SOUTHEASTERN GEOLOGY

PUBLISHED AT DUKE UNIVERSITY DURHAM, NORTH CAROLINA

VOL. 13 NO. 3 OCTOBER, 1971

SOUTHEASTERN GEOLOGY

PUBLISHED QUARTERLY

AT

DUKE UNIVERSITY

This journal welcomes original papers on all phases of geology, geophysics, and geochemistry as related to the Southeast. Transmit manuscripts to S. DUNCAN HERON, JR., BOX 6665, COLLEGE STATION, DURHAM, NORTH CAROLINA. Please observe the following:

(1) Type the manuscript with double space lines and submit in duplicate.
(2) Cite references and prepare bibliographic lists in accordance with the method found within the pages of this journal.
(3) Submit line drawings and complex tables as finished copy.
(4) Make certain that all photographs are sharp, clear, and of good contrast.
(5) Stratigraphic terminology should abide by the Code of Stratigraphic Nomenclature (AAPG, v. 45, 1961).

Proofs will not be sent authors unless a request to this effect accompanies the manuscript.

Reprints must be ordered prior to publication. Prices are available upon request.

* * * * * *

Subscriptions to Southeastern Geology are $5.00 per volume. Inquiries should be addressed to WM. J. FURBISH, BUSINESS AND CIRCULATION MANAGER, BOX 6665, COLLEGE STATION, DURHAM NORTH CAROLINA. Make check payable to Southeastern Geology.

SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 13, No. 3

1971

TRANSMISSIVITY TRACTS IN THE COASTAL PLAIN AQUIFERS OF MARYLAND

By

Harry J. Hansen
Maryland Geological Survey
Baltimore, Maryland 21218

ABSTRACT

The Coastal Plain sediments of Maryland exhibit three major transgressive (→)-regressive (←) sedimentary cycles. In these cycles three major lithotopes are repeated. The first cycle consists of the Lower Cretaceous Potomac Group (fluvial) → Upper Cretaceous Magothy Formation (fluvio-marine) → Upper Cretaceous Matawan and Monmouth Formations (marine) ← Paleocene Brightseat Formation (marine) ← and Paleo-Eocene Aquia (marine) Formation. The second cycle consists of the Eocene Marlboro Clay (fluvio-marine) → Eocene Nanjemoy Formation (marine) ← Eocene Piney Point Formation (marine) ← and Oligo-Miocene Lacuna (unconformity); the third cycle is defined by the Oligo-Miocene Lacuna → Miocene Chesapeake Group (marine) ← Miocene Yorktown Formation (fluvio-marine) ← and Columbia Group (nonmarine).

In Maryland regional transmissivity tracts (areas of high transmissivity) characteristic of the nonmarine lithotope (e.g. Potomac Group, Columbia Group) are generally subparallel to the regional dip. Formations of the marine lithotope, such as the Aquia and Piney Point, exhibit tracts subparallel to the regional strike. The former suggests a predominance of fluvial input during deposition, the latter of marine redistribution.

Transmissivity tracts in Coastal Plain Maryland generally occupy the more axial portions of a major landward extension of the Baltimore Canyon Trough (Maher, 1967), suggesting coarse detrital input from a long existing proto-Susquehanna drainage system. The Baltimore Canyon Trough was an active depocenter during much of Cretaceous and Tertiary time.

INTRODUCTION

The Coastal Plain formations of Maryland crop out in a series of concentric bands subparallel to the Fall Zone (Cleaves and others, 1968). They dip southeasterly, generally less than 1°, and thicken seaward to form a sedimentary wedge that exceeds 10, 000 feet beneath the continental shelf (Maher, 1967). The formations range in age from Lower Cretaceous to Pleistocene and consist of unconsolidated beds of clay, silt, sand and gravel; ferrugenous or calcareous cementation is stratigraphically restricted. In the Coastal Plain two subsurface systems control the occurrences of ground water. The geologic system is static and functions as a framework whereas the interstitial fluid system is dynamic and responds rapidly to stresses resulting from groundwater pumpage.

This paper is concerned with the geologic framework or static portion of the subsurface. Insofar as it affects hydraulic gradients within the fluid system, one aspect of the geologic framework is particularly important to the ground-water hydrologist. This is transmissivity, an hydrologic parameter that is equal to the product of field permeability and thickness; it is essential for predicting the time-yield-drawdown relationships associated with ground-water withdrawals. In the Maryland Coastal Plain high transmissivity values are associated with thick, sand and gravel facies. Therefore, knowledge of the depositional processes responsible for the distribution of these facies aids in the exploration for productive ground-water sources.

GEOLOGIC FRAMEWORK

Previous Investigations

The first hydrogeologic studies of the Maryland Coastal Plain were those initiated by Darton (1896, 1902) and Clark and others (1918). Since 1949 a series of county reports has been published by the Maryland Geological Survey (1970). The earlier investigations erected a hydrogeologic framework for the Coastal Plain. The later reports stressed quantitative aspects of the ground-water system.

Regional stratigraphic studies of the Atlantic Coastal Plain include those by Richards (1948), LeGrand (1961), Maher (1965, 1967), Emery (1967) and U. S. Geological Survey (1967); extensive bibliographies are contained in Murray (1961) and Glaser (1968). Studies by Back (1966) and Upson (1966) have investigated the regional ground water flow patterns occurring in the North Atlantic Coastal Plain; these studies discuss the hydrodynamic factors controlling the distribution of hydrochemical facies, including salt-fresh water boundaries.

The Major Sedimentary Cycles of the Maryland Coastal Plain

The Coastal Plain sedimentary of Maryland were deposited in a wide range of sedimentary environments that may be conveniently grouped into three major classes or lithotopes (as defined by Dunbar and Rogers, 1957, p. 137): Nonmarine (or fluvio-deltaic), Fluvio-marine (or strand zone), and Marine (or shelf).

As shown by their depositional products, the stratigraphic succession of environments may occur transgressively or regressively (Curray, 1964). The term transgression connotes a temporal migration of the strand zone in a landward direction and results in the superposition of nonmarine, fluvio-marine, and lastly, marine sediments. Conversely, the term regression connotes migration in a seaward direction and results in the overlapping of marine deposits by fluvio-marine and nonmarine deposits.

In the Maryland Coastal Plain three major sedimentary cycles define pre-Pleistocene stratigraphy (Glaser, 1968). Figure 1 is a schematic diagram showing the tripartite cyclicity exhibited by the sediments; the relationship between depositional environment and upper range in transmissivity is shown. Figure 1 is meant to provide an analytical framework and is not a complete empirical generalization.

The transgressive phase of the first cycle begins at the base of the Coastal Plain and is represented by the fluvial Potomac Group of Cretaceous age (Clark, et al, 1911; Glaser, 1967). In Maryland the Potomac Group includes, from oldest to youngest, the Patuxent, Arundel and Patapsco Formations. Overlying the Potomac Group is the Upper Cretaceous Magothy Formation, a strand-zone deposit of fluvio-marine origin (Clark, 1916, p. 87 ; Overbeck and Slaughter, 1958, p. 55). Marine deposits are represented by the Upper Cretaceous Matawan and Monmouth Formations (Goldman, 1916, p. 173; Overbeck and Slaughter, 1958, p. 59) which overlap the Magothy and mark the end of the transgressive phase of the first cycle.

The regressive phase begins with the Brightseat Formation, a marine unit of Paleocene age (Bennett and Collins, 1952; Nogan, 1964). The Overlying Aquia Formation of Paleocene-Eocene age (Hazel, 1968) reflects regressive sedimentation (Glaser, 1968, p. 17) with the upper part of it deposited in very shallow water, perhaps littoral (Drobnyk, 1965). Overlying the Aquia Formation in updip areas is the commonly lignitic, pink to gray Marlboro Clay (Darton, 1948), a fluvio-marine unit (Nogan, 1964).

The Nanjemoy Formation of Eocene age (Shifflet, 1948) overlies the Marlboro Clay. This Formation was deposited during a rapid marine transgression. Both the mineralogy (glauconite) and fossil content suggest neritic deposition (Cooke, 1952). The Piney Point Formation of upper Eocene age overlies the Nanjemoy Formation (Otton, 1955). This coarsely arenaceous unit occupies the same sedimentological niche as the older Aquia Formation and represents shallow neritic deposition during the concluding (regressive) phase of the second sedimentary cycle.

To date sediments of Oligocene age have not been identified in Maryland (Cleaves, et al, 1968). In the absence of contrary evidence it is assumed that the Oligocene record occurs in Maryland as a missing stratigraphic interval, or lacuna as defined by Wheeler (1964,

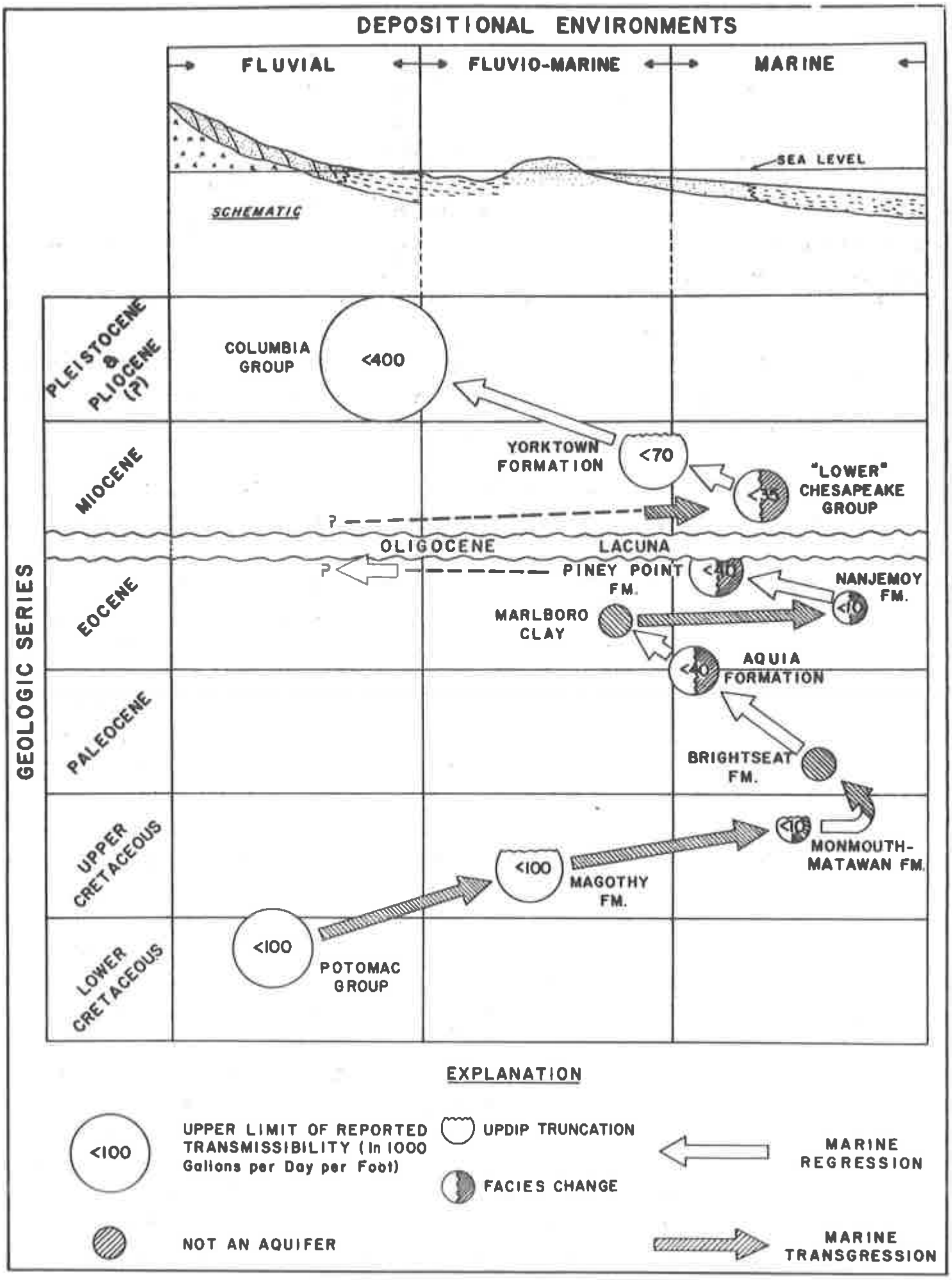


Figure 1. Diagram showing tripartite cyclicity exhibited by three Coastal Plain lithotopes in Maryland.

p. 602). The transgressive phase of the third sedimentary cycle begins with the Middle Miocene Calvert Formation. The basal bed of this

Formation is an arenaceous unit containing small quartz pebbles, suggesting to Dryden and Overbeck (1948, p. 53) a very near shore origin. Based on lithologic and faunal evidence, beds of the overlying Calvert, Choptank, and St. Marys Formations were deposited in a shallow marine setting (Cooke, 1952, p. 34; Gibson, 1962, p. 60-66); Gibson (1962) has noted that the Miocene Formations exhibit transgressive-regressive cyclicity, although of a smaller magnitude than the three major cycles shown in Figure 1.

