. 9C) are present sporadically throughout the unit. Bryozoans are rare, which is in great contrast to the previously discussed lithofacies of the Santee.

The matrix of the unit is micrite which usually supports the allochems; locally, however, foraminifers are so abundant that they support the rock. The allochems, even the sparse pelecypod component, are all very small (Fig. 9B). Locally, the micrite is peloidal. The terrigenous content of the lithofacies is relatively low, but is higher than that of the bryozoan facies of the Santee (Fig. 9D). In places, the rock contains large amounts of very fine monocrystalline subangular quartz (loc. OR-9); elsewhere it is clean of terrigenous quartz. The clay-sized fraction of the insoluble residue dominates. Glauconite commonly replaces fine sediment within the chambers of foraminifers.

Aragonitic allochems now occur as molds, though these are not abundant. Where the unit contains larger molluscs, the rock is well-indurated due to precipitation of low magnesian calcite derived from dissolution of the aragonitic shell material.

Caw Caw Member

Stratigraphic Revision: The name "Caw Caw", introduced by Sloan (1908, p. 457-58) is used for a siliceous molluscan-mold mudstone facies which is a biostratigraphic equivalent of the upper Santee Limestone (*Pteropsis lapidosa* - *Cubitostrea sellaeformis* zone). This unit has recently been given the formal status of Caw Caw Member of the Santee Limestone (Powell and Baum, 1982). It has previously been assigned to the McBean Formation, which has a type section in Georgia (Cooke and MacNeil, 1952; Pooser, 1965; Colquhoun and others, 1969). However, the unit bears no lithologic resemblance to the true "McBean" lithostratigraphic unit exposed at the type section of Veatch and Stephenson (1911). Recent studies (Huddlestun and others, 1979) have shown that clastic units originally included in the upper type section of the McBean are actually part of the upper Eocene Dry Branch Formation of the Barnwell Group (Huddlestun, 1982). The McBean Formation of current usage is restricted to the carbonate units (often containing *C. sellaeformis*) at the type section. Thus, the Caw Caw Member is not the lithostratigraphic equivalent of the McBean Formation of current usage (Huddlestun, 1982), nor is it the lithostratigraphic equivalent of sands of the Dry Branch Formation which were originally included in the type McBean by Veatch and Stephenson (1911). Therefore, the name McBean is inappropriate for the molluscan mudstones referred to the Caw Caw Member in this study.

Lithology and Distribution: The Caw Caw Member is a dull yellowish gray (5Y 7/2) moldic mudstone characterized by the small pelecypod *Pteropsis lapidosa* (Conrad). The outcrop belt lies updip of the Chapel Branch Member in northern Orangeburg County (Fig. 1). This unit is the "conspicuous component...that appears to be a marl from which all the lime mud has been leached" (Cooke and MacNeil, 1952). It was described by Cooke and MacNeil as well as Cooke (1936) as a light-weight sandstone. Actually, the rock is composed of clay and silt-sized material and may be an altered diatomite quite similar to the altered diatomites of the Congaree Formation and Black Mingo Group (see Weaver and Wise, 1974; Wise and Weaver, 1973).

The Caw Caw Member is quite homogenous and varies only in its glauconite content, quantity of megainvertebrates, and state of fossil preservation. Lithologies assigned to the Caw Caw Member may grade updip into non-marine to marginal marine sands and kaolinitic clays referred to the middle Eocene by Willoughby (in preparation, personal communication, 1984). The type section of the Caw Caw Member (loc. OR-5, see appendix) consists of 16 ft (5 m) of molluscan mudstone that becomes slightly glauconitic towards the base. Pockets of silicified molluscs are reported by Pooser (1965) from this unit but they were not observed in this study.

The Caw Caw crops out northwest of the Orangeburg escarpment. Its lower contact with the Chapel Branch Member runs through the Orangeburg vicinity and from there swings to the southwest toward the Southeast Georgia embayment. Its maximum thickness is approximately 30 ft (9 m).

The fauna of this unit has been studied by C. Wythe Cooke and T. W. Vaughn (in Cooke, 1936). It has an abundant and diverse molluscan fauna in its type area (loc. OR-5; see Cooke, 1936, p. 62-64). In addition to *Pteropsis lapidosa* (Conrad), molluscs such as *Turritella, Tellina,* and *Venericardia* are the most common faunal components.

DEPOSITIONAL ENVIRONMENT

Bryozoan Facies

The depositional environment of the bryozoan facies of the Santee Limestone has been previously discussed in Banks (1977). Also, Baum (1980) has discussed in detail the depositional environment of equivalent bryozoan lithosomes in the Castle Hayne Limestone of North Carolina. Conclusions of this study agree quite well with these works and are discussed in more detail in Powell (1981).

The bryozoan biosparrudite lithosome represents a shallow, above wave base intertidal to shallow subtidal (0-35 m) environment based on the combined presence of moderate angle planar crossbedding and large micrite intraclasts (mud polygons). The grainstone texture of the unit is consistent with a shallow water position. This unit basically corresponds to Wilson's (1975) facies belt 6 (standard microfacies 11, 12).

The bryozoan biomicrudite lithosome represents a deeper, below wave base (>35 m) middle shelf environment generally within the photic zone. This conclusion is based on the abundance of numerous eschariform and lunulitiform bryozoans, branching red algae, and a relatively shallow shelf (35-50 m) ostracod fauna (Sharon Lyon, personal communication, 1981). In addition, the wackestone-packstone texture of the unit indicates a position below effective wave base. The overall faunal diversity of the unit (see Powell and Baum, 1982), particularly in the echinoids (Kier, 1980), suggests a stable, open marine shelf, perhaps situated in a subtropical to tropical biogeographic setting. This unit corresponds to Wilson's (1975) facies belt 2 (standard microfacies 8).

Chapel Branch and Caw Caw Member

The recognition of a foraminiferal biomicrite lithosome (Chapel Branch Member) updip and laterally adjacent to bryozoan lithosomes of the Santee allows for a more complete description of the depositional framework of this formation. The prevalence of micritized benthonic foraminifera and ostracods in contrast to the diverse faunal assemblage of the downdip laterally adjacent bryozoan lithosome, suggests a semi-restricted condition for the Chapel Branch Member. The wackestone-packstone and locally peloidal texture of the unit is compatible with an environment partially restricted from effective marine circulation. Presumably, the Chapel Branch was deposited in shallow coastal bays and/or lagoons. The foraminiferal biomicrite equates with Wilson's (1975) facies belt 8 (standard microfacies 19).

The Caw Caw Member reflects a similar semi-restricted, shallow marine or perhaps even brackish environment, laterally adjacent to the middle Eocene coastal plain. The absence of stenohaline marine forms (echinoderms, corals, etc.) in the unit

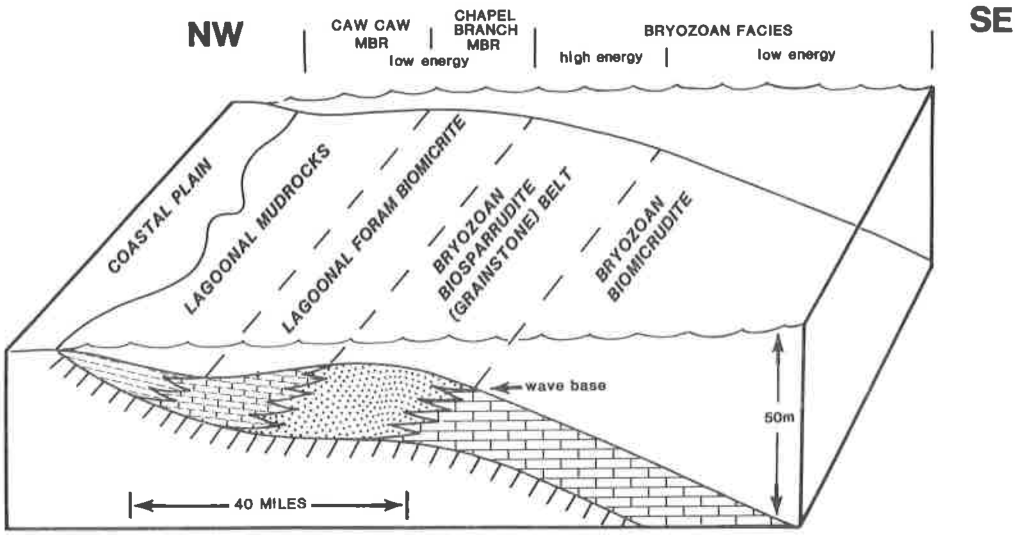


Figure 10. Lateral facies relationships of the Santee Limestone during early Cubitostrea sellaeformis *time.*

suggests brackish conditions, though this could result from an increased terrigenous influence adjacent to the coastal plain. The Caw Caw Member apparently represents an environment of bays or lagoons similar to that of the Chapel Branch Member, but with an increased terrigenous, and possible freshwater, influence.

Depositional Model

The depositional setting of the Santee Limestone appears more closely associated with the carbonate ramp model (Ahr, 1973) than with the carbonate platform model (Wilson, 1975). Reef trends are apparently absent in the Santee, and the regional paleoslope of the deep water facies to shallow water facies is much less than 1°.

The lateral facies relationships in the Santee consist of a deep water bryozoan biomicrudite, a shallow water bryozoan biosparrudite grainstone belt, and a semi-restricted lagoonal belt composed of foraminiferal biomicrites grading updip to terrigenous molluscan mudstones (Fig. 10). The low energy lagoonal belt was most likely created as a result of shoaling of the bryozoan grainstone belt. At times the grainstone belt may have been locally subaerially exposed. A similar carbonate ramp facies association of deep water wackestones, shallow water grainstones, and lagoonal packstones-wackestones has been described from the Jurassic Smackover Formation of Florida (Sigsby, 1976) and the Jurassic Portland Limestone of southern England (Sellwood, 1978, p. 294-295).

A rapid increase in the relative rate of sea level rise resulted in decreased sedimentation with subsequent submarine cementation and hardground development at the top of the shoaling-upward grainstone facies. At sea level maximum, the deeper water bryozoan packstone-wackestone facies was deposited over the hardground-capped (Fig. 6C) bryozoan grainstone belt.

DEPOSITIONAL SEQUENCE FRAMEWORK

An overwhelming volume of literature has accumulated in recent years on seismic stratigraphy, which in many cases emphasizes various aspects of depositional sequence analysis (for example, Vail and others, 1977) and sequence development of passive margins (Pitman, 1978). As discussed in Vail and others, the depositional sequence is the fundamental unit for establishing a comprehensive stratigraphic framework. This concept can be readily applied to seismic studies; however, it can also be applied to subsurface core and well log studies, and surface geology from outcrops. As stated by Vail and others, the physical relationship between lithostratigraphic units is the principle criterion for sequence analysis. In this respect, the surface or subsurface geologist has a major advantage--he can usually "put his hand" on critical physical relationships, and if his area is regional in scope, can develop a sequence framework for the stratigraphic section.