The Upper Miocene Yorktown Formation, which contains coarse sandy facies, was deposited in a shallow marine to perhaps deltaic or estuarine environment, (Rasmussen and Slaughter, 1955, p. 43). It represents a major regressive phase of the third sedimentary cycle. In Maryland regressive sedimentation continued into the Pliocene which is represented chiefly by thin nonmarine (fluvial) deposits (Schlee, 1957) or by a lacuna (missing interval).

The Pleistocene epoch was characterized by geologically rapid shifts in sea level which resulted in several transgressive-regressive oscillations, although not of the same temporal scale as the tripartite cyclicity described above. Except for eastern portions of the Delmarva Peninsula, most Pleistocene sediments preserved in the stratigraphic record are fluvial (Jordan, 1964; Hansen, 1966).

Tectonic Setting

The Chesapeake-Delaware Embayment is the major structural feature of the northern Atlantic Coastal Plain (Murray, 1961, p. 92). It extends from Cape Fear, North Carolina to about Long Island, New York as a broad downwarping of the basement and overlying sediments. The Baltimore Canyon Trough is a major offshore element of the Chesapeake-Delaware Embayment (Maher, 1967, plate 4). Contours drawn on top of the basement rocks show that a landward extension of this feature trends beneath the northern portion of the Delmarva Peninsula (Figure 2).

The transmissivity tracts discussed below are generally associated with the axial portions of this depocenter. By determining where in the embayment major fluvial inputs would occur, one can infer that basinal subsidence has controlled the sedimentary patterns developed during Mesozoic and Tertiary deposition. It is interesting to note that the modern Susquehanna River, which has nearly twice the discharge of any river draining the East Coast of the United States (Wilson, 1967), occupies a near axial position in the embayment (see Maher, plate 4). Apparently the Susquehanna River has had antecedents of similar magnitude.

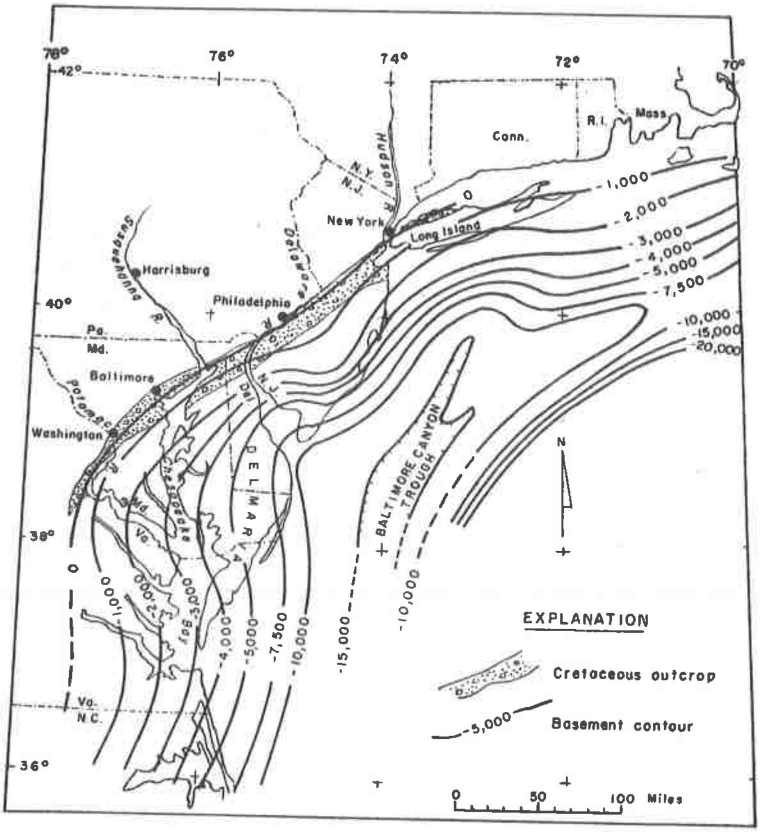


Figure 2. Regional structure contour map, showing occurrence Baltimore Canyon Trough (adapted from Maher 1967, pl. 4).

RELATIONSHIP BETWEEN DEPOSITIONAL ENVIRONMENT AND TRANSMISSIVITY

The Coastal Plain aquifers of Maryland can be categorized into three environmentally determined lithotopes that are stratigraphically repeated to form three major sedimentary cycles of the transgressive-regressive type (Figure 1). Of the three lithotopes, sandy facies associated with nonmarine (fluvial) sediments have the highest transmissivity values.

Two stratigraphic units, the Potomac Group and the Plio-Pleistocene Series, exemplify the nonmarine (fluvial) lithotope. The Potomac Group is a multi-aquifer unit used chiefly in the inner Coastal Plain

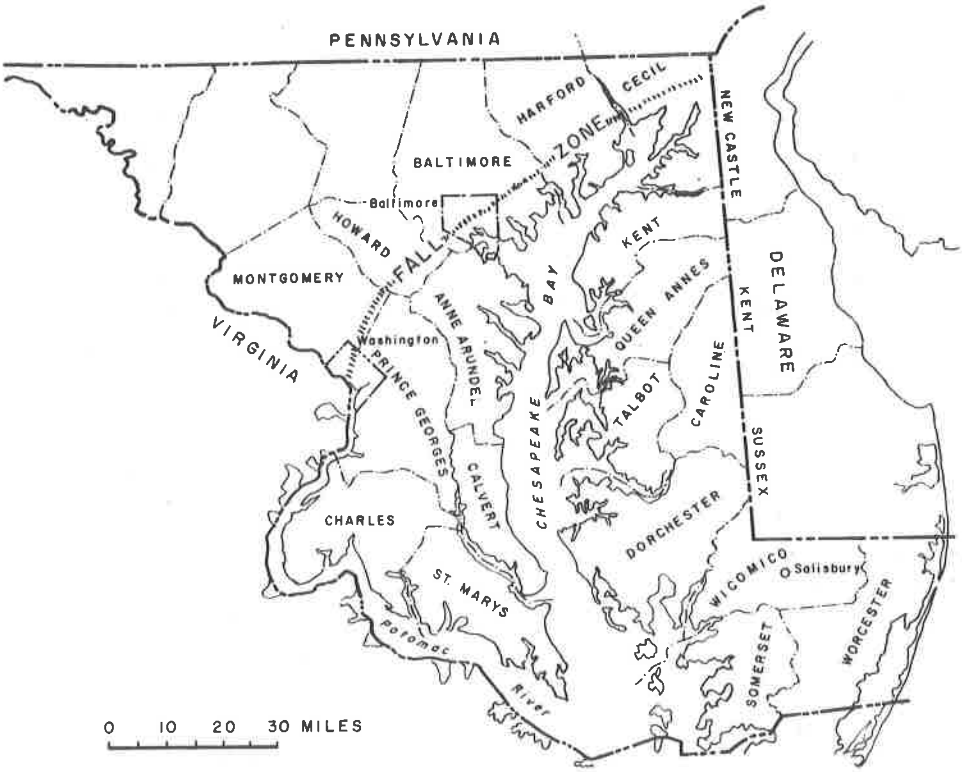


Figure 3. Map of Maryland showing Coastal Plain Counties east of Fall Zone.

counties of Cecil, Harford, Baltimore, Anne Arundel, Prince Georges and Charles (Figure 3). Maximum transmissivity values are reported in southern Baltimore County and northern Anne Arundel County; the values range up to 80,000 gpd/ft. and may, perhaps, reach 100,000 gpd/ft. (Mack, 1962; Bennett and Meyer, 1952; Otton, et al, 1964); both north and south of this area, transmissivities decrease significantly, rarely exceeding 20,000 gpd/ft. (Otton, 1955; Overbeck and Slaughter, 1958; Mack, 1966).

Sediments of the Pleistocene Series represent an important source of ground water in the Eastern Shore Counties of Caroline, Dorchester, Wicomico, and Worcester. Like the Potomac Group, Pleistocene sediments are chiefly fluvial and have transmissivities that exceed 100,000 gpd/ft. (Rasmussen and Slaughter, 1955; 1957) and in some cases 400,000 gpd/ft. (Mack and Thomas, 1968, p. 53).

In Maryland the Upper Cretaceous Magothy Formation and the Miocene Yorktown Formation are fluvio-marine lithotopes, In axial areas of the embayment, where coarse detritus from a proto-Susquehanna source was deposited, both the Magothy (Anne Arundel County) and Yorktown (Worcester County) Formations exhibit transmissivities exceeding 50, 000 gpd/ft. (Mack, 1962). However, in more distal areas values are usually less, sometimes ranging below 10, 000 gpd/ft.

The third major lithotope consists of marine sediments. Aquifers grouped into this category include the Matawan, Aquia, Nanjemoy, and Piney Point. These may be subdivided into deeper water shelf sediments and shallower water, near shore sediments. The former, which include the Upper Cretaceous Matawan and Eocene Nanjemoy Formations, generally have transmissivities less than 10, 000 gpd/ft.; the latter, which are, in part, coarsely arenaceous, may have transmissivities ranging as high as 40, 000 gpd/ft. The Aquia and Piney Point aquifers are examples.

Nonmarine (Fluvial) Lithotope

A. Potomac Group. As reviewed by Beerbower (1964, p. 34), nonmarine fluvial sediments exhibit a complex range of sedimentary facies. These may be summarized as follows:

Environmental Elements	Predominant Lithologic Characteristics
1. Channel	
a. Floor	gravel to sand
b. Point Bar	gravel to sand
2. Natural Levee	fine sand to silt
3. Abandoned Channel	silt, clay and organic material
4. Flood Plain	
a. Back Swamp	silt, clay and some organic material
b. Swamp	clay and large amounts of organic material
c. Lakes	clay and some organic material

In modern environments alluvial facies can be readily identified by field mapping and closely spaced test borings. This is not possible in the case of buried ancient sediments because their facies must be studied using an open, loosely controlled data grid. Generally speaking, however, major facies trends of regional significance can be described using subsurface data.