This type of analysis works quite well when applied to the Santee Limestone and associated Eocene units of South Carolina (Fig. 11). The basal unconformity separating the Santee Limestone from older early Tertiary units (Fig. 4) represents the Ta/Tb supercycle boundary of Vail and others (1977). This contact is also considered to be a Type 1 unconformity (Vail and Todd, 1981) representing a supersequence boundary. A Type 1 unconformity is a result of a relative fall in sea level below the shelf edge, exposing the shelf to subaerial erosion and subjecting the deep water basin to concomittant point source type sedimentation (deep sea fans, low stand deltas). Theoretically, a Type 1 unconformity should also represent a major lithostratigraphic and chronostratigraphic break, such as is seen at the Black Mingo/Santee contact and the Santee/Cross contact.

A Type 2 unconformity results from a stillstand or slight drop in sea level with subsequent subaerial erosion in updip areas. Sea level remains above the shelf edge allowing for a conformable relationship in downdip shelfal areas. In general, a Type 2 unconformity should represent only a minor lithostratigraphic break, such as that between the lower bryozoan biomicrudite and bryozoan biosparrudite of the Santee. This erosional surface is known to be a minor unconformity by the absence of the *Cubitostrea smithvillensis* faunal zone which occurs between the *C. lisbonensis* and *C. sellaeformis* zones (see Stenzel, 1949; Powell and Baum, 1982).

Within the Santee supersequence the shallow water facies (biosparrudite) is disconformably overlain by the deeper water facies (upper bryozoan biomicrudite) (Fig. 6C). Available data indicates a rapid relative rise in sea level for this biomicrudite; immediately above the unconformity a relatively deep water environment existed, suggesting the unconformity is a non-depositional marine hiatus rather than a subaerial erosion surface. Thin section studies show the underlying surface to have been extensively cemented prior to deposition of the overlying biomicrudite.

Such extensive submarine cementation is to be expected when an area is subjected to a rapid rise in sea level with concomittant cessation of, or tremendously reduced, sedimentation. This hiatus commonly represents a condensed section (Vail and Todd, 1981; Vail, Hardenbol and Todd, 1982). The juxtaposition of an offshore facies unconformably over nearshore deposits, which is not normally thought to occur in subaerially developed unconformities, is of particular interest.

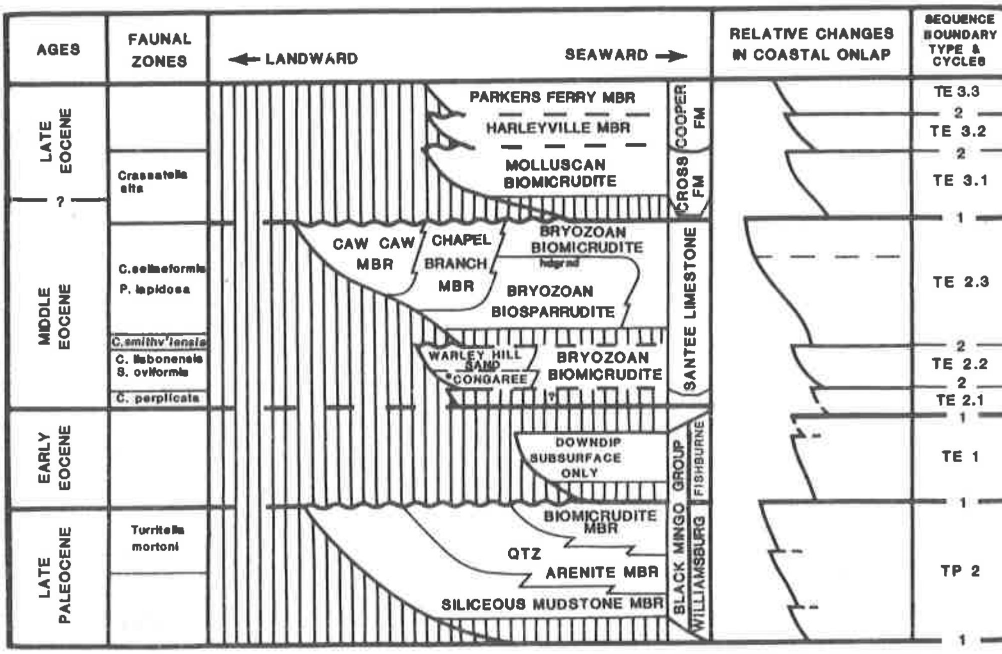
Thus, the Santee Limestone may be interpreted as a supersequence bounded by major lithostratigraphic breaks which are also Type 1 unconformities. Within this supersequence, second order sequences are formed by minor relative changes in coastal onlap (Type 2 unconformity) and sediment response to rapid rise in sea level (condensed interval). This sequence analysis agrees quite well with the overall lithostratigraphic and biostratigraphic analysis, with the equivalent stratigraphic framework of the Gulf Coastal Plain, and with the eustatic model of Vail and Mitchum (1979) and Vail and Hardenbol (1979) (see Powell and Baum, 1982).

SUMMARY

The Lutetian-Bartonian Santee Limestone consists of three regional lithofacies in its area of outcrop and subcrop: a bryozoan biomicrudite/biosparrudite, a foraminiferal biomicrite, and a molluscan-mold mudstone (Caw Caw Member). The foraminiferal biomicrite is a previously unrecognized, distinct, mappable lithostratigraphic unit and is designated the Chapel Branch Member of the Santee Limestone.

The depositional and lithostratigraphic framework of the Santee Limestone is quite similar to the carbonate ramp model: the lateral facies relationships in the Santee consist of a deep water (35-50 m) open shelf bryozoan biomicrudite, a shoaling upward shallow subtidal to intertidal (0-35 m) bryozoan biosparrudite, a semi-restricted lagoonal (>10 m) foraminiferal biomicrite (Chapel Branch Member), and a semi-restricted lagoonal (>10 m) terrigenous molluscan mudstone (Caw Caw Member). The overall faunal association within the Santee suggests a subtropical environment.

The Santee Limestone is a highstand supersequence bounded by major regional



* may represent TE 2.1 cycle (Tallahatta equivalent)

Figure 11. Chronostratigraphic section of emerged Coastal Plain of South Carolina with relative changes in coastal onlap (modified from Powell and Baum, 1982; unconformity types after Vail and Todd, 1981). References for age assignments and contact relationships are as follows: Black Mingo Group (Van Nieuwenhuise and Colquhoun, 1982; Hazel and others, 1977; Powell and Baum, 1981; Baum and others, 1979); Santee Limestone (Hazel and others, 1977; Powell and Baum, 1982; Garry Jones, personal communication); Cross Formation (Huddlestun and others, 1974; Huddlestun and Hetrick, 1979; Hazel and others, 1977; Powell and Baum, 1982; Harris and Zullo, 1980; Fullager and others, 1980; Paul Huddlestun, pers. comm., 1984); Cooper Formation (Hazel and others, 1977; Ward and others, 1979). Note: the placement of the middle Eocene-upper Eocene boundary is questioned due to a late middle Eocene placement of the Cross Formation based on nannoplankton by Laurel Bybell (personal communication, 1983). The age of the Cross Formation is problematic, and as stated by Powell and Baum (1984), the age of the basal TE 3.1 cycle may be somewhat older (within NP 17?) than proposed by Vail and Mitchum (1979) and Vail and Hardenbol (1979).

(Type 1) unconformities which bracket the middle Eocene (Lutetian-Bartonian) Tb supercycle of Vail and Hardenbol (1979). A minor (Type 2) unconformity occurs within this sequence at the base of the *Cubitostrea sellaeformis* zone. Within the *C. sellaeformis* cycle, a non-depositional marine hiatus occurs at the contact between the shallow water biosparrudite and the overlying deeper water bryozoan biomicrudite. This diastem is equated with rapid increase in the rate of sea level rise and concommittant sediment starvation in a seaward direction, resulting in a pronounced landward shift in marine onlap.

ACKNOWLEDGEMENTS

I wish to thank Walter H. Wheeler and Daniel A. Textoris (University of North Carolina at Chapel Hill) for reading and criticizing earlier drafts of the manuscript, and for their advice during the study. I am particularly indebted to Gerald R. Baum (Exxon Production Research Company) who shared many ideas on stratigraphic relationships and assisted me during several field excursions as well as in the identification of fauna. Sharon K. Lyon (Amoco Production Company) shared her knowledge of ostracod paleoecology and Garry D. Jones (Union Oil Company) examined several samples for planktonic foraminifers and shared his age determinations. Discussions with Laurel Bybell (U. S. Geological Survey) and Paul Huddlestun (Georgia Geological Survey) were very beneficial.

The study was supported by Sigma Xi and the Smith Fund of the University of North Carolina. Cores and access to quarry operations were kindly given to me by Martin Marietta Company, Santee Portland Company, and the U. S. Army Corps of Engineers.

I am deeply grateful for the constructive criticism of the manuscript given by Robert Mixon (U. S. Geological Survey), Joseph Carter (University of North Carolina, Chapel Hill), William B. Harris and Victor A. Zullo (University of North Carolina, Wilmington), and Ralph Willoughby (South Carolina Geological Survey).

Finally, I would like to thank my wife, Lynn, who not only helped prepare the manuscript, but read it several times and made many suggestions that improved it.

APPENDIX

Selected Measured Sections

Descriptions of all data points (Fig. 1) are given in Powell (1981).

OR-5

Roadcut on U. S. 21, 5.5 miles (8.85 km) north of Orangeburg, just south of Early Branch, a tributary of Caw Caw Swamp. Type section of the Caw Caw Member of the Santee Limestone (Powell and Baum, 1982).

Surficial Soils 3.0 ft (0.9 m)
Unconformity
Santee Limestone
Caw Caw Member 16.0 ft (4.9 m)

1 - porous molluscan-mold mudstone containing *Pteropsis lapidosa, Tellina, Venericardia,* and *Turritella;* buff-yellow grading to light pale green at base.

Remarks: Pooser (1965, auger hole 38-5) reports greenish-yellow limestone 35 feet below the surface here. The limestone is probably the Chapel Branch Member. This locality is the same as Cooke's (1936) loc. 117 (Early Branch). As suggested by Cooke (1936) and Pooser (1965), it may be the same section observed by Sloan (1908) at "Pooser's Hill" (sur. no. 344).

OR-9

South shore of Lake Marion, Santee State Park, access by unnumbered paved road, 1.5 mi (2.41 km) east of crossing of County Roads 105 and 82, 1.5 mi (2.41 km) northwest of south shore of Lake Marion crossing U. S. 301. Type section of Chapel Branch Member of the Santee Limestone.

Surficials 2.0 ft (0.6 m)
Unconformity
Santee Limestone
Chapel Branch Member 7.5 ft (2.3 m)

1 - buff-yellow, slightly glauconitic foraminiferal biomicrite; echinoids, pelecypods; *Pteropsis lapidosa, Cubitostrea sellaeformis, Protoscutella, Cidaris pratti, Recurvaster, Ostrea,* and *Crassatella* cf. *C. texalta;* in general, poorly consolidated but containing numerous, well-lithified nodules due to solution of aragonitic shell material and reprecipitation as cement.

Remarks: This outcrop consists of a long wall over 100 ft (31 m) in length. The type specimens of *Pteropsis lapidosa* (Conrad) and *Ostrea radians* Conrad (= *C. sellaeformis)* are from the Chapel Branch Member of the Santee Limestone at Vances Ferry (Sloan, 1908, sur. no. 707), a locality very near loc. OR-9 (see Palmer and Brann, 1965). Garry Jones (personal communication, 1979) examined planktonic foraminifera from this locality and considers the assemblage to indicate a middle Eocene age.

CA-5

Abandoned limestone quarry 0.8 mi (1.29 km) east of railroad crossing S. C. 6, on the farm of Robert Edwards. Auxiliary reference section for Chapel Branch Member.

Surficials 4.5 ft (1.4 m)
Unconformity
Santee Limestone
Chapel Branch Member 20.0 ft (6.1 m)
1 - buff-yellow poorly fossiliferous foraminiferal biomicrite; Cubitostrea *sellaeformis* abundant.

Remarks: Garry Jones (personal communication, 1979) examined planktonic foraminifera from this locality and considers the assemblage to indicate a middle Eocene age. This locality appears to be the same as Sloan's (1908) sur. no. 697 where he reports 16.0 ft (5 m) of Santee "Marl".

BE-O

Martin Marietta Berkeley Quarry 1.5 mi (2.4 km) south of the intersection of S. C. 6 and County Road 59. Composite section of quarry and quarry site core (for detail see Baum and others, 1980, Banks, 1977). Neostratotype (redefined) of the Santee Limestone (see remarks). Type section of the Cross Formation (Ward and others, 1979; Baum and others, 1980).

Cross Formation 3.0 ft (0.9 m)
5 - buff-yellow molluscan-mold foraminiferal biomicrudite.
Unconformity
Santee Limestone
4 - creamy white to dark gray bryozoan biomicrudite containing C. *sellaeformis*. 20.5 ft (6.2 m)
Diastem
3 - light gray to dark gray poorly cemented bryozoan biosparrudite with some interbeds of biomicrudite containing C. *sellaeformis*. 17.5 ft (5.3 m)
Unconformity
2 - creamy to dark gray poorly sorted bryozoan biomicrudite containing C. *lisbonensis* in that part of the bed which is exposed in the quarry. 16.0 ft (4.9 m)
Unconformity
Black Mingo Group (Williamsburg Formation) 23.0 ft (7.0 m)
1 - dark gray-green pelecypod-mold biomicrudite.

Remarks: Base of quarry section occurs 3 ft (1 m) below the top of bed 2. The quarry section was selected by Ward and others (1979) as the neostratotype of the Santee Limestone. Ward and others (1979) report a 23 ft (7 m) quarry section for the Santee and Cross formations (Cross Member of Ward and others, 1979); however, Banks (1977) and Baum and others (1980) report 47.5 ft (14.5 m) and 48.0 ft (14.6 m) respectively for this interval exposed in the quarry. Faunal components of beds 2, 3, and 4 are discussed in Powell and Baum (1982).

BE-4

Wilson's Landing, Santee River, northern Berkeley County, just below Santee Dam. Reference section for lower Santee Limestone.

Santee Limestone 10.5 ft (3.2 m)
1 - very glauconitic buff yellow bryozoan biomicrudite; abundant *Santeelampas*

oviformis.

Remarks: Garry Jones (personal communication, 1979) examined planktonic foraminifera and considers the assemblage to indicate a middle Eocene age.

GE-1

Martin Marietta Georgetown Quarry, 3.0 mi (4.8 km) NNW of Jamestown, just west of S. C. 41. Reference section for Santee Limestone (see Fig. 4 for detail). Composite section of quarry and quarry site core.

Santee Limestone	
4 - buff-yellow bryozoan biomicrudite.	19.0 ft (5.6 m)
Diastem	
3 - light gray well-cemented to unconsolidated bryozoan biosparrudite.	28.5 ft (8.8 m)
Unconformity	
2 - light gray to buff-yellow bryozoan biomicrudite.	17.0 ft (5.0 m)
Unconformity	
Black Mingo Group (Williamsburg Formation)	11.0 ft (3.4 m)
1 - dark gray-green pelecypod mold biomicrudite.	

Remarks: Ward and others (1979) referred bed 4 to the Cross Formation (Member of authors). See Baum and others (1980) for discussion. Base of quarry occurs 7 ft (2.1 m) above the top of bed 2. Faunal components of beds 3 and 4 are discussed in Powell and Baum (1982).

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STRATIGRAPHY, DEPOSITIONAL ENVIRONMENTS AND REGIONAL DOLOMITIZATION OF THE BRASSFIELD FORMATION (LLANDOVERIAN) IN EAST-CENTRAL KENTUCKY

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ABSTRACT

Early Silurian paleogeography in east-central Kentucky is reconstructed on the basis of stratigraphic, sedimentologic and paleoecologic analysis of the Brassfield Formation (Early-Middle Llandovery). The Brassfield in east-central Kentucky, is composed predominantly of biogenic dolostones but can be traced northward along the outcrop belt on the eastern flank of the Cincinnati Arch into a predominantly limestone facies. Brassfield deposition took place on a shallow carbonate platform that dipped gently eastward into the subsiding Appalachian Basin. Widely correlatable lithofacies represent lagoonal, shoal and shallow open-marine environments that migrated westwardly as the Silurian seas transgressed onto the Cincinnati Arch region. Cyclic sea-level fluctuations during Brassfield time resulted in periodic exposure causing local nondeposition and erosion of some Brassfield units. These sea-level fluctuations were either influenced by local uplift on the "proto-Cincinnati Arch", which had begun by Early Silurian time, or by regional eustatic changes caused by glaciation in the southern hemisphere. Uplift and exposure was greatest southwest of Bath County, Kentucky, where the Brassfield thins considerably. Shoaling dominated such high areas during submergent periods when sea-levels were low, whereas erosion probably took place during extremely low-water periods. Rising sea-level following local regressive interludes resulted in an onshore to offshore pattern of depositional environments and communities typical of Early Paleozoic carbonate platform settings.

Sea-level fluctuation during and after Brassfield time may also explain the regional dolomitization of the unit. Dolomitization probably took place in a fresh water-sea water mixing zone that moved laterally with transgression and regression near exposed areas on the proto-arch. Shifting carbonate sand bodies atop exposed areas would have acted as conduits for freshwater. The greater vertical and lateral extent of dolomitization near high areas may be related to the cumulative effects of multiple mixing zones accompanying cyclic transgressions and regressions in the area.

Early Silurian transgression in the Cincinnati Arch region culminated in deposition of the deeper-water shales of the Crab Orchard Formation (Middle-Late Llandovery). Additional minor transgressive and regressive events during Crab Orchard time reflect continuing sea-level fluctuations throughout the region during all of Early Silurian time.

INTRODUCTION

The Brassfield Formation (Early-Middle Llandovery) is widely recognized in the Cincinnati Arch region as a distinctive stratigraphic unit marking the base of the Silurian system. In east-central Kentucky where the type section is located (Foerste, 1906), the Brassfield consists of a series of predominantly carbonate and shale lithofacies which can be traced along most of the narrow Silurian outcrop belt on the eastern flank of the Cincinnati Arch; these lithofacies reflect deposition approximately parallel to the regional depositional strike (Figure 1). Regional biostratigraphic and lithostratigraphic studies have demonstrated that the Brassfield transgressively onlaps Upper Ordovician strata from southeast to northwest in the Cincinnati Arch region (Rexroad and others, 1965; O'Donnell, 1967; Berry and Boucot, 1970; Boucot, 1975), and both petrographic and geochemical data have shown that the Brassfield changes rather

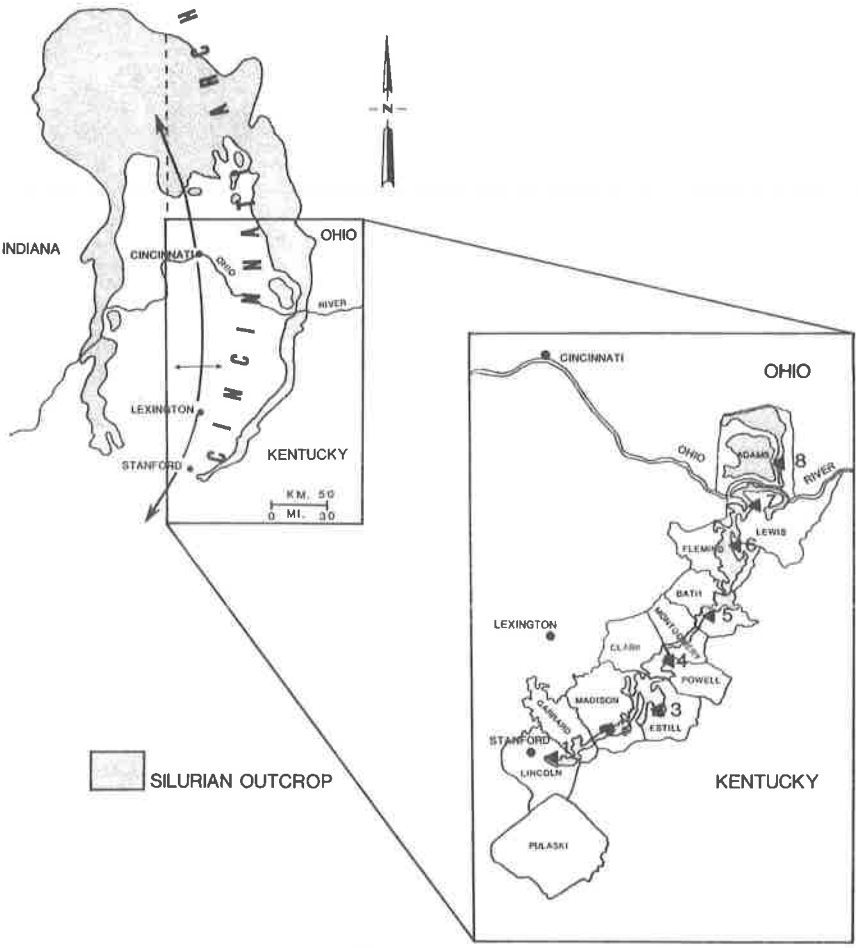

Figure 1. Location map showing Silurian outcrop pattern in Kentucky, Indiana and Ohio. Numbers refer to measured sections discussed in text.

abruptly from a dolomitic facies south of the Ohio River to a progressively more limestone-rich facies north of the river (O'Donnell, 1967; Stith and Stieglitz, 1979; Gordon, 1980).

This study began as a group project by former University of Kentucky graduate students to determine Brassfield paleoenvironments and interpret the general Early Silurian paleogeography of the eastern Cincinnati Arch region in Kentucky. Primary objectives of the project were: 1) to interpret the stratigraphy along the east-central Kentucky outcrop belt from numerous, well exposed sections, 2) collect and identify fossils from each lithofacies to determine the paleoecology of the depositional environment, and 3) examine petrographic data in order to ascertain primary depositional textures (Dunham, 1962) and the degree of alteration of primary textures as a result of dolomitization (Powers, 1962). In this report, we propose a cyclic sea-level fluctuation model for Brassfield deposition and suggest a possible mechanism for regional dolomitization.