An example is a subsurface study by Hansen (1969) of the Potomac Group, an alluvial sequence of lensoidal sands and clays. Lithologic properties such as sand percentage, sand thickness, and transmissivity were shown to decrease southward from Baltimore to the Potomac River. Additional data demonstrate a similar trend northward to the Chesapeake and Delaware Canal in Cecil County, Maryland (e.g., Sundstrom, et al., 1967). These data suggest that the updip portions of the Potomac Group contain facies characteristic of a fluvio-deltaic system having its axial drainage in the vicinity of Baltimore. The gravelly sands occupying the axial portion of this system are thick, permeable units, having vertical profiles diagnostic of sediments deposited by a braided bed load river (Hansen, 1969, fig. 11). These deposits are areally extensive and occur as irregularly bounded sheets. Both north and south the Potomac Group changes to a marsh or swamp facies. As Kolb and Van Lopik (1966, p. 58) note these facies usually contain less than 20 percent sand, and this is generally confined to abandoned drainage channels. Because low gradient suspended load channels draining a swamp lack the energy necessary for lateral erosion and migration, their stratigraphic record upon abandonment consists of fine sandy pods or ribbons, rather than areally extensive coarse sandy sheets (Schumm, 1968). Thus, the deposits of suspended load streams do not make productive aquifers.

Figures 4 and 5 show the transmissivity trends that characterize the Patuxent and Patapsco Formations of the Potomac Group. Because of their depositional histories, both formations exhibit greater transmissivity values in the Baltimore-Anne Arundel County area (greater than 25, 000 gpd/ft.) than north and south where transmissivities decrease to less than 20, 000 gpd/ft. and, in some cases, consistently less than 5, 000 gpd/ft. (Charles County, for example; Slaughter and Otton, 1968).

The structural dip (east and southeast) of the Potomac Group in Maryland is subparallel to the former paleoslope (i. e., direction of depositional transport). In the broadest terms, sediments of the nonmarine (fluvial) lithotope exhibit lithologic trends normal to the strike direction of the paleoslope. In the case of the Potomac Group, trend arrows shown in Figures 4 and 5 suggest a southeasternly transmissivity tract originating near Baltimore.

The downdip portions of the Potomac Group are inadequately explored. Thus, the extent to which the transmissivity tracts shown in Figures 4 and 5 extend southeastward is unkown. Modern fluvio-deltaic sequences are constructed as a series of imbricated lobes, reflecting shifts in axial drainage (Scruton, 1960, p. 100-102). To some extent the Potomac Group may also have been deposited in this manner. Therefore, beneath the Delmarva Peninsula transmissivity tracts not directly related to the one described above, may characterize the Potomac Group.

B. Plio-Pleistocene Sediments. Plio-Pleistocene sediments are widely distributed over the Maryland Coastal Plain, overlying unconformably the truncated edges of older Tertiary and Cretaceous formations. The nonmarine Columbia Group consists of coarse-grained sand and gravel, occasionally interbedded with finer textured materials. The sediments are largely the product of bed load deposition by braided streams coalescing across a wide flood plain.

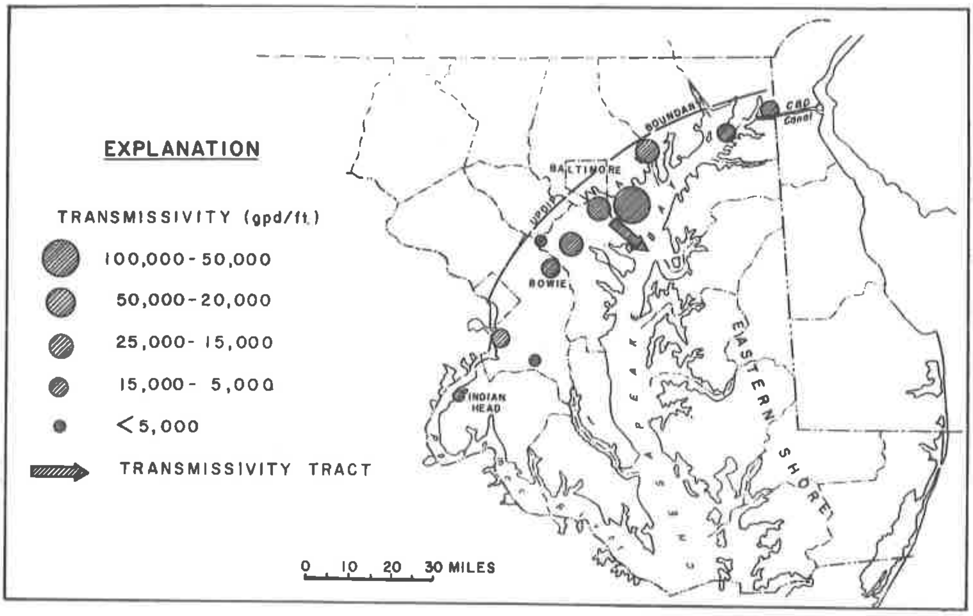


Figure 4. Map showing transmissivity distribution of the Patuxent Formation in Maryland.

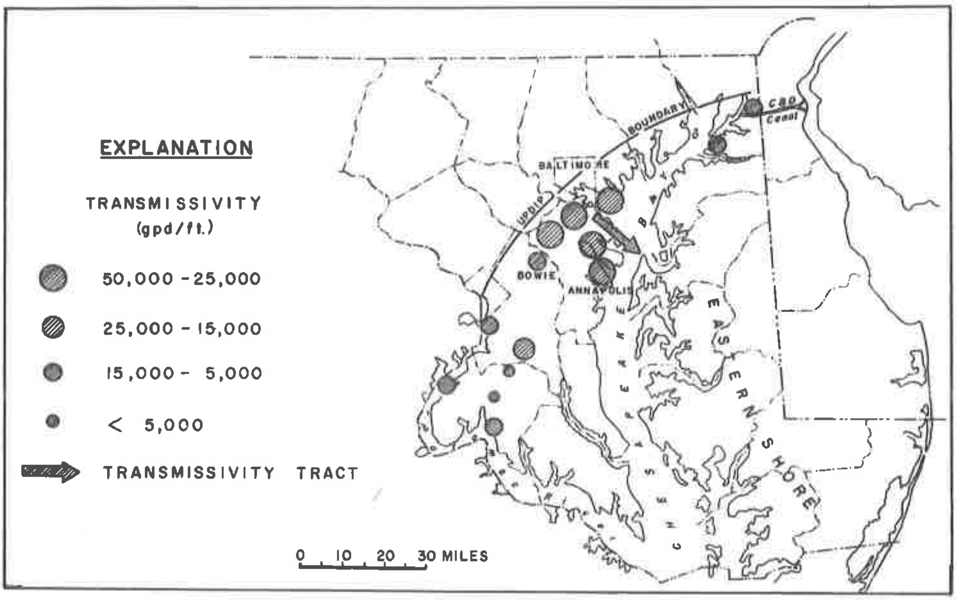


Figure 5. Map showing transmissivity distribution of the Patapsco Formation in Maryland.

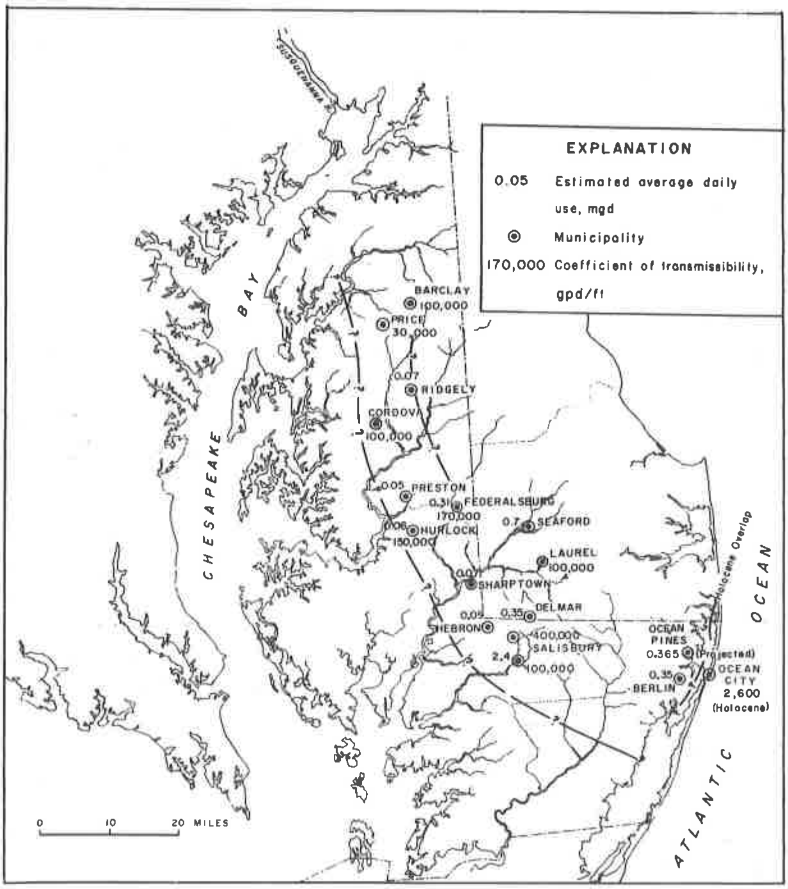


Figure 6. Map showing transmissivity distribution and municipal useage of the Plio-Pleistocene Series.

Being surficial the sediments function as a water table aquifer. In the upland areas west of Chesapeake Bay Plio-Pleistocene sediments are thin and often drained during dry periods. Transmissivity values are low; in Charles County for example, values are reported to be less than 1,000 gpd/ft. (Slaughter and Otton, 1968, p. 39).

Plio-Pleistocene sediments are of greatest importance on the Eastern Shore, particularly in a broad tract extending from Queen Annes County to Worcester County, and including adjacent parts of Delaware (Rasmussen and others, 1960). Here the saturated sediments are relatively thick with reported transmissivity values ranging between 30,000 gpd/ft. in Queen Annes County and 400,000 gpd/ft. in Wicomico County (Mack and Thomas, 1968, p. 53). Maryland communities pumping from the Plio-Pleistocene aquifer are confined to the belt depicted in Figure 6.

The southeast trending transmissivity tract described by the Columbia Group sediments points toward the existence of a major consequent stream of pre-Wisconsin age. The belt of high transmissivity values shown in Figure 6 lies along a projection extending from the mouth of the present Susquehanna River to the Washington and Norfolk Submarine Canyons, features that scallop the continental shelf and mark major fluvial outfalls (Belding and Holland, 1970).

Marine Lithotope

Visher (1965, p. 56) suggests that sedimentation in the neritic marine environment results in two first-order facies. These are:

Environmental Elements	Predominant Lithologic Characteristic
1. Neritic (Below Wave Base)	clay and silt
2. Neritic (Above Wave Base)	silt and sand

The clayey nature of marine beds deposited below wave base results in their being poor aquifers or aquicludes. In Maryland portions of the Nanjemoy, Brightseat, and Matawan Formations represent this facies; their transmissivities are consistently less than 10,000 gpd/ft. (Otton, 1955; Overbeck and Slaughter, 1958).

The sandy facies of both the Aquia and Piney Point Formations were deposited above wave base during periods of marine regression (Figure 1). Their coarse texture suggests inner neritic to, perhaps, littoral deposition. Both formations consist of fine to coarse grained, glauconitic quartz sand, containing subordinate beds of greenish glauconitic clay and indurated, calcite cemented shell debris. Transmissivity tracts characteristic of the Aquia and Piney Point Formations tend to parallel the strike direction, reflecting the action of winnowing and redistribution by long shore currents.

A. Aquia Formation. On the Western Shore the Aquia Formation outcrops in an irregularly-shaped belt extending from Sandy Point in Anne Arundel County to Indian Head in Charles County; It has a maximum width of about ten miles. On the Eastern Shore the Aquia is generally buried beneath a Pleistocene veneer, outcropping only in stream valleys. The subcrop trend traverses Kent and Cecil Counties before entering Delaware.