STRATIGRAPHY

The Brassfield and younger Silurian rocks in Kentucky and neighboring states form a narrow "horseshoe shaped" outcrop belt on the flanks of the Cincinnati Arch (Figure 1). The area covered by this report stretches from the southern limit of continuous

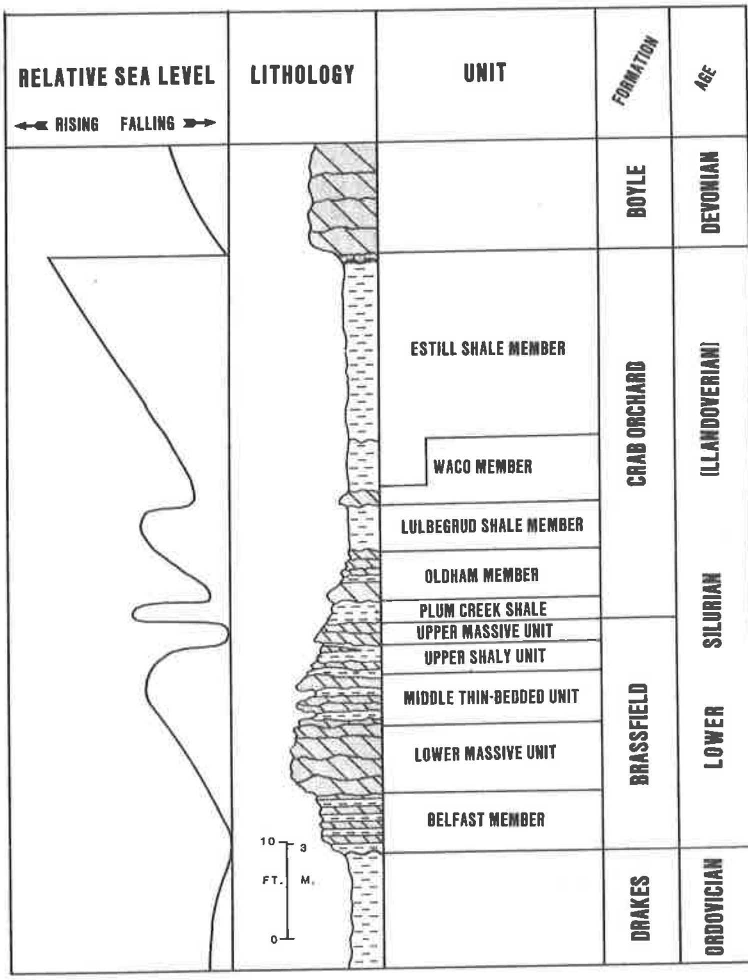

Figure 2. Composite stratigraphic column of Lower Silurian rocks in east-central Kentucky showing the ten informal and formal members recognized in this study. Fluctuating sea-levels during Early Silurian time are represented by the alternating sequence of carbonate and shale lithofacies.

Silurian exposures in east-central Kentucky (Section 1) to exposures just north of the Ohio River in Ohio (Section 8). Except for sections north of the Ohio River where the Brassfield is predominantly a limestone-rich facies, Lower Silurian strata in the study area consist predominantly of interbedded dolostones and shales. This sequence disconformably overlies the Upper Ordovician Drakes Formation (Weir and Peck, 1968) and underlies Middle to Upper Devonian dolostones (Boyle) or black shales (Ohio and Chattanooga) below a major regional unconformity. Recently completed geologic mapping in east-central Kentucky by the joint United States Geological Survey-Kentucky Geological Survey Mapping Program has shown that, except for scattered outliers, this unconformity has removed the entire Silurian section south of Lincoln County in Kentucky (McDowell, 1979). Silurian strata gradually thicken northeastward from pinchout on the unconformity to a maximum of approximately 300 ft (100m) near the Ohio River.

Lower Silurian rocks in east-central Kentucky can be subdivided into two formations and several widely correlatable members (Figure 2). The top of the upper massive lithofacies of the Brassfield was used as a reference datum in constructing a regional cross-section (Figure 3) and a schematic lithofacies diagram (Figure 4). North

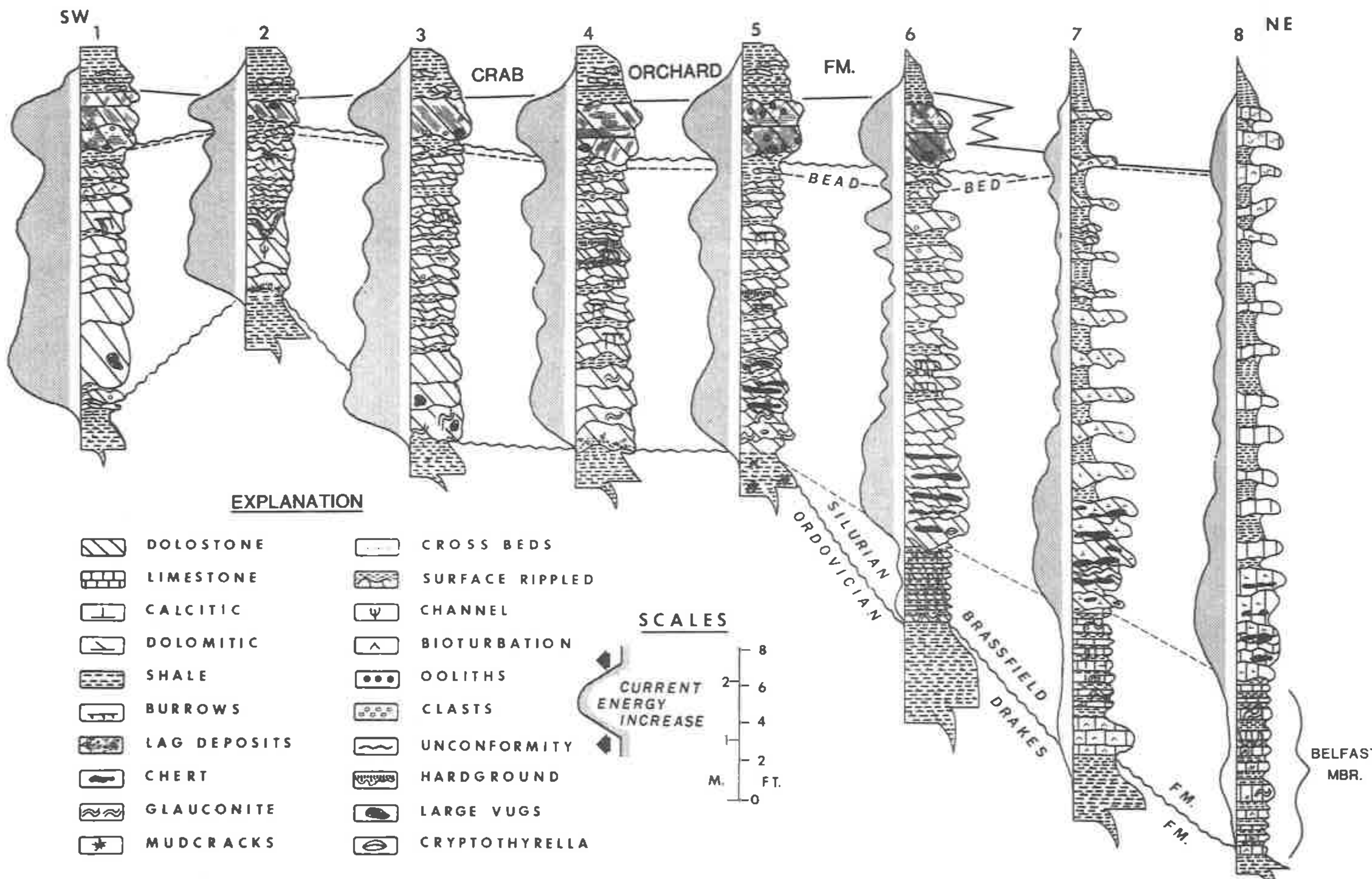

Figure 3. Brassfield stratigraphy in the study area. (See Figure 1 for locations of measured sections.) Top of the Brassfield is considered to be the top of the highest massive dolostone bed, which lies below the predominantly shale member of the Crab Orchard Formation. Horizontal not scaled.

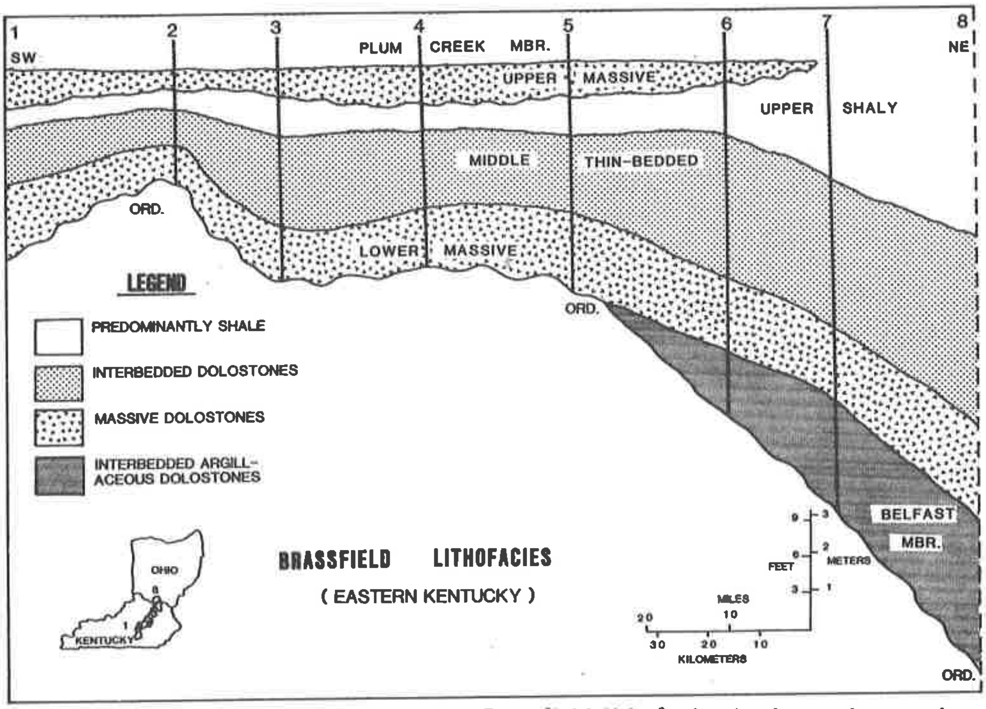

Figure 4. Schematic diagram illustrating Brassfield lithofacies in the study area based on the lithologic profiles shown in Fig. 3. (See Figure 1 for locations of measured sections.)

of Fleming County, Kentucky, where this unit is not well defined, a prominent biostratigraphic marker horizon of large crinoid columnals, known as the "Bead Bed" or "Cogwheel Bed" (Figure 3) marks the top of the Brassfield (Rexroad and others, 1965). Normally this unit occurs in shale just below the upper massive unit.

The Brassfield Formation is composed predominantly of reddish-yellow dolostones that are interbedded with greenish-gray shales. The Belfast Member, a distinctive basal unit of the Brassfield, consists of thin-bedded, argillaceous, dolomitic mudstones containing Silurian fossils (Foerste, 1896), and pinches out beneath the more typical Brassfield dolostones southwestward along the outcrop belt near Bath County, Kentucky (Figures 3 and 4). The upper part of the Brassfield can be subdivided into a lower massive lithofacies, a middle thin-bedded lithofacies, an upper shaly lithofacies and an upper massive lithofacies (Figure 2). In the study area the Brassfield attains its maximum thickness of 52 feet (16m) in Adams County, Ohio (Section 8) and gradually thins southwestward to 20 feet (6m) near Crab Orchard, Kentucky (Section 1). The overlying Crab Orchard Formation consists predominantly of greenish-gray shales but has two mappable carbonate members, the Oldham and the Waco (Foerste, 1906; Rexroad and others, 1965) (Figure 2). Only the Brassfield Formation and adjacent parts of the Crab Orchard Formation were studied in detail for this report.