Sandy beds of the Aquia Formation are thickest beneath southern Queen Annes County and northern Talbot County, being 165 feet at Wades Point. In Southern Maryland the cumulative thickness of Aquia sand beds is less, being generally less than 100 feet in Charles, Calvert, and St. Mary's Counties. Downgradient a pronounced sandy to clayey facies change occurs. As a consequence the Aquia section is chiefly clayey (or silty) at Point Lookout, Cambridge, and Denton where it no longer functions as an aquifer.

The downgradient, sandy to clayey facies change is common in the stratigraphic record, the former representing the depositional product of a near shore, littoral to shallow marine environment, the latter a deeper, marine setting, perhaps below wave base. The Aquia facies change extends from north of Point Lookout to north of Dover, Delaware, subparalleling the strike of the formation.

Aquia transmissivity values range from less than 1,000 gpd/ft. to nearly 40,000 gpd/ft. The highest values occur in an elliptical-shaped tract underlying most of southern Queen Annes and northern Talbot Counties (Figure 7). Representative values include 37,500 gpd/ft. at Queenstown (Queen Annes County) and 25,000 gpd/ft. at Wades Point (Talbot County). The long-axis of the Aquia transmissivity tract trends northeasterly, subparalleling both the updip outcrop belt and the downdip facies change. Values decrease along strike both northeast and southwest of the high transmissivity tract, as for example at Smyrna, Delaware (16,500 gpd/ft.) (Sundstrom and Pickett, 1968) and Chesapeake Beach in Calvert County (10,000 gpd/ft.) (Otton, 1955).

In Southern Maryland values range from less than 1,000 gpd/ft. at La Plata in Charles County to 10,000 gpd/ft. at Chesapeake Beach in Calvert County. In southern Calvert and St. Marys Counties Aquia transmissivity values dimish downdip toward the facies boundary, being, for example, 8,000 gpd/ft. at Lusby and 7,500 gpd/ft. at Lexington Park (Otton, 1955).

The Aquia transmissivity tract has a significantly different orientation than the tracts characteristic of the Potomac and Plio-Pleistocene aquifers. The Aquia tract is parallel to strike, the other tracts perpendicular. The Aquia tract reflects marine processes which act subparallel to an ancient strand line. Conversely the fluvial tracts reflect processes acting subnormal to an ancient strand line.

Although oriented differently, the tracts, nevertheless, occur along a line extending southeasterly through the central portion of the Maryland Coastal Plain. This supports the assumption of a long-existing (Mesozoic-Tertiary) sediment source, perhaps representing a proto-Susquehanna drainage system.

B. Piney Point Formation. Like the Aquia Formation, the Piney Point is a marine unit that was deposited in a sublittoral to shallow neritic environment. The updip portion of the formation consists of sand and shell detritus. Downdip the Piney Point becomes finer-grained, exhibiting the sandy (above wave base) to clayey (below wave base) facies change characteristic of many marine formations.

The Piney Point Formation was deposited during a time of marine regression. The formation coarsens upwards, reflecting shoaling conditions and a seaward shift of the strandline. As the strandline migrated eastward to some indeterminate position, erosion replaced deposition in updip areas, thus introducing an unconformity into the stratal sequence. As a result, the Piney Point Formation is truncated

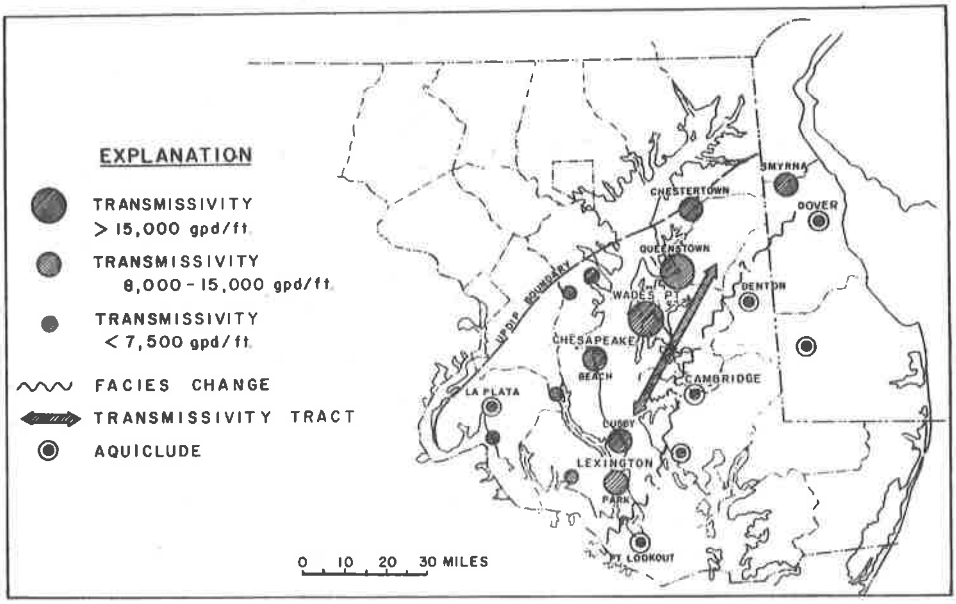


Figure 7. Map showing transmissivity distribution of the Aquia Formation in Maryland.

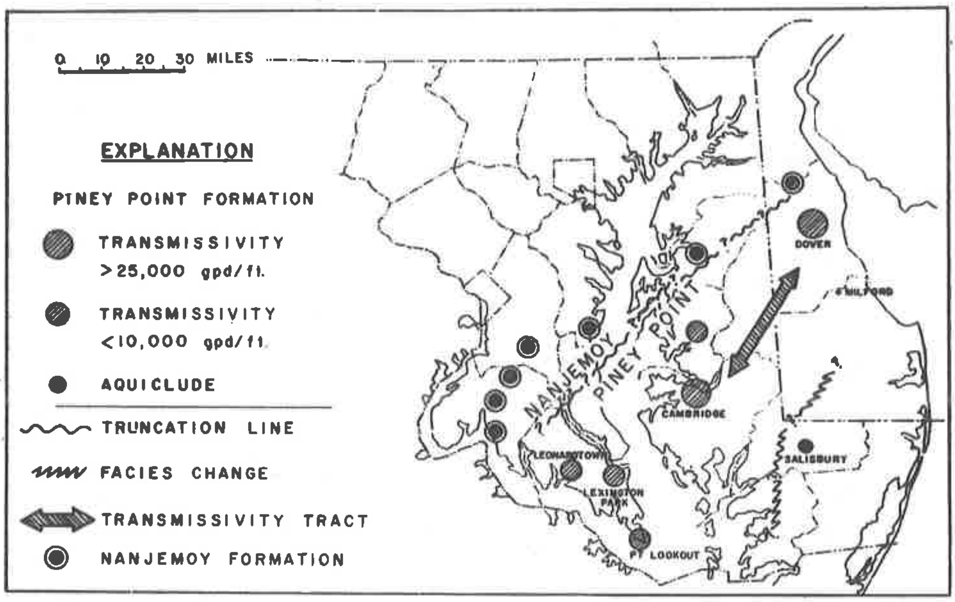


Figure 8. Map showing transmissivity distribution of the Piney Point Formation in Maryland.

in the subsurface and, consequently, does not outcrop.

The Piney Point Formation is truncated in the subsurface along an irregular line extending from north of Leonardtown in St. Mary's County to northern Carolina County. The updip portion of the Piney Point Formation was erosionally bevelled during an Oligo-Miocene withdrawal of the sea (Figure 1). During the following marine transgression (Middle Miocene) truncated beds of the Piney Point Formation were overlapped by basal strata of the Calvert Formation.

Sandy beds of the Piney Point Formation thicken northeasterly in a pod-like accumulation extending roughly from Point Lookout (90 feet) in St. Mary's County to Dover (220 feet) in Kent County, Delaware. The sandy facies of the Piney Point Formation exceeds 100 feet at Cambridge in Dorchester County, Denton in Carolina County, and Milford in Sussex County, Delaware. The Piney Point sands thin updip by erosional truncation, downdip by facies change.

Reported Piney Point transmissivity values range from less than 5,000 gpd/ft. to values as high as 40,000 gpd/ft. The highest values are reported from the Eastern Shore and coincide, not surprisingly, with the area of thickest sand accumulation (Figure 8). High values are reported at Cambridge (40,000 gpd/ft.) and Dover, Delaware (37,500 gpd/ft.) (Rasmussen and Slaughter, 1957; Sundstrom and Pickett, 1968).

The long-axis of the Piney Point transmissivity tract trends northeasterly, subparalleling the Aquia tract; however, the Piney Point tract is offset southeasterly, traversing Dorchester and southern Caroline Counties, rather than southern Queen Annes and northern Talbot Counties.

Reported transmissivity values are appreciably less in Southern Maryland as, for example, at Leonardtown (6,500 gpd/ft.), Lexington Park (6,500 gpd/ft.), and Point Lookout (4,000 gpd/ft.) (Otton, 1955).

Like the Aquia, the Piney Point transmissivity tract (Figure 8) is subparallel to the strike of the formation. Although offset to the southeast, the Piney Point tract also occupies the central portion of the Maryland Coastal Plain. Both Aquia and Piney Point transmissivity values decrease into Southern Maryland and the lower Eastern Shore. Unlike the Aquia, however, the sandy facies of the Piney Point Formation continues to exhibit high transmissivity values northeasterly across Delaware, not diminishing until southern New Jersey (Richards and others, 1962, p. 32).

Fluvio-Marine Lithotopes

The Fluvio-marine lithotope represents a marginal facies separating the two types discussed above. Visher (1965, p. 45) suggests that fluvio-marine sediments can be subdivided into the following facies:

Environmental Element	Predominant Lithologic Characteristic
1. Tidal Flat	clay and silt, occasionally sandy channels
2. Lagoon-Bay	clay and silt, occasionally sandy rims
3. Dune	sand, fine to medium
4. Littoral (beach area between extreme high and low tide limits)	sand, medium to coarse

Fluvio-marine sediments may be associated with either marine transgression or marine regression (Curray, 1964, p. 191). Both examples occur in Maryland; the Upper Cretaceous Magothy Formation is representative of transgression, the Miocene Yorktown Formation of regression. Generally speaking, regressive or offlap sequences are better preserved in the stratigraphic record than transgressive or onlap sequences (Visher, 1965, p. 55; Swift, 1967, p. 444). The fluvio-marine lithotope in a regressive sequence tends to be thicker because most of the environmental elements listed above are buried, rather than eroded, during offlap sedimentation. On the other hand, relatively thin, but areally extensive, littoral zone sediments characterize the transgressive sequence. During onlap wave erosion often destroys the marginal facies represented by beach, lagoonal, and tidal flat sediments; consequently these facies are usually preserved, if at all, as sporadic lenses.

A. Magothy Formation. The transgressive Magothy Formation overlies the fluvio-deltaic Potomac Group and underlies the neritic marine Matawan-Monmouth sequence (Figure 1). It is areally extensive, occurring over much of the Coastal Plain as a blanket-like deposit. Ground water is pumped from the Magothy in parts of Charles, Prince Georges, Calvert, and Anne Arundel Counties as well as the Eastern Shore counties located west of Delaware.

The Magothy lithology is coarsely arenaceous, commonly lignitic, and rarely shelly; although clayey interbeds are common, they are generally lensoidal and thin (less than 10 feet). Most of the unit is suggestive of littoral deposition with the other facies of the fluvio-marine lithotope being subordinate. The Magothy Formation rarely exceeds 75 feet, except in Anne Arundel County where it is considerably thicker.