GEOLOGIC SETTING

The disconformable surface upon which Lower Silurian rocks in east-central Kentucky were deposited was a slightly irregular carbonate platform that dipped gently eastward. This carbonate platform was part of a wide belt of predominantly carbonate deposition covering most of the North American mid-continent in Early Silurian time (Ham and Wilson, 1967; Johnson, 1980; Ziegler and others, 1977), and in Kentucky it represented the western flank of the subsiding Appalachian basin. The eastern flank of the basin, marginal to the uplifted Taconic Mountains and now present in West Virginia and western Virginia, experienced clastic progradation in Late Ordovician-Early

Silurian time (Dennison, 1970). Terrigenous clastics equivalent to the Silurian carbonates of east-central Kentucky are the Clinch and Tuscarora sandstones (Horvath, 1967), which are well exposed on the Pine Mountain thrust along the West Virginia-Kentucky border (Butts, 1941).

Minor relief on the platform prior to Silurian transgression is indicated by stratigraphic relationships observed at the systematic boundary with the Ordovician. The Belfast Member, the basal unit of the Brassfield, is missing southwest of Bath County, Kentucky apparently due to nondeposition or erosion near the Cincinnati Arch (Figure 4). Locally, in Montgomery and Clark counties where the Belfast is absent, a basal conglomerate consisting of thin layers of quartz sand, phosphate pebbles, glauconite and other reworked Ordovician material marks the base of the Brassfield (see Byrne, 1961). Farther southwest in Madison County, local thickness variations in the lower massive facies seem to reflect minor relief on the pre-Brassfield surface. Basal conglomerates observed in the same general area consist of reworked fragments of the underlying Drakes Formation indicating that significant erosion may have taken place. Gray and Boucot (1972) presented palynological evidence from the Ordovician-Silurian contact at the Ohio Brush Creek section (Section 8) that indicated only a moderate hiatus in that part of Ohio prior to Belfast deposition. These stratigraphic relationships seem to confirm the presence of uplifted areas on the Late Ordovician-Early Silurian platform in Kentucky. Similar areas of Early Silurian uplift on the mid-Continental platform have been reported by earlier workers in neighboring states to the west (Rexroad, 1967; Johnson, 1980).

Pre-Silurian rocks, now represented by Upper Ordovician rocks in the Cincinnati Arch region, were deposited in shallow regressive seas prior to the major westerly Silurian transgression onto the platform. Weir and Peck (1968) suggested that the Drakes Formation, which underlies the Brassfield in the study area, was deposited on extensive tidal flats. Interregional studies seem to confirm a sea-level lowstand at the close of Ordovician time (Ham and Wilson, 1967; Hatfield, 1968; Martin, 1975; Dennison, 1976; Johnson, 1980; Johnson and others, 1981), and some authors have attributed this sea-level drop to glacial activity in north Africa (see Beuf and others, 1971; Sheehan, 1975; Dennison, 1976).

LOWER-SILURIAN PALEOENVIRONMENTS

The lithology, primary sedimentary structures, and fossils generally characteristic of each Brassfield lithofacies were the primary criteria used for recognizing depositional environments (Figure 5). Inferences regarding the regional extent of depositional environments across the shallow carbonate platform were made largely by observing the vertical and lateral variations in lithologic and paleontologic character of each facies and their stratigraphic relationships with adjacent units. For example, by applying an energy-index value to each lithofacies based on lithic and faunal associations, cyclic changes in energy across the platform could be detected and correlated (Figure 3). (For a description of the energy index concept see Plumley and others, 1962; or Catalov, 1972). Each facies was also correlated with one of the five communities typically found in Llandoverian platform sequences and named after the dominant brachiopod member (Ziegler, 1965; Ziegler and others, 1968) (Figure 5). This concept of relating facies, energy and fossil communities was orginally proposed by Anderson (1971) and is used here only to establish ecologic zones and their position relative to a hypothetical shoreline. After establishing these relationships, a pattern of cyclic sea-level fluctuations in the Brassfield sequence emerged (Figure 2).

The sequence of depositional environments and communities recognized in the Brassfield and lower part of the Crab Orchard Formation is indicative of relatively minor sea-level fluctuations that are characteristic of shallow, transgressing epeiric seas. These shallow-water environments probably existed contemporaneously across the carbonate platform, migrating westward with transgression and eastward with regression. The relative onshore to offshore position of environments and communities follows the same general pattern noted in previous studies of depositional environments and communities in epeiric seas (Shaw, 1964; Irwin, 1965; Laporte, 1969; Walker and Laporte, 1970; Anderson, 1971; Johnson, 1980) (Figure 5). The vertically stacked sequence of Brassfield lithofacies in east-central Kentucky represents two minor

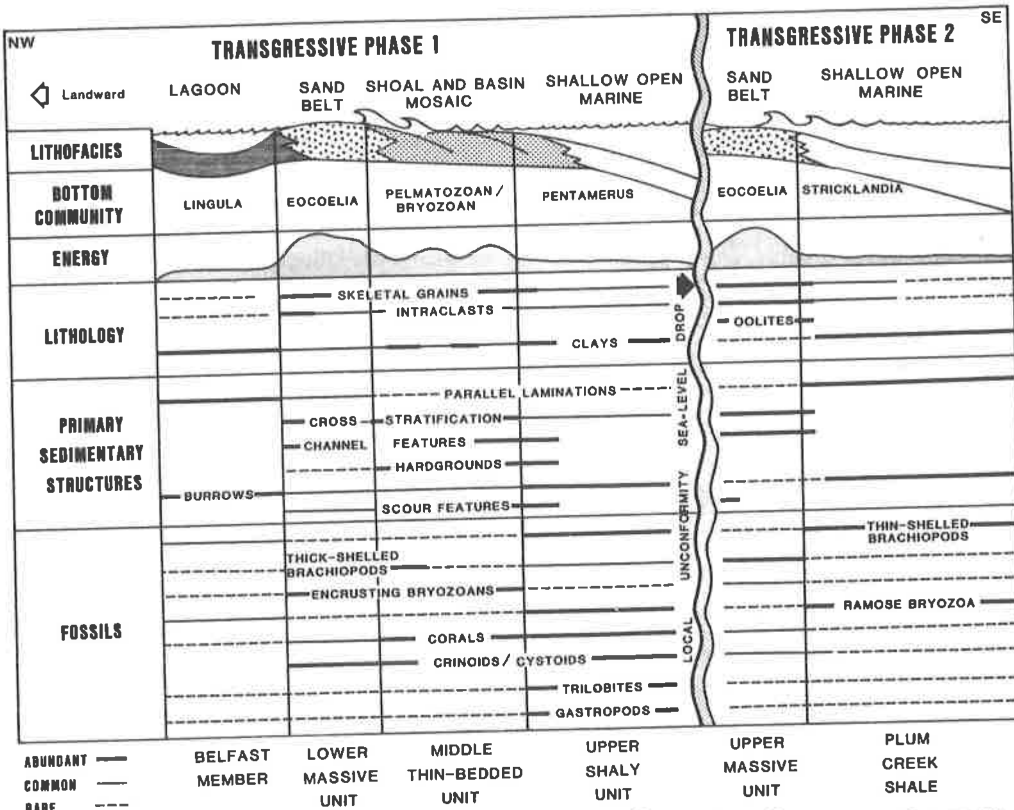

Figure 5. Hypothetical reconstruction of Brassfield depositional environments. Locally, an unconformity separates the Brassfield into two distinct transgressive cycles.

transgressive events separated by a short, local regression. These apparent sea-level fluctuations may be related to glacial events in the southern hemisphere or perhaps to local uplift of the Cincinnati Arch beginning in Early Silurian time (McDowell and others, 1980). On a much broader scale of sea-level change, the Brassfield, predominantly a carbonate lithofacies, represents near-shore environments formed at the onset of a major regional transgression. The overlying Crab Orchard Formation, predominantly a fine-grained detrital lithofacies, represents the deeper-water environments formed as Early Silurian seas approached peak transgression. The Oldham and Waco carbonate members of the Crab Orchard represent minor regressive interludes.

TRANSGRESSIVE PHASE I ENVIRONMENTS

Sediments deposited during the initial Silurian transgression of east-central Kentucky are represented in ascending order, by the Belfast Member, the lower massive facies, the middle thin-bedded facies and the upper shaly facies. Each of these facies is remarkably distinctive along the outcrop belt suggesting that they were formed in spatially uniform depositional environments on the former epicontinental platform (Figure 5). The vertical arrangement of these stacked lithofacies is thought to be the result of slow regional subsidence contemporaneous with westwardly directed transgression.

The earliest phase of Silurian sedimentation in the region is represented by the Belfast Member, which was deposited in the shallow areas on the shelf in a bay or lagoonal environment similar to that of Florida Bay (Ginsburg, 1956) or the shelf lagoon west of Andros Island in the Bahamas (Purdy, 1963). Protected from wave action by a sand-bar belt seaward, the lagoonal environment was relatively quiet, and hence ideal for the deposition of muds and poorly-sorted skeletal sands indicated by the mudstone and wakestone lithologies. The prevalence of bioturbated, gently rippled or parallel beds, indicates low-energy conditions associated with slow deposition from suspension. The high degree of infaunal activity (burrows, scolecodonts) and the occurrence of organisms best suited for life on a muddy bottom (strophomenid brachiopods) further suggest a low-energy, lagoonal environment. Gray and Boucot (1972) stated that the Belfast Member contains *Lingula* community equivalents. We suggest that lagoonal Belfast environments were probably also present southwest of their present-day preserved distribution (Figures 3 and 4), although smaller in extent. Erosion by high-energy processes on the adjacent sand-bar or shoal belt (Figure 5) near the Cincinnati Arch probably destroyed the lagoonal sediments.

Following Belfast deposition, high-energy shoal environments represented by the lower massive facies encroached upon the area. The low percentage of muddy matrix in these pelmatozoan and bryozoan grainstones is characteristic of winnowed skeletal sands like those common in Bahamian-type shoal-water environments (Imbrie and Buchanan, 1965). Rippled bedding surfaces, cross-stratification and intraformational conglomerates attest to turbulence. Channels and "bar-like" features are similar to those of the modern carbonate shoals described by Imbrie and Buchanan (1965). The low faunal diversity and ecology of the invertebrates (crinoids, thick-shelled brachiopods and gastropods) suggest a community capable of enduring high-energy conditions on a sea-floor carpeted by constantly shifting sands. The lower massive facies probably represents a chain of linear, subtidal sand bars breached by tidal channels and situated seaward of a shelf lagoon. The faunal associations and *Cryptothyrella* brachiopod community found in this lithofacies equate the sand-belt environment with the *Eocoelia* Community of the Upper Llandovery (Berry and Boucot, 1970; Boucot, 1975).

The middle thin-bedded interval contains similar features characteristic of shoaling environments. This unit does, however, contain a greater percentage of shale interbeds. The current structures and packstone/grainstone textures of the dolostone beds indicate periods of intense energy. However, the shales and intense burrowing characteristic of the unit indicate that low-energy conditions prevailed at other times, perhaps due to short-lived local transgressions. These factors indicate moderate agitation in a depositional environment, which is typical of sandy shoals and adjacent shallow basins near and occasionally above normal wave base.

The paleoecology of the middle thin-bedded unit is typical of Early Silurian "shoaly" facies. In-place hardground communities (Gordon and Ettonsohn, 1980) are common on some upper bedding-plane surfaces. This community is very diverse, being dominated by pelmatozoans, encrusting bryozoans, corals and stromotoporoids, but conspicuous in its lack of large brachiopod communities. The diversity and abundance of columnals, plates and holdfasts suggest that the echinoderm fauna was especially prolific. Cystoids, including *Brockocystis*, probably attached themselves to other epifaunal elements or to the bedrock surface by small button-shaped holdfasts. Even more common are thick, heavy, rootlike holdfasts that conformed to the irregular scoured surface of the hardground. These holdfasts probably belonged to crinoids and indicate the necessity for firm attachment in a high-energy environment. Similar hardground communities were described by Halleck (1973) from the Silurian Laurel Limestone in southern Indiana.

The pelmatozoan-bryozoan-coral communities which are common in the middle thin-bedded facies are also similar to the coral-algal communities described by Johnson (1980) from the Lower Silurian of Iowa. *Favosites, Halysites, Syringopora* and *Aulopora* are abundant in most beds although they are only rarely well preserved because of extensive dolomitization. Many flat, disc-shaped stromatolites were also recorded but are generally even more difficult to identify. Other prominent members of the communities belonging to the middle thin-bedded facies included large encrusting colonies of the coral *Heliolites* and numerous well preserved species of rugosan corals, including several interesting encrusting types. The pelmatozoan-bryozoan-coral dominated communities common in Lower Silurian platform rocks are probably equivalent to the shallow open-marine *Eocoelia* or *Pentamerous* communities (Anderson, 1971; Kaljo, 1972; Johnson, 1980; Colville and Johnson, 1982).

The upper shaly facies is very fossiliferous; corals, cystoids, crinoids, gastropods, trilobites and especially brachiopods abound. The brachiopods occur in highly bioturbated dolostone beds that originally consisted of poorly sorted, wackestones or mudstones. The abundance of terrigenous muds in this unit indicates very little winnowing, and the abundance of burrowing also reflects the low energy levels and low persistence of currents. These factors indicate intermittently agitated, quiet waters below wave base. The abundance and relatively higher faunal diversity within this facies indicates opportunistic conditions characteristic of shallow-shelf, open-marine environments (Heckel, 1972). The brachiopod associations are very similar to the *Linoporella* community described by Rubel (1970). Boucot (1975) equates the *Linoporella* Community with the *Pentamerus* Community; dalmanellids, leptaenoids and triplesiads are characteristic members.

TRANSGRESSIVE PHASE II ENVIRONMENTS

The environments and communities represented by the upper massive facies and the lower part of the Crab Orchard Formation suggest local regression followed by renewed transgressive deposition. Shallow, open-marine environments persisted north of Lewis County, Kentucky, where the upper massive beds pinch out into shales, and evidence for regression is missing in that area (Figure 3). Phosphatic, glauconitic lags and conglomerates in the lower part of the upper massive facies south of Lewis County suggest an erosive episode; however, the regressive interlude was probably rapid and the sea withdrew and returned in a relatively short period of time.

The upper massive facies was predominantly composed of pelmatozoan-bryozoan grainstones before dolomitization and is conspicuous for its lack of thick shale interbeds. This facies reflects return of high-energy, sandbelt environments following the presumed period of brief exposure. Megaripples, crossbedding and ooliths locally common in this facies support our contention that this facies represents a shoaling environment similar to that proposed for the lower massive facies. The major faunal constituents of the upper massive facies are massive encrusting bryozoans, rugose corals, thick-stemmed crinoids and communities of large *Cryptothyrella* brachiopods (see Gauri and Boucot, 1975). Hence the upper massive facies contains equivalents of the *Eocoelia* community.

The upper massive facies grades abruptly upward into the predominantly shaly rocks of the Plum Creek Shale Member of the Crab Orchard Formation. Thin beds of fossiliferous and bioturbated dolomitic mudstones and wackestones in the lower part of the member indicate the return of shallow, open-marine conditions over the entire platform. The monotonous sequence of shales in the upper part of the member reflects slightly deeper water.

The Plum Creek Member is dominated by thin-shelled strophomenid brachiopods including *Leptaena*, *Strophonella* and *Fardenia*, small orthids, dalmanellids, rhynchonellids and delicate ramose bryozoans such as *Helopora*. Walker (1975) reconstructed a community based on a similar assemblage. He suggested that the flat, high-surface area, light-weight strophomenids pioneered the quiet-water muddy bottom and later acted as a shell pavement for the attachment of ramose bryozoans, rhynchonellids and orthids. Several rock slabs collected from the lower Plum Creek suggest *in situ* preservation of a similar community. The brachiopod associations and the lithology of the Plum Creek Member suggest deeper-water communities than those in the underlying upper massive facies of the Brassfield. This environment is probably equivalent to the *Stricklandia* community of Ziegler and others, (1968).

PALEOGEOGRAPHIC RECONSTRUCTION AND DOLOMITIZATION

The fluctuating sea-level model for Brassfield-Crab Orchard deposition was derived by interpretation of a cross-section of widely correlatable, stacked lithofacies along the east-central Kentucky outcrop belt. By incorporating this two-dimensional interpretation into previously reported regional studies including isopachs, lithofacies maps and surface-to-subsurface cross-sections (O'Donnell, 1967; Horvath, 1967; Horvath and others, 1970), the Early Silurian paleogeography of the region can be approximated. The block diagram depicted in Figure 6 attempts to capture an instant in time corresponding to the close of Brassfield deposition in east-central Kentucky. The front panel of this diagram represents the sequence of stacked lithofacies as interpreted from outcrops. Deposition of the upper sand-bar belt facies is shown to be contemporaneous with deposition of the open-marine shale facies downdip along the line of the profile. The side panel is an inferred cross-section based on O'Donnell's (1967) regional lithostratigraphy and studies by Stith and Stieglitz (1979) on the extent and thickness of the Belfast Member in southwestern Ohio. It is clear from their interpretations and the distribution of Brassfield units along the outcrop belt (Figures 3 and 4), that the section depositionally thins and probable erosion in an updip direction took place along the eastern flank of the Cincinnati Arch during Early-Silurian time. The term "proto-Cincinnati Arch" is suggested because thinning updip seems to parallel the present-day structural axis of the arch. The thicker, more complete sections to the northeast near the Ohio-Kentucky border reflect subsidence or increased distance from the uplifted arch. High-energy shoals normally dominated areas to the southwest.

The fluctuating sea-level model also may be useful in explaining the regional dolomitization of the Brassfield. Petrographic examination of over 100 thin sections from the Brassfield along the outcrop belt revealed a gradual increase in the extent of dolomitization from the Ohio-Kentucky border area southwestward towards the inferred "proto-arch" (Gordon, 1980) (Figure 6). Relict textures from the dolostones and from the cherts embedded within them strongly suggest that all rocks were originally subtidally deposited biogenic limestones. We feel that the stratigraphic and petrographic relationships observed in the Brassfield suggest that some dolomitization models are inappropriate. The results of this study indicate that most Brassfield dolomites probably were formed during the mixing of carbonate waters.

The fluctuating sea-level model proposed in this study does not require long periods of severe restriction and intense evaporation during Brassfield-Crab-Orchard time to provide the supersaturated dolomitizing brines essential for the seepage-reflux (Adams and Rhodes, 1960) and capillary-transpiration mechanisms (Bush, 1974). With the exception of the Belfast Member, the presence of shallow marine environments during most of Early Silurian time and the absence of features indicating prolonged exposure, such as mudcracks, algal mats, evaporites and solution breccias preclude Brassfield dolomitization in supratidal or restricted subtidal environments. Furthermore, slow growth from diluted solutions produces relatively large, well ordered dolomite crystals. In contrast, rapid crystallization from brines produces very small (crystals less than 0.05 mm in diameter), poorly ordered, Ca-rich dolomite crystals (Dunham and Olson, 1978, 1980). In east-central Kentucky the Brassfield dolostones are coarsely crystalline (0.06 mm to 0.6 mm), and are thus more akin to those commonly associated with a mixing-zone origin.

Hanshaw and Back (1971) presented a model for dolomitization by mixing through the circulation of groundwater. Their model shows that the influx of sea water provides a continual source of magnesium for dolomitization in the zone of mixing. Studies by Plummer (1975), Badiozamani (1973), and Folk and Land (1975) have shown that mixtures of sea water and carbonate-rich groundwater could be supersaturated with respect to dolomite and undersaturated with respect to calcite, so that there is a thermodynamic potential for dolomitization of limestone by replacement. Land (1973) demonstrated that this process could have produced dolomite in Pleistocene reefs off the Jamaican coast.

Exposed areas on the Early Silurian "proto-arch" platform in Kentucky could have formed freshwater sources creating the potential for a mixing-zone type of dolomitization. The merging of sand bars in the sand-bar belt (Figure 6) due to bar migration in a high-energy zone would have created a large sheet of sand dipping very gently southeastward away from the "proto-arch". The lower massive facies (Figures 4 and 6) represents another sand sheet formed earlier in a similar migrating sand-bar belt (Figure 5). Exposure of any of these sand bodies near the "proto-arch" would have created conduits for down-dip mixing of fresh water with sea water, resulting in a brackish mixing zone near coastal areas (Figure 6). The widespread extent of the dolomitization indicates that both sand sheets probably acted as acquifers in the mixing zone, although for ease of illustration, this process is shown schematically in Figure 6 using only one bar. Movement of the mixing zone near the arch during cyclic transgression and regression could have caused successive dolomitizing events. This multiplying effect may account for the greater intensity of dolomitization observed near uplifted areas.