Over much of the Coastal Plain reported transmissivity values for the Magothy aquifer range between 5, 000 gpd/ft. and 15, 000 gpd/ft. Generally speaking, no regional transmissivity tracts are discernible (Figure 9). This suggests that littoral reworking along a transgressive strand zone has obliterated, except locally, the depositional patterns resulting from fluvial input to the basin. Tracts parallel to the depositional strike are not noticeably strong because the unit has transgressed

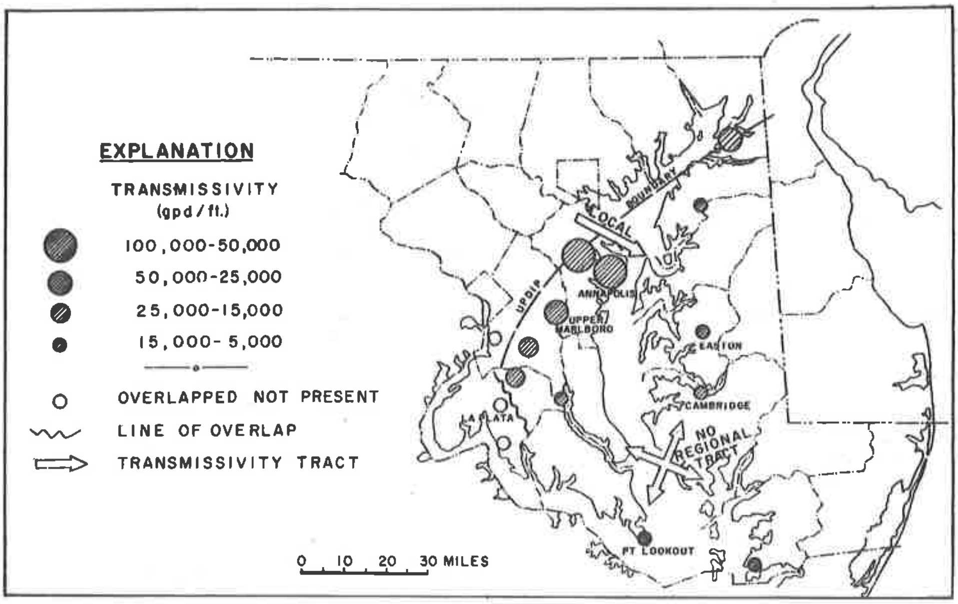


Figure 9. Map showing transmissivity distribution of the Magothy Formation in Maryland.

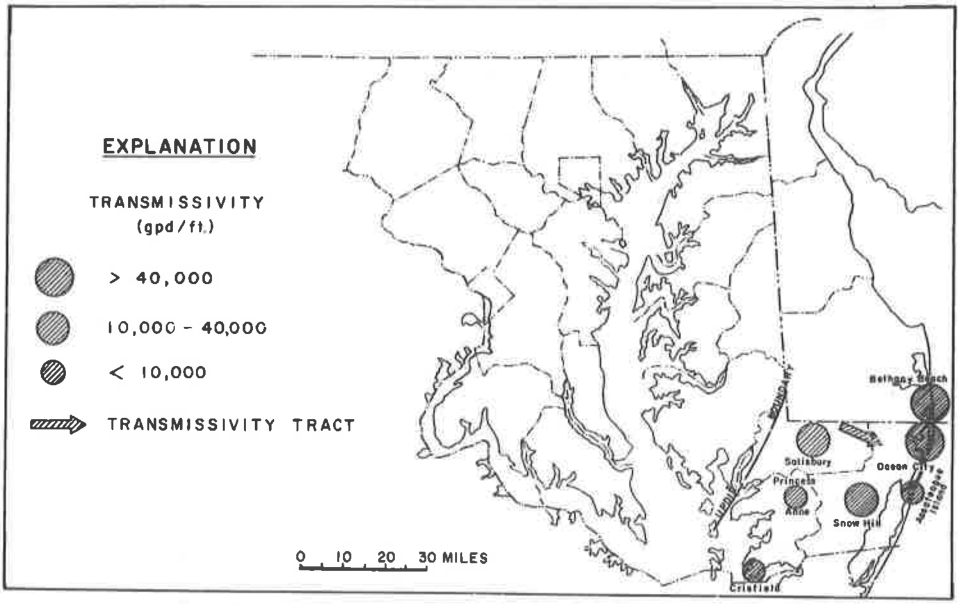


Figure 10. Map showing transmissivity distribution of the Yorktown Formation in Maryland.

most of the Coastal Plain, apparently without major stillstands.

A local, but important, exception is well documented in eastern Anne Arundel County (Mack, 1962) (Figure 9). Here the Magothy Formation exceeds 150 feet in thickness and consists chiefly of coarse sand and, occasionally, gravel. Transmissivity values generally exceed 50,000 gpd/ft. and may reach as high as 100,000 gpd/ft.

As shown in Figure 9, the orientation of this locally significant transmissivity tract is perpendicular to the strike of the paleoslope, a configuration diagnostic of nonmarine (fluvial) influence. During Magothy deposition, this area was proximal to a major source of fluvially derived sediments. Because littoral activity was insufficient to redistribute the arenaceous fraction, a transmissivity tract is preserved in the stratigraphic record.

Interestingly, the Magothy tract coincides in general with those characteristic of the Potomac Group (Figures 4 and 5). Evidently the fluvial system that dominated Potomac sedimentation persisted into Magothy time, having its greatest effect nearest the Piedmont source area.

B. Yorktown Formation. The Yorktown Formation of Upper Miocene age is restricted in Maryland to the eastern Delmarva Peninsula where it subcrops beneath a mantle of Plio-Pleistocene sediments. Although stratigraphic knowledge of this unit is incomplete, lithologic and faunal data discussed by Rasmussen and Slaughter (1955, p. 93) suggest a regressive, fluvio-marine lithotope, containing, perhaps, neritic intertongues.

The Yorktown Formation averages 250 feet in thickness (Rasmussen and Slaughter, 1955, p. 43; Slaughter, 1962, p. 3). Although it is a lensoidal unit, correlations are possible, at least locally, because portions of the formation tend to be either predominantly arenaceous (Manokin and Pocomoke aquifers) or predominantly argillaceous. The gray, quartzose sand beds are usually fine to medium grained, but may be coarse or gravelly in places; shell beds have been reported, but only rarely. The argillaceous beds are dark gray to greenish black, commonly shelly, and occasionally lignitic.

The highest transmissivity values for the Manokin Aquifer (a subdivision of the Yorktown Formation) are reported from the Ocean City area of Worcester County and adjacent areas of Sussex County, Delaware (Figure 10). Values decrease southward being generally less than 10,000 gpd/ft. at Princess Anne and beneath much of Assateague Island; however, Snow Hill (Worcester County), where a value of 40,000 gpd/ft. was reported, is an exception.

Spot data suggest that a transmissivity tract extends between Salisbury in Wicomico County and Ocean City (Figure 10). Such an orientation is normal to the depositional strike of the Manokin Aquifer. It is interesting to note that the overlying, fluvially-derived, Plio-Pleistocene sediments exhibit a similar trend, suggesting that pre-Holocene drainage may have been of the consequent-type, flowing

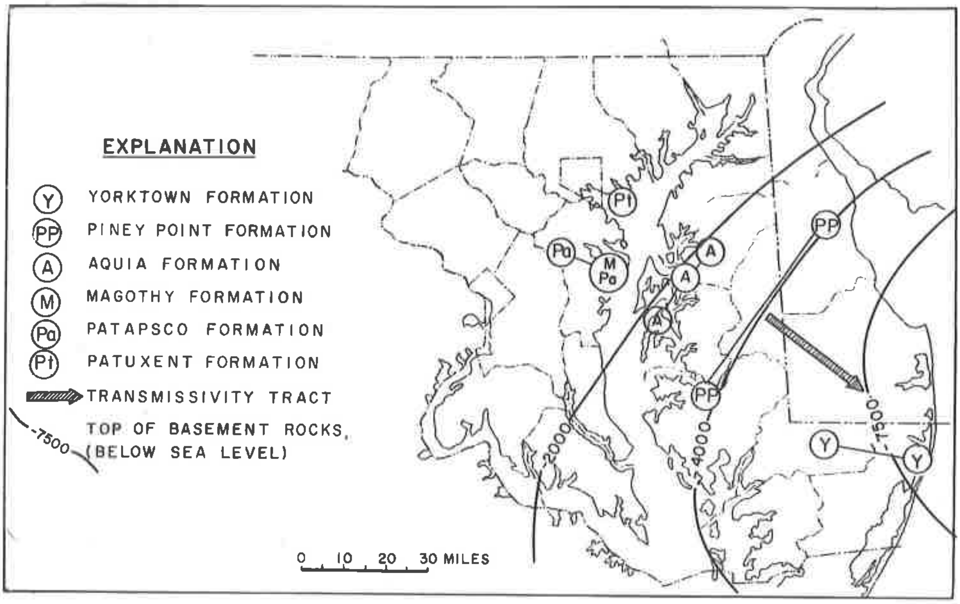


Figure 11. Map showing relationship between transmissivity tracts of major Coastal Plain aquifers and axis of structural trough.

southeasterly across the present Delmarva Peninsula.

CONCLUSIONS

An inspection of subsurface data from the Maryland Coastal Plain suggests the following:

1. The Coastal Plain sediments exhibit a tripartite transgressive-regressive cyclicity that has resulted in the repetition of three major lithotopes. These are: Nonmarine (fluvial) (Potomac Group, Plio-Pleistocene Series), Fluvio-marine (Magothy Formation, Marlboro Clay, Yorktown Formation), and Marine (Monmouth-Matawan Formation, Nanjemoy Formation, pre-Yorktown Chesapeake Group). The highest transmissivity values are associated with the nonmarine lithotope, the lowest with the marine.

2. Transmissivity tracts characteristic of the marine lithotope are generally parallel to the depositional strike; conversely, in aquifers of the fluvial lithotope tracts are generally parallel to the depositional slope. The former reflects the predominance of marine redistribution, the latter of fluvial input.

3. Transmissivity tracts conform generally with the axial portion of a major landward extension of the Baltimore Canyon trough (Maher, 1967). This supports the assumption of a long existing (Mesozoic-Tertiary) sediment source related to a proto-Susquehanna drainage system (Figure 11).

REFERENCES

Back, W., 1966, Hydrochemical facies and ground-water flow patterns in northern part of Atlantic Coastal Plain: U. S. Geol. Survey Prof. Paper 498-A, 41 p.

Beerbower, J. R., 1964, Cyclothems and cyclic depositional mechanisms in alluvial plain sedimentation: In Symposium in cyclic sedimentation, ed. D. F. Merriam, Kansas Geol. Survey Bull. 169, Vol. 1, p. 31-42.

Belding, H. F., and Holland, W. C., 1970, Bathymetric maps eastern continental margin U. S. A., Atlantic Ocean north of Cape Hatteras: Amer. Assoc. Petrol. Geologists, Tulsa, Okla.

Bennett, R. R., and Collins, G. G., 1952, Brightseat Formation, a new name for sediments of Paleocene age in Maryland: Wash. Acad. Sci. Jour., v. 42, p. 114-116.

__________, and Meyer, R. R., 1952, Geology and ground-water resources of the Baltimore area: Maryland Dept. Geology, Mines and Water Resources Bull. 4, 573 p.

Clark, W. B., 1916, The Upper Cretaceous deposits of Maryland, in Upper Cretaceous volume: Maryland Geol. Survey, p. 1-109.

__________, Bibbins, A. B., and Berry, E. W., 1911, The Lower Cretaceous deposits of Maryland, in Lower Cretaceous volume: Maryland Geol. Survey, p. 1-98.

__________, Mathews, E. B., and Berry, E. W., 1918, The surface and underground water resources of Maryland, including Delaware and the District of Columbia: Maryland Geol. Survey, Spec. Pub., V. 10, Part 2, 542 p.

Cleaves, E. T., Weaver, K. N., Edwards, J., and Glaser, J. D., 1968, Geologic Map of Maryland: Maryland Geol. Survey.

Cooke, C. W., 1952, Sedimentary deposits in Geology and water resources of Prince Georges County: Maryland Dept. Geology, Mines and Water Resources Bull. 10, p. 1-53.

Curray, J. R., 1964, Transgression and regressions: in Shepard Commemorative Volume ed. R. L. Miller, MacMillan Co., New York, p. 175-203.

Darton, N. H., 1896, Maryland in Artesian well prospects in the Atlantic Coastal Plain region: U. S. Geological Survey Bull. 138, p. 124-135.

__________, 1902, Preliminary list of deep borings in the United States, Part I, Alabama-Montana: U. S. Geological Survey Water-Supply Paper 57, 60 p.

__________, 1948, The Marlboro Clay: Econ. Geol. v. 43, p. 154-155.

Drobnyk, J. W., 1965, Petrology of the Paleocene-Eocene Aquia formation of Virginia, Maryland and Delaware: Jour. Sed. Petrology, v. 35, p. 626-642.

Dryden, L., and Overbeck, R. M., 1948, The physical features of Charles County: Maryland Dept. Geology, Mines and Water Resources, 267 p.

Dunbar, C. O., and Rogers, J. R., 1957, Principles of stratigraphy: John Wiley and Sons, Inc., New York, 356 p.

Emery, K. O., 1967, The Atlantic continental margin of the United States during the past 70 million years: Geol. Assoc. Canada Sp. Paper No. 4, p. 53-70.

Gibson, T. G., 1962, Benthonic foraminifera and paleoecology of the Miocene deposits of the Middle Atlantic Coastal Plain: unpubl. PhD Dissertation, Princeton University, 194 p.

Glaser, J. D., 1967, Nonmarine Cretaceous sedimentation in the Middle Atlantic Coastal Plain: unpubl. PhD Dissertation, Johns Hopkins University, 359 p.

__________, 1968, Coastal Plain geology of Southern Maryland: Maryland Geol. Survey Guide Book No. 1, 56 p.

Goldman, M. E., 1916, The petrography and genesis of the sediments of the Upper Cretaceous of Maryland, in Upper Cretaceous volume: Maryland Geol. Survey p. 111-182.

Hansen, H. J., 1966, Pleistocene stratigraphy of the Salisbury area, Maryland and its relationship to the lower Eastern Shore: Maryland Geol. Survey Rpt. Invest. No. 2, 56 p.

__________, 1969, Depositional environments of subsurface Potomac Group in Southern Maryland: Amer. Assoc. Petrol. Geol. Bull., Vol. 53, No. 9, p. 1923-1937.

Hazel, J. E., 1968, Ostracodes from the Brightseat Formation (Danian) of Maryland: Jour. Paleotology, v. 42, p. 100-142.

Jordan, R. R., 1964, Columbia (Pleistocene) sediments of Delaware: Delaware Geol. Survey Bull. 12, 59 p.

Kolb, C. R., and Van Lopik, J. R., 1966, Depositional environments of the Mississippi River deltaic plain - southeastern Louisiana, in Martha Shirley, ed., Deltas in their geologic framework: Houston Geol. Society, p. 19-61.

LeGrand, H. E., 1961, Summary of geology of Atlantic Coastal Plain: Bull. Amer. Assoc. Petrol. Geologists, v. 45, no. 9, p. 1557-1571.

Mack, F. K., 1962, Ground-water supplies for industrial and urban development in Anne Arundel County: Maryland Dept. Geology, Mines and Water Resources Bull. 26, 90 p.

__________, 1966, Ground water in Prince Georges County: Maryland Geological Survey Bull. 29, 101 p.

__________, and Thomas, W. O., 1968, Hydrologic testing of buried valley near Salisbury, Maryland: U. S. Geol. Survey Prof. Paper 600A, p. 53.

Maher, J. C., 1965, Correlation of subsurface Mesozoic and Cenozoic rocks along the Atlantic Coast: Am. Assoc. Petrol. Geol., Tulsa, Okla., 18 p.

__________, 1967, Geologic framework and petroleum potential of the Atlantic Coastal Plain and Continental Shelf: U. S. Geol. Survey open file rept., 280 p.

Maryland Geological Survey, 1970, List of Publications: Maryland Geological Survey, Baltimore, Md. 21218.

Murray, G. E., 1961, Geology of the Atlantic and Gulf Coastal Province of North America: Harper and Bros., New York, 692 p.

Nogan, D. S., 1964, Foraminifera, stratigraphy, and paleoecology of the Aquia Formation of Maryland and Virginia: Cushman Foundation for Foraminiferal Research Spec. Publ. No. 7, 50 p.

Otton, E. G., 1955, Ground-water resources of the Southern Maryland coastal plain: Maryland Dept. Geology, Mines and Water Resources Bull. 15, 347 p.

__________, and others, 1964, Water resources of the Baltimore area, Maryland: U. S. Geol. Survey Water-Supply Paper 1499-F, 105 p.

Overbeck, R. M., and Slaughter, T. H., 1958, The water resources of Cecil, Kent, and Queen Annes Counties: Maryland Dept. Geology, Mines, and Water Resources Bull. 21, 478 p.

Rasmussen, W. C., and Slaughter, T. H., 1955, The water resources of Somerset, Wicomico, and Worcester Counties: Maryland Dept. Geology, Mines and Water Resources, Bull. 16, 533 p.

______________, 1957, The water resources of Caroline, Dorchester, and Talbot Counties: Maryland Dept. Geology, Mines and Water Resources Bull. 18, 465 p.

Rasmussen, W. C., Wilkens, R. A., and Beall, R. M., 1960, Water resources of Sussex County, Delaware: Delaware Geol. Survey Bull. 8, 228 p.

Richards, H. G., 1948, Studies on the subsurface geology and paleontology of the Atlantic Coastal Plain: Acad. Natural Sci. Philadelphia Proc., v. 100, p. 39-76.

______________, Olmsted, F. H., and Ruhle, J. L., 1962, Generalized structure contour maps of the New Jersey Coastal Plain: New Jersey Geol. Survey, Geol. Rept. Series No. 4, 38 p.

Schlee, J. S., 1957, Upland gravels of Southern Maryland: Geol. Soc. America Bull., v. 68, p. 1371-1409.

Schumm, S. A., 1968, Speculations concerning paleohydrologic controls of terrestrial sedimentation: Geol. Soc. Amer. Bull., vol. 79, no. 11, p. 1573-1588.

Scruton, P. C., 1960, Delta building and the deltaic sequence: in Recent sediments, northwest Gulf of Mexico ed. F. P. Shepard, Amer. Assoc. Petrol. Geologists, p. 82-102.

Shifflett, E., 1948, Eocene stratigraphy and foraminifera of the Aquia Formation: Maryland Dept. Geology, Mines and Water Resources Bull. 3, 93 p.

Slaughter, T. H., 1962, Beach area water supplies between Ocean City, Maryland and Rehoboth Beach, Delaware: U. S. Geol. Survey Water-Supply Paper 1619-T, 10 p.

__________, and Otton, E. G., 1968, Availability of ground water in Charles County: Maryland Geol. Survey Bull. 30, 100 p.

Sundstrom, R. W., and others, 1967, The availability of ground water from the Potomac Formation in the Chesapeake and Delaware Canal area: Water Resources Center, Univ. Delaware, Newark, Del., 95 p.

__________, and Pickett, T. E., 1968, The availability of ground water in Kent County, Delaware, with special reference to the Dover area: Univ. Delaware Water Resources Center, 123 p.

Swift, D. J. P., 1967, Coastal erosion and transgressive stratigraphy: Jour. Geol. Vol. 76, p. 444-456.

United States Geological Survey, 1967, Engineering geology of the northeast corridor Washington, D. C. to Boston, Massachusetts: Coastal Plain and surficial deposits: U. S. Geol. Survey Misc. Geol. Invest. Map I-514-B.

Upson, J. E., 1966, Relationship of fresh and salty ground water in the northern Atlantic Coastal Plain of the United States: U. S. Geol. Survey Prof. Paper 550-C, p. 235-243.

Visher, G. S., 1965, Use of vertical profiles in environmental reconstruction: Am. Assoc. Petrol. Geol. Bull., vol. 49, no. 1., p. 41-61.

Wheeler, H. E., 1964, Baselevel, lithosphere surface, and time-stratigraphy: Geol. Soc. Amer. Bull., vol. 75, p. 599-610.

Wilson, A., 1967, River discharge to the sea from the shores of the conterminous United States: U. S. Geol. Survey Hydrol. Invest. Atlas HA-282.

A STRUCTURAL AND PETROGRAPHIC ANALYSIS OF THE BANK'S CREEK (NORTH CAROLINA) SERPENTINITE

By

Kenneth R. Neuhauser

and

John R. Carpenter
Department of Geology
University of South Carolina
Columbia, South Carolina 29208

ABSTRACT

A structural and petrographic study of a completely serpentinized ultramafic body in Yancey County, North Carolina indicates that an in situ serpentinization process occurred. A "mesh" texture formed by serpentine pseudomorphs of antigorite and chrysotile with minor magnetite is observed. The enclosing country rock consists of middle-grade metamorphics of quartz-plagioclase (An_{20})-biotite gneisses and amphibolitic gneisses containing a blue-green hornblende, plagioclase (An_{38}) and quartz with minor epidote, garnet and biotite.

Strikes and dips of country rock foliation surfaces and axial plane surfaces, along with lineation plunge and bearings indicate a synformal structure with a gentle southwesterly plunge. The serpentinite lies near the synformal trough axis on the lower western flank or almost upon the axis.

It is felt that an in situ serpentinization process combined with a chemical potential hypothesis serves as a model explanation for the Bank's Creek serpentinite.

INTRODUCTION

The origin of ultramafic bodies associated with orogenic belts, the so-called alpine-type ultramafics, has caused much controversy in the field of petrology, and recent investigations have brought about other additional interpretations concerning this subject. The controversy lies basically in attempting to establish a mode of emplacement to explain the diverse types of ultramafics found in this tectonic setting.

The ultramafic belt of the Southern Appalachians is generally characterized by the presence of dunites and peridotites. The Bank's

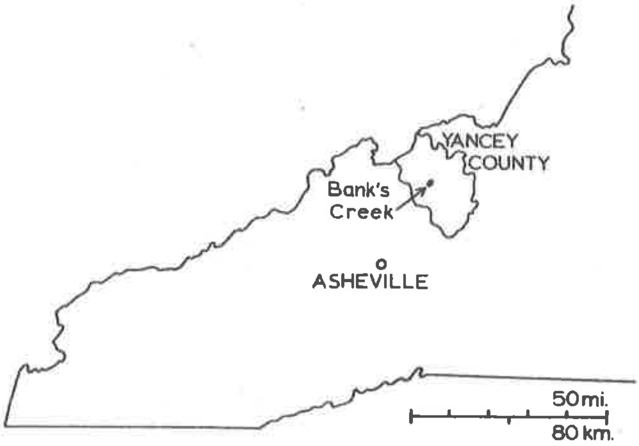


Figure 1. Index map of western North Carolina.

Creek deposit of Yancey County, North Carolina (Figure 1) is one of a very few completely serpentinized bodies in this province and this investigation attempts to establish the structural and petrologic relationships between it and the surrounding country rocks. This objective involved a complete structural and petrological study of the area; to establish the relationship between the serpentinite and the structural fabric of the country rock, and to establish the metamorphic grade of the enclosing country rock.