Pronounced zoning in many of the dolomite crystals may reflect such multiple

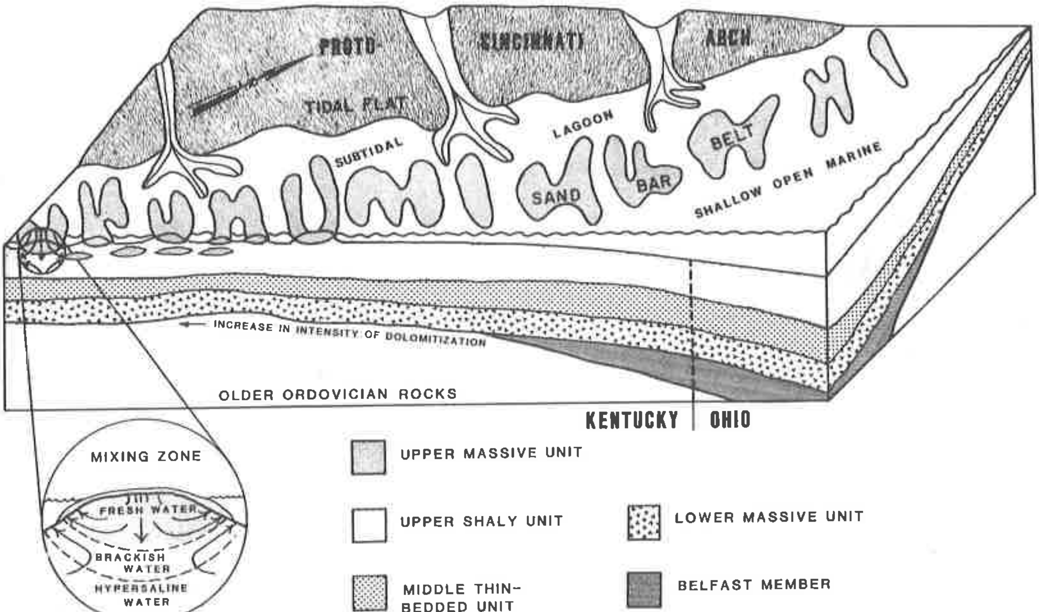

Figure 6. Paleogeographic reconstruction of east-central Kentucky during Late Brassfield time. Front panel shows sequence of Brassfield lithofacies along the eastern Kentucky outcrop belt (See Figure 4). Side panel is inferred based on earlier studies by Stith and Stieglitz (1979) and O'Donnell (1967). Mixing zone model of dolomitization is modified after Hanshaw and others, (1971).

episodes of dolomitization. Not only would this process have worked during the cyclic transgressions and regressions of the Silurian, but also during the Early and Middle Devonian when the Cincinnati Arch in Kentucky was subjected to long periods of exposure and erosion followed by periods of incomplete transgression across the arch. Middle and Upper Devonian rocks unconformably overlie the Brassfield and Crab Orchard formations along large parts of the Cincinnati Arch in Kentucky, and the erosional gaps increase greatly to the southwest (McFarlan, 1943). We suggest that the Early and Middle Devonian may have been an even more opportune time for mixing-zone dolomitization of the Brassfield. Uplift on the arch was apparently greater in intensity and duration at this time.

The lack of pervasive dolomitization in the Brassfield of Ohio may indicate that the brackish mixing zone never extended that far away from the arch in the subsurface. The increased distance of the Brassfield outcrop belt in Ohio from the arch is apparent on maps (Figure 1), but is also indicated by the thicker, more complete sections to the northeast and the dominance of shale indicating deeper-water environments in that area (Figure 3). Decrease in the size, number and proximity of carbonate sand layers (Figure 3) useful as groundwater conduits may have been equally important in the decreased dolomitization to the northeast of the arch. Finally, distance from the arch would have reduced the effects of cyclic sea-level fluctuations and largely precluded the cumulative dolomitization effects suggested by the intensely altered textures observed southwestward toward the arch.

SUMMARY AND CONCLUSIONS

Stratigraphic, sedimentologic and paleoecologic evidence indicates that the Lower Silurian Brassfield Formation exposed on the eastern flank of the Cincinnati Arch in east-central Kentucky was deposited in shallow, carbonate platform environments during a regional westerly transgression of seas over an uplifted "proto- Cincinnati Arch". Relatively minor fluctuations in sea-level during Brassfield and Early Crab Orchard time resulted in lateral shifts of depositional environments and faunal communities observed in the vertical sequence of lithofacies and biofacies. These sea-level changes may be related to glacial activity in the southern hemisphere or to local uplift along the "proto- Cincinnati Arch". Shoaling environments dominated high areas near the "proto-arch" south of Bath County, Kentucky, during Brassfield deposition, and minor unconformities in the same area indicate occasional exposure. The Early Silurian sea approached maximum extent in east-central Kentucky during deposition of the upper Crab Orchard sediments, when the relatively thick Estill and Lulbegrud shales were deposited.

Sea-level fluctuations may also explain the dolomitization of primary biogenic limestones on uplifted areas during the Silurian and perhaps more importantly during the Early and Middle Devonian. Areas exposed during sea-level lowstands could have provided freshwater aquifers that mixed downdip with sea-water along coastal areas so as to form a brackish mixing zone. Migration of the mixing zone with transgression and regression could account for the pervasive dolomitization of the Brassfield near uplifted areas.

ACKNOWLEDGMENTS

The writers gratefully acknowledge the assistance that was furnished by many persons during the course of this project. Former University of Kentucky graduate students, Mike Miller, Dennis Swager and Donald Chestnut, contributed ideas that significantly influenced the nature and outcome of the investigation. Dr. Lois J. Campbell and Dr. William C. MacQuown contributed ideas which significantly improved the presentation. Dr. John Thrailkill brought to our attention ideas concerning mixing-zone dolomitization and critically reviewed portions of the manuscript dealing with this subject. D. E. Hattin, and an anonymous reviewer provided many comments that significantly improved the paper. Tom Dugan, a consulting geologist in Lexington, Kentucky, assisted the senior author in the field.

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FLUVIAL TERRACES AND LATE PLEISTOCENE TECTONISM IN GEORGIA

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ABSTRACT

There are six identifiable fluvial terraces along the major rivers of Georgia. Remnants of the terraces lie at the following heights, in feet, above present river levels: 10-20, 30-50, 60-80, 110-130, 140-160, and 170-190. The terraces appears to be correlative to coastal terraces in Georgia as indicated in Table 2.

Gradients of Coastal Plain sections of the terraces are consistent with modern river gradients, confirming that the terraces have been correctly identified. Terraces extending along the Chattahoochee River from 30 to 70 valley miles below Columbus, Georgia are elevated 50 to 100 feet above average gradient. The area appears to have been uplifted in post-Sangamon time, possibly by a fault near Ft. Gaines, Georgia. Abnormally high terrace gradients on the Flint River, where it crosses Pine Mountain, suggest southward tilting of the area in post-Sangamon time.

INTRODUCTION

Although the existence of fluvial terraces in the Coastal Plain of Georgia has been recognized since the classic work of Veatch and Stephenson (1911), little attention has been given to the subject in recent years, probably because large scale topographic maps of much of the area were available only after the mid to late 1970s.

In 1978 the senior author conducted a seminar on fluvial sedimentation and geomorphology for a group of archaeologists. Subsequent discussions and field trips established that fluvial terrace remnants, presumably of Pleistocene age, extend well into the Piedmont province, specifically in the Oconee River valley at Wallace Reservoir. We were unaware of any report of river terraces in the Piedmont province and the fact of their existence was so surprising, and interesting, that Susan A. Waters undertook a study of fluviatile terraces in Georgia as her honors thesis project in 1978 (see Waters, 1979). Subsequently, Brook (1981) described two terraces on the Oconee River in the vicinity of the Wallace Reservoir.

PREVIOUS WORK

Veatch and Stephenson (1911; Stephenson and Veatch, 1915) first reported fluvial terraces in Georgia. They recognized two coastal terraces, physiographically the Okeefenokee Plain and the Satilla Coastal Lowland, each with fluvial terrace equivalents extending up the local rivers to the Fall Line. Stephenson and Veatch (1915) defined the fluvial correlatives of their Okeefenokee Plain as lying 50 to 100 feet above present river level, with only remnants surviving erosion "...that formed the lower terrace." The fluvial correlatives of the lower terrace, or Satilla Coastal Lowlands, were said to lie 15 to 50 feet above present river levels and to be conspicuous at Columbus, Macon, and Augusta. They also refer to the lower terrace as "second bottoms" (quotes in original), a term that is still used by Coastal Plain residents.

Teas (1921), following Veatch and Stephenson's (1911) terminology, described the physical properties of the "Okeefenokee formation" and "Satilla formation" at numerous localities in the Coastal Plain in connection with his report on sand and gravel resources.

Woodruff and Parizek (1956) reported two erosional terraces in valleys of the Georgia Piedmont, and Dennis noted that terraces in the Piedmont occur at 9-12 m (30-40 feet), 18-24 m (60-80 feet), and 30-46 m (100-150 feet) above the present floodplain. These workers did not relate the Piedmont terraces to the Coastal Plain terraces described by Veatch and Stephenson.

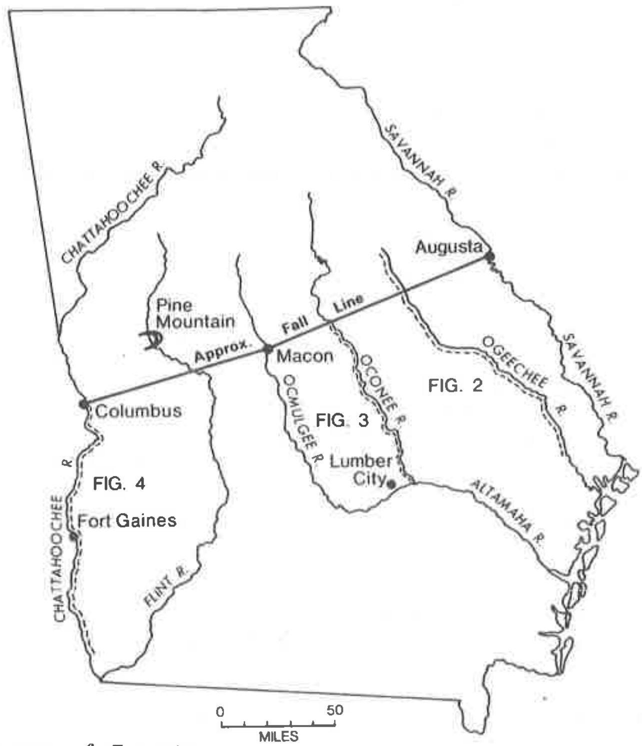

Figure 1. Index map of Georgia.

Roberts (1958) defined four terraces on the Chattahoochee River between Ft. Gaines, Georgia and Chattahoochee, Florida, employing Paulin altimeter surveys and Corps of Engineers preliminary "blue line" topographic maps in Georgia, and 1:24,000 U.S. Geological Survey topographic maps in Florida. Roberts defined the terraces on the basis of their height above the surface flooded in 1929, the modern floodplain. According to Roberts the terraces are 10 to 20, 30 to 50, 70 to 110, and 135 to 160 feet above the modern floodplain.

Roberts (1958, p. 5) noted that the Chattahoochee River is entrenched to a depth of 40 to 50 feet in the northern part of his study area, and that tributaries are deeply entrenched in the modern floodplain (p. 18), but found a consistent down-valley terrace and floodplain slope of 1.3 to 1.4 feet per mile and concluded that there has been no "regional tilting" since formation of the highest and oldest terrace (p. 44). Veatch and Stephenson (1911, p. 41 and 63-64), however, cite the entrenchment of the Chattahoochee and its tributaries, the near absence of floodplain, and the narrowness of the drainage basin, compared to that of the adjacent Flint River, as evidence for uplift along the Chattahoochee anticline, a southward pitching anticline with its axis coincident with the Chattahoochee River. The resolution of this conflict in interpretation lies, apparently, in the fact that the northern limit of Robert's study areas lies very near the southern limit of recent uplift reported in this paper.