Previous Work

Previous petrographic and mineralogic investigations of the rocks in the Bank's Creek area were carried out in conjunction with economic investigations on the feasibility of mining such deposits as chromite, graphite, soapstone, talc, corundum, marble, and magnetite (Keith, 1905).

Acknowledgments

We greatly appreciate the comments and criticisms of Donald T. Secor and W. E. Sharp of the University of South Carolina who read various drafts of this manuscript. We would also like to acknowledge financial support from the North Carolina Department of Conservation and Development, Division of Mineral Resources, Stephen Conrad, State Geologist, for field expenses. The investigation was partially financed through National Science Foundation grant GA-4518 to J. R. Carpenter.

THEORIES ON ORIGIN

Alpine-type serpentinites apparently can originate either by intrusion of "solid" serpentine, or by in situ hydration of dunites and peridotites. The solid emplacement hypothesis suggests that the rocks were emplaced as masses of more or less plastic serpentine or partially serpentinized olivine. This solid (plastic) intrusion of serpentine would be characterized by flow structures, shear zones or other evidence of plastic flow, and this is commonly seen in tectonically emplaced serpentinites, such as the Joaquin Ridge serpentinite in the southern coast ranges of California (Cowan and Mansfield, 1970).

In the case of tectonic emplacement the serpentinite is found in a totally different environment from that in which it presently is observed, and one must attempt to ascertain the origin of the serpentinite. Dietz (1963) accepted the idea that alpine serpentinites were tectonically emplaced in the solid state and suggested that they may have been fragments of the sea floor derived from the oceanic layer. Then as the sea floor thrust toward the continent, probably coincident with sea-floor spreading, the continental-rise prism was folded into a eugeosynclinal prism. Subsequently, pods of serpentine derived from the sea-floor sima underlying the eugeosyncline would be caught up in the folding process, and later occur as linear swarms of en echelon masses along axes of old and new belts of folded mountains.

Hess (1960) proposed that the oceanic layer three consisted of partially serpentinized peridotite. This statement was based on: (1) the sampling of serpentinized peridotite on scarps of the mid-ocean ridges and the north slope of the Puerto-Rico Trench, and (2) a similarity with serpentinites found in the island arc system and within the basement under Puerto-Rico. Cann (1968) however, has argued that this oceanic layer supplying bits to the folded prisms is amphibolite. To date, further analyses of the composition of the oceanic layer is required, and no definite conclusion has been reached as to the exact material composing it except to categorize it as being ultrabasic (Vogt, et al., 1969). Le Pichon (1969) feels that there are no definitive arguments either for or against the serpentinite composition of layer three.

Still another possibility is that these serpentinites represent hydrated upper mantle material. Chase and Hersey (1968) argue that these ultramafic bodies came from below layer three and were intruded along faults. Hess (1966), Vogt, et al., (1969) also believed that parts of the mantle are caught up in the compression of the earth's crustal layers and do not return to the mantle.

There is another group of alpine-type serpentinites that exhibit a "mesh" texture, where serpentine pseudomorphically replaces olivine and/or pyroxene retaining the original grain outline. In general these exhibit little or no evidence of tectonic deformation except at the peripheries. This has been reported from the serpentinites in Lancaster and Chester Counties, Pennsylvania and northern Cecil County, Maryland (Lapham, 1967); the Roxbury serpentinites of Vermont (Jahns, 1967); and also the Burro Mountain serpentinites in California (Page, 1967). Clearly these bodies, which can still be described as alpine-type, represent in situ hydration of pre-existing olivine and/or pyroxene-rich rock. Serpentinite thus formed probably would not reflect the complete emplacement history of the peridotite or dunite from which they have been derived.

In the case of in situ serpentinization one must be concerned with the nature of the hydration reaction. Did volumetric and/or compositional changes accompany serpentinization? The two reactions most generally cited are:

(1) $3\ Mg_2SiO_4 + 4H_2O + SiO_2 = 2Mg_3Si_2O_5(OH)_4$

and

(2) $5Mg_2SiO_4 + 4H_2O = 2Mg_3Si_2O_5(OH)_4 + 4MgO + SiO_2$

Reaction (1) requires that a very large volume increase percentage take place, but does not require extensive metasomatism only the introduction of silica and water. Reaction (2) is a constant volume reaction resulting in large quantities of MgO and SiO_2 being given off to the country rock.

Thayer (1966, 1967) strongly adheres to an essentially constant-volume metasomatic process to explain in situ serpentinization. Madison and Condie (1969) take a more neutral position, arguing that the serpentinite of the Webster-Addie (North Carolina) complex involved both volumetric and compositional changes. They suggest another reaction:

(3) $5Mg_2SiO_4 + SiO_2 + 6\ H_2O = 3Mg_3Si_2O_5(OH)_4 + MgO$

with a concomitant volume increase of 50 percent, if the reaction proceeds to completion.

PETROGRAPHY

Country Rock

The country rocks enclosing the Bank's Creek serpentinite are part of a middle-grade sequence of meta-igneous and meta-sedimentary rocks known collectively as the Ashe Formation (Rankin, 1969). Two dominant lithologies were observed in this area, a quartz-plagioclase (An_{20})-biotite gneiss and an amphibolite gneiss containing a blue-green hornblende, plagioclase (An_{38}), and quartz with minor epidote, garnet and biotite. The modal analyses of these two rock types vary both regionally and locally and tend to grade into one another both laterally and vertically.

In general the gneisses are more quartz rich near the serpentinite deposit but this phenomenon is also observed away from the serpentinite along strike. Analysis of the plagioclase by the albite twin law indicates an over-all range of An_{18} to An_{43} with the average centering around An_{26}. It was noted that near the serpentinite the distinct gneissose foliation formed by mineral banding gives way to a more schistose foliation. Hornblende-rich gneisses were generally observed farther from the serpentinite and form indistinct bands paralleling regional strike. Foliation is less distinct in the amphibolitic gneisses due to the decrease of muscovite and biotite. It was observed that an increase in the Ca-content in the plagioclase occurs as the country rock becomes more amphibolitic.

Serpentinite

Thin section analysis of serpentinite samples indicate a "mesh-like" texture, i.e., reticulate fibers of serpentine pseudomorphically replacing some other mineral. The mineral replaced in this case appears to be olivine, based on the crystal grain outline. Fractures also exist in and along the pseudomorphic structures. They are irregularly curved and tend to have fine-grained magnetite along these fractures. It is believed that this magnetite separates during the alteration of olivine to serpentine and moves into the fractures (Moorehouse, 1959). In some portions of the serpentinite small veinlets of chrysotile asbestos cut across the structure. Based on thin section description and x-ray diffraction analysis the pseudomorphic serpentine is antigorite, and the asbestos veinlets are chrysotile.

Photomicrographs of tectonically emplaced serpentinites from Japan were observed to compare any textural likenesses or differences. No "mesh-like" texture was observed in the tectonically emplaced serpentinites but rather a layered texture reflecting flow. Flow structures characteristic of tectonically emplaced serpentinite bodies were not observed in the Bank's Creek complex. The evidence suggests very strongly that this deposit formed by means of an in situ hydration process.

Based upon textural, x-ray diffraction, and float data, the four serpentinite outcrops sampled appear to be part of the same deposit. Lines along regional strike were traced between each serpentinite outcrop, and the distribution indicated that a talc belt parallels and encircles the serpentinite. Actual areal dimensions were difficult to ascertain by this method but they roughly suggest a pod-like structure (Figure 2). Two of the serpentinite outcrops studied were once mine shafts, now filled, excavated by Johns Mansville Mining Company in the early 1940's. One deep shaft encountered talc and country rock at 65 feet, but the true depth of the Bank's Creek deposit is not known.

STRUCTURE

In order to ascertain the attitude of the Bank's Creek serpentinite a structural investigation of the surrounding country rock was

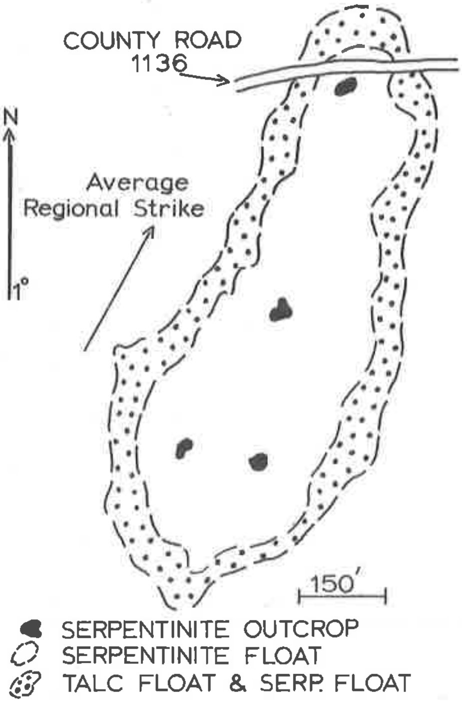


Figure 2. Generalized geologic map of serpentinite complex.

conducted. Data were uniformly obtained within the area to accord a more accurate representation of the fabric elements. Structural data were plotted on standard 20 cm. Schmidt stereographic nets. Structural interpretations of the stereographic projection data were based on Turner and Weiss's structural analyses of metamorphic tectonites (1963). Their metamorphic tectonite analyses offered a method applicable to interpreting the country rock surrounding the Bank's Creek deposit.

Field observation showed that the schistose and gneissose country rock has been intensely deformed. Strikes and dips were measured on foliation surfaces formed by platy mineral banding. It was observed at several exposures, however, that this "foliation" is not a "bedding plane" foliation but rather an axial plane foliation. Competent, thin arenaceous units in some exposures show evidence of tight isoclinal folding. Associated well-foliated micaceous units in these exposures do not show evidence of this folding. The compositional banding between these units is interpreted to be a "bedding" foliation. As a result of this folding episode, "bedding" foliation in the less competent mica-

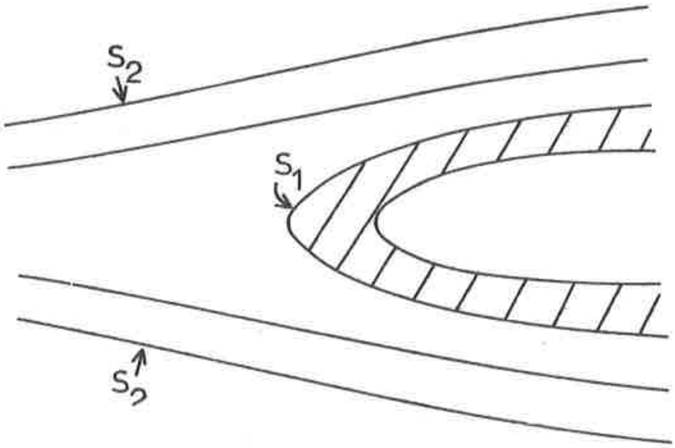


Figure 3. Diagrammatic sketch showing relationships between compositional banding (S_1) and axial plane foliation (S_2).

ceous units was destroyed. The foliation that developed in the mica units parallels the axial planes of the tight isoclinal folds in the arenaceous units. Therefore two planar fabric elements have been observed in the country rock and can be distinguished as follows:

S_1 - compositional banding between meta-arenite and mica-schists

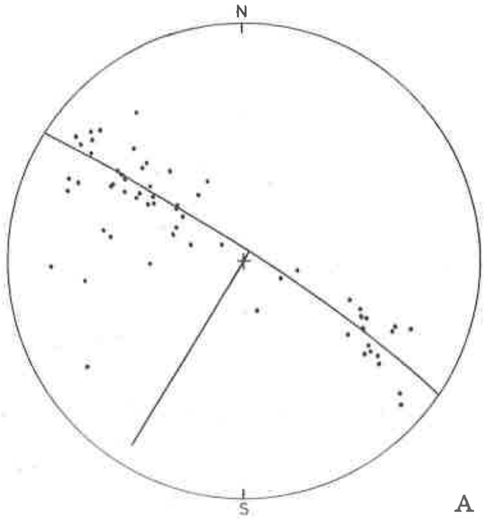
S_2 - axial plane foliation in mica schists.