RESEARCH METHODS

The Piedmont and Coastal Plain sections of the Savannah, Ogeechee, Oconee, Ocmulgee, and Flint Rivers, and the Coastal Plain section of the Chattahoochee River drainages (Figure 1) were studied. Terraces were identified by topographic expression on U.S. Geological Survey 1:24,000 (7 1/2 minute) topographic maps, aerial photographs, and Landsat photographs. Terraces were defined on the basis of their altitude above present river level (bank elevation) and plotted on base maps prepared from 1:250,000 U.S. Army Topographic Map Command maps.

Terrace gradients are based on actual terrace elevations and down valley

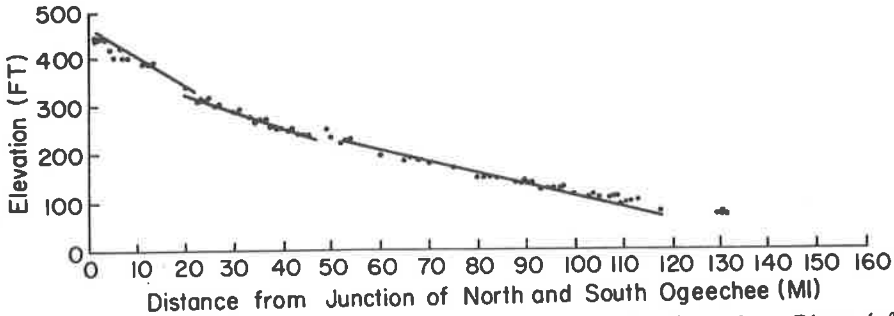

Figure 2. Terrace profile for the 10-20 foot terrace on the Ogeechee River (after Waters, 1979).

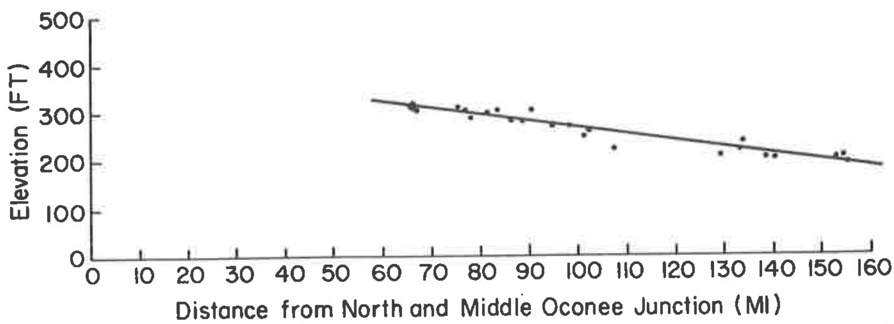

Figure 3. Terrace profile for the 60-80 foot terrace on the Oconee River (after, Waters, 1979).

distances, while river gradients are based on river elevations, as indicated by contours crossing the river, and down-valley distances. There is some scatter of data in the river and terrace profiles; due to actual differences in terrace elevation, small errors inherent in the topographic maps, the subjective nature of down-valley distance, and the concavity of strain profiles. There therefore is some subjectivity in the choice of slope line, but, based on replication, the gradients reported here are thought to be precise within 0.2 feet per mile. Waters (1979) was able to confidently determine terrace gradients from 28 of the 26 terrace profiles she prepared. Two profiles, typical of the 28, are illustrated in Figures 2 and 3.

All of the terraces can be traced into the Piedmont, but they are not areally extensive. Near the Coast the higher terraces are absent and the lower terraces merge with the coastal marine terraces.

GRADIENTS

Gradients of the terraces are higher in the Piedmont than in the Coastal Plain and are considerably more variable. Gradients for Piedmont terraces, where sufficient terrace remnants to determine gradients occur, range from 2.2 feet per mile (Oconee River, 10-20 foot terrace) to 5.5 feet per mile (Ogeechee River, 10-20 foot terrace), averaging about 4 feet per mile. Variation in Piedmont gradients are assumed to be a function of the erodability of the underlying rock.

There is one interesting exception to the above generalization. Where the Flint River cuts through Pine Mountain, about 40 valley miles south of Jonesboro, the gradient of the 10-20 and 60-80 foot terraces appears to be about 12 feet per mile. There are too few remnants of the other terraces to determine gradient. This is an interesting fact that is discussed in a subsequent section of this paper.

Coastal Plain terrace gradients and modern rivers gradients for the upper to middle Coastal Plain are presented in Table 1. The consistency of gradients, considering that the precision is probably about 0.2 feet per mile, gives confidence in our identification of six terraces.

Table 1. Gradients of modern rivers and Pleistocene fluvial terraces in feet per mile, upper and middle coastal plain area.

	River					
	SAV	OGE	OCO	OCM	FLT	CHA
Modern Gradient	1.2	2.1	1.4	1.4	1.8	1.3
Terrace Gradient						
10-20	1.4	1.9	1.3	1.4	1.8	1.3
30-50	1.3	1.7	1.3	1.4	1.7	1.3
60-80	1.3	1.9	1.4	1.3	1.8	1.3
110-130	1.3	1.7	1.3	1.3	1.8	1.3
140-160	ND	ND	1.3	1.5	ND	ND
170-190	1.2	ND	1.5	ND	ND	ND

ND = Not determined, insufficient data
SAV = Savannah River
OGE = Ogeechee River
OCO = Oconee River
OCM = Ocmulgee River
FLT = Flint River
CHA = Chattahoochee River

Table 2. Correlation of fluvial and coastal marine terraces. Elevations of the upper three terraces are from Colquhoun (1974), the remainder are from Hoyt and Hails (1974). Because of the concavity of river profiles, fluvial terraces may be somewhat higher above present river levels than coastal terraces are above sea level.

Coastal Terrace	Elevation	River Terrace (feet above present river)
Coharie	245 ft.	——
Sunderland	170	170-190
Okeefenokee	135	140-160
Wicomico	95-100	110-130
Penholoway	70-75	60-80
Talbot	40-45	30-50
Pamlico	24	10-20
Princess Anne	13	
Silver Bluff	4.5	*

* Correlatives lie within modern floodplain

CORRELATION WITH COASTAL TERRACES

It is assumed that the river terraces described here are ultimately correlative with the coastal marine terraces that have received so much attention since their recognition by Lyell (1845). Each of the terraces should represent the state of a given river at the time of formation of one of the major coastal marine terraces, but correlation is difficult because differences in elevation between modern river level and the terraces decreases downstream, due to concavity of the valley profiles.

Hoyt and Hails (1974) and Winker and Howard (1977) have shown that areas of the Atlantic Coastal Plain north and south of Georgia have been uplifted since formation of the Pamlico Terrace (Chatham sequence of Winker and Howard); present elevations of older terraces are therefore not useful in correlation over distances greater than about 300 miles. Opdyke and others (1984) report that Pleistocene-Holocene uplift in northern Florida and southern Georgia is due to isostatic rebound associated with limestone solution. They calculate that at least 36 m (118 feet) of uplift has occurred in north-central Florida. The marine terraces of Georgia, except for those in extreme south Georgia, appear to have been unaffected, or only slightly affected, and correlation of the marine and fluvial terraces probably can be most reliably determined in Georgia.

A correlation of marine and fluvial terraces is presented in Table 2. No equivalent of the Coharie Terrace was observed. If a fluvial correlative exists, it may be too eroded to be recognizable. The 10-20 foot fluvial terrace probably is correlative

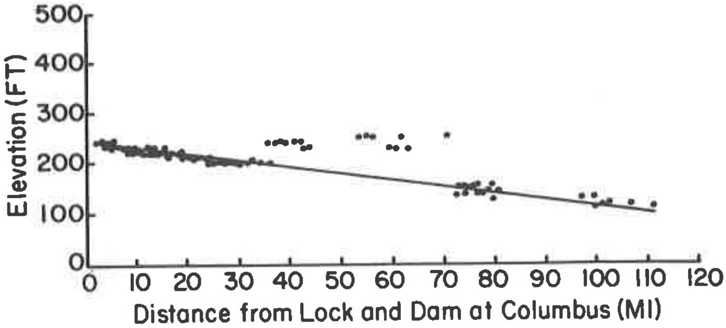

Figure 4. Profile of the 40-50 foot terrace on the Chattahoochee River. The distribution of elevations suggests 50 to 100 feet of uplift, ending abruptly near Ft. Gaines, Georgia (after Waters, 1979).

with both the Pamlico and Princess Anne marine terraces, and the correlatives of the Silver Bluff Terrace probably are indistinguishable from the modern floodplain.

Voorhies (1974) found Pliocene horse teeth at a height of approximately 120 feet above the Flint River at Reynolds, Georgia (about 10 miles south of the Fall Line), but the locality is not on a feature that we identify as a terrace. Further, the terraces may be in part erosional. Fossils in the material of erosional terraces date the material, but not the terrace.

LATE PLEISTOCENE TO HOLOCENE UPLIFT

The central and upper Coastal Plain sections of nearly all terrace profiles for the rivers discussed here are essentially straight lines but slightly concave upward (Figures 2 and 3). Slopes are consistent with modern river gradients (Table 1), suggesting that there has been no tectonic activity since formation of the oldest recognizable terrace.

Exceptions occur on the Chattahoochee River from 30 to 70 valley miles below the lock and dam at Columbus. Here the terraces are elevated 50 to 100 feet above the average and expected profiles. This is recognizable in profiles for the lower four terraces, there being too few data for the higher terraces, but is best illustrated by the profile of the 30-50 foot terrace (Figure 4). Because the few remnants of the 10-20 foot terrace in the upper Coastal Plain section appear to be affected, the uplift must have postdated the formation of the 10-20 foot terrace. Colquhoun (1974) estimates that the Princess Anne marine terrace (termed cyclic unit by Colquhoun, but formation by Hoyt and Hails, 1974) is of Sangamon age and, if the correlation of Table 2 is correct, the uplift must be post Sangamon.

The change in terrace elevation near valley mile 70 of Figure 4 appears abrupt and suggests a fault, down to the south, at that location. Valley mile 70 is near Fort Gaines and Walter F. George Lock and Dam. The Chattahoochee River is entrenched above and at least 18 miles below this point, and we have found no geomorphologic or geologic evidence for a fault. Nonetheless, Patterson and Herrick (1971) find little evidence for the Chattahoochee Anticline, of Veatch and Stephenson (1911) and a fault may therefore best explain the entrenchment of the Chattahoochee River and uplift of the fluvial terraces.

A second unusual feature of the terrace profiles, previously mentioned, is the steep gradient of the 10-to-20 and 60-to-80 foot terraces where the Flint River crosses Pine Mountain. Pine Mountain is arcuate, concave to the west, and the Flint River has cut 500 foot gorges through the northern and southern limbs of the arc (Figure 1). Terrace gradients that can be determined in this area are about 12 feet per mile, while the present river gradient is between 6.5 and 9 feet per mile. This suggests some southward tilt, or uplift of the northern part of the area, in post-Sangamon time. Interestingly, an earthquake that occurred on October 30, 1982, was felt in Harris and Muscogee Counties. The western end of the Pine Mountain thrust sheet is in Harris County; Columbus, Georgia is in the next county south, Muscogee County. There apparently is a currently active fault in the vicinity of the uplifts indicated by the terrace data.

ACKNOWLEDGEMENTS

I wish to thank Hatten Howard, Joann Stewart, Charles Winker, and Fredric Pirkle for their thoughtful and helpful reviews of the original typescript.

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