See sketch in Figure 3 for details and relationships.

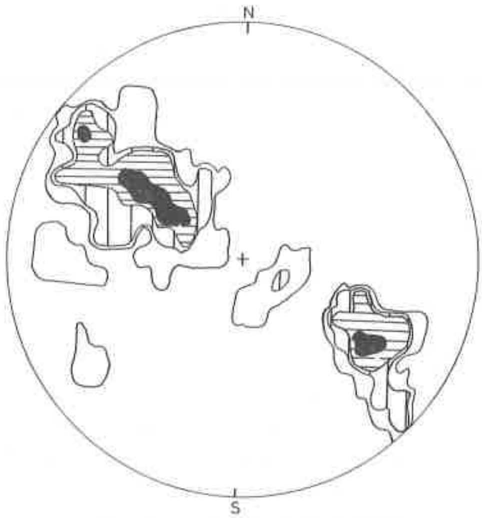
The isoclinally folded meta-arenaceous units are generally thin and sparsely distributed throughout the predominant micaceous units and therefore, most observed "foliation" surfaces are either S_1 surfaces, or S_2 surfaces parallel to S_1. All were plotted as S_1 surfaces.

Sixty-two strike and dip measurements of S_1 were recorded with a strike range of N 2° W to N 65° E, and an average of N 32° E. Seventy-five percent of these strike measurements fell within seven degrees of the average, suggesting some regularity to the regional strike. Stereonet plots of S_1-poles define a girdle which coincides with a great circle (Figure 4a). This pattern is frequently characteristic of a single episode of folding, or an episode of folding not later disturbed. The pole to the girdle is thought to define the fold axis with a bearing and plunge of S 32° W at 10 degrees. Similar plunge and strike data were obtained by Brobst (1962) on the country rocks in the Spruce Pine District approximately ten miles northeast of the Bank's Creek serpentinite.

Northwesterly dips range from 52 degrees to 84 degrees, and southeasterly dips range from 26 degrees to 80 degrees. Visual inspection suggested that two families of dip directions existed. This



A



B

Figure 4. Stereonet plots of S_1: A. - normals to S_1 showing girdle; B. -contour of normals to S_1.

idea was supported by contouring the S_1-poles, whereby two maxima formed approximately 120 degrees apart along the girdle pattern. Dips average 60° SE and 63° NW for these two families (Figure 4b).

Sixteen planar S_1-segments were constructed and their point

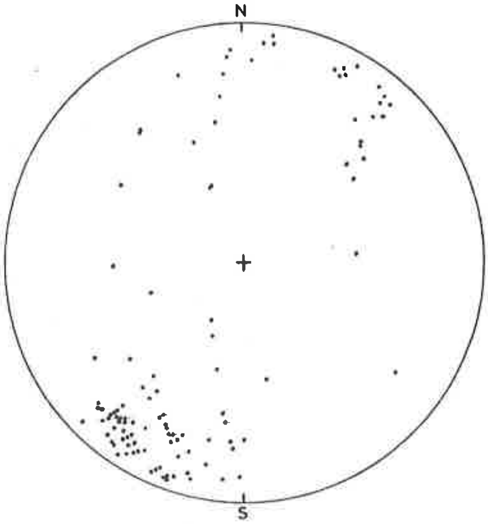


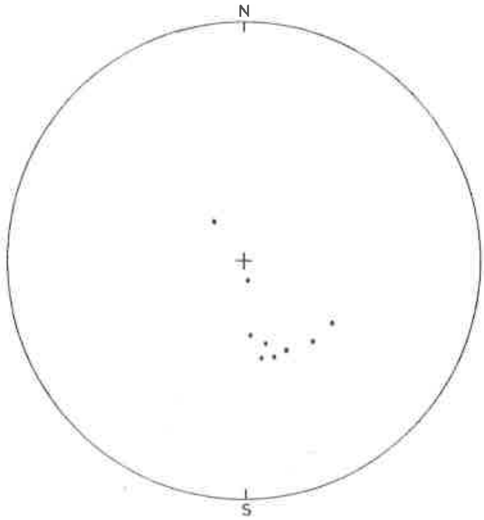
Figure 5. Stereonet plot of β -diagram.

intersections plotted (Figure 5). These intersection points, termed by Turner and Weiss (1963) as β, statistically define a fold axis. It was noted that a β -maximum coincided almost identically with the pole axis of the S_1 girdle (Figures 4a and 5).

Axes of the tight isoclinal folds are referred to as L_2. Nine measurements of L_2 bearing and plunges were recorded (Figure 6a). Stereonet plots of L_2 show a broad indistinct maximum of S 22° E at 48 degrees. Because of the very small number of data points and the scatter, it is believed that this point maximum is not statistically significant.

Superimposed on the S_1 (S_1 and S_2) surfaces of both the arenaceous and micaceous units is evidence of a second folding episode. This linear element is in the form of axes of open folds or crenulations with an amplitude of 1/16 inches to 4 inches and are referred to as L_3. Thirty-seven plunge and bearing measurements were recorded for L_3. Plots of L_3 are more highly concentrated forming a point maximum having a bearing and plunge of N 34° E at 4 degrees. Bearings of L_3 change from N 34° E to S 32° W due to the near horizontality of the plunge. Plunge in the case of L_3 varies only 15 to 20 degrees along a NE-SW directional trend (Figure 6b).

A close examination indicated that where S_1 dips to the SE, axial planes of the L_3 folds dip to the NW at about 55-60 degrees. Furthermore, where S_1 dips to the NW, axial planes dip to the SE, again at about 55-60 degrees. The sense of shear associated with L_3 folds also is related to the dip direction of S_1. Where S_1 dipped to the SE, the sense of shear was left-lateral; where S_1 dipped NW, the sense of shear



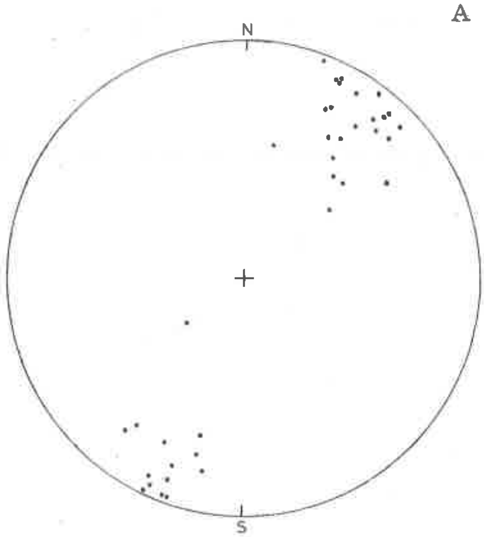


Figure 6. Stereonet plots of lineation elements: A. - L_2; B. - L_3.

was right-lateral. These data, combined with the bimodal concentrations of S_1-pole points in Figure 4b and L_3 point maximum in Figure 6b strongly suggest a synformal structure with the major fold axis striking approximately N 32° E. An idealized cross section is shown in Figure 7.

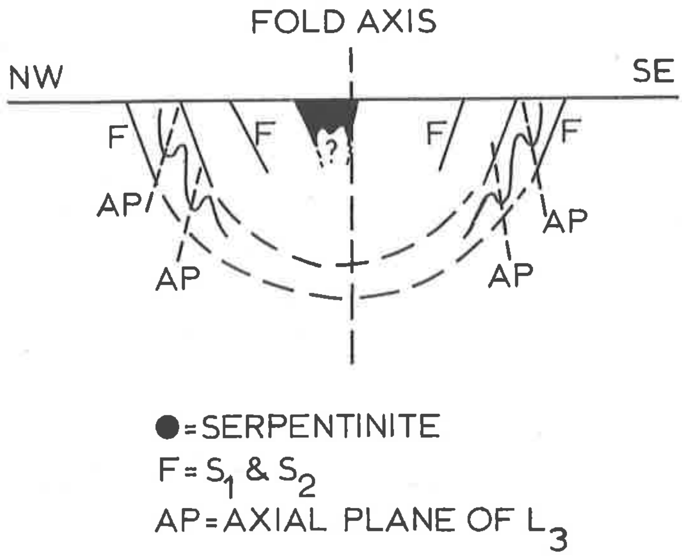


Figure 7. Diagrammatic cross-sectional sketch.

CONCLUSIONS AND SPECULATIONS

The most distinct physical characteristics exhibited by the Bank's Creek serpentinite that could lead to a genetic interpretation are the mesh texture and an apparent lack of shear or flow textures. The mesh texture is formed by retaining the original crystal grain outline by pseudomorphic replacement by serpentine. Criteria such as these, strongly imply an in situ serpentinization of some pre-existing olivine-rich rock.

Previous discussion concerning reactions involved in serpentinization raises questions about metasomatism, i. e., why is no MgO or SiO_2 metasomatism observed in the country rocks? Madison and Condie (1969) state that SiO_2 added to dunites during serpentinization may have come from silica-rich gneisses and schists surrounding the ultramafic complex; and that the MgO apparently "dribbled out" in solution into adjacent country rocks. Thayer (1966) believes that serpentinization is essentially a constant volume metasomatic process with removal of SiO_2 and MgO, and that the metasomatic diffusion is limited to a few inches or a few feet at the most. Coleman (1971) states that metamorphic serpentinites appear to have formed by volume-for-volume replacement with a loss of MgO or an addition of SiO_2; but no definite conclusion is drawn to explain the silica source or the secondary position of the magnesium because it is felt that serpentinization represents more than one period of reconstitution.

The structural data of this study suggest at least two episodes of deformation. One episode resulted in the deformation of compositional banding, producing isoclinal folds in appropriately competent units and axial plane foliations in less competent units. Evidence for a second deformational episode include the development of a more open folding which affected all units. The scatter in the orientation of isoclinal fold axes (L_2) in Figure 6a, mentioned earlier, can be explained by a later episode of folding. Therefore, it is concluded that the development of S_2 and L_2 were genetically related to an early deformational episode. The open folding that resulted in L_3 followed at a later time. The relationships between deformational episodes and fabric elements are given in Table 1. Evidence from the surrounding country rocks suggests a tightly folded synformal structure with a gentle southwesterly plunge of approximately ten degrees. According to this structural interpretation the Bank's Creek serpentinite lies near the trough axis on the lower-western flank, or almost upon that axis (Figure 8).

Table 1. Fabric Elements.

	Planar	Linear
	S_1 - compositional banding between meta-arenites and mica schists.	
I	S_2 - axial plane foliation in mica-schists.	L_2 - axes of tight isoclinal folds in meta-arenites; not seen in mica-schists.
II		L_3 - open folds on all planar surfaces, generally with horizontal axes.

I - folding episode
II - folding episode

It has been suggested that pressure gradients established by folding in the course of regional metamorphism could bring about chemical potential gradients in folded rocks (Carpenter, 1964, 1968). Chemical potential gradients in turn could lead to the migration of the more mobile constituents of the folded rocks, especially water, away from high pressure structural sites, or the fold flanks, to the low pressure sites, the fold crests or fold troughs. The higher concentration of water then brought about an *in situ* hydration of the minerals present. By this mechanism an olivine-rich rock which found itself in a fold crest or trough during regional metamorphism would quite likely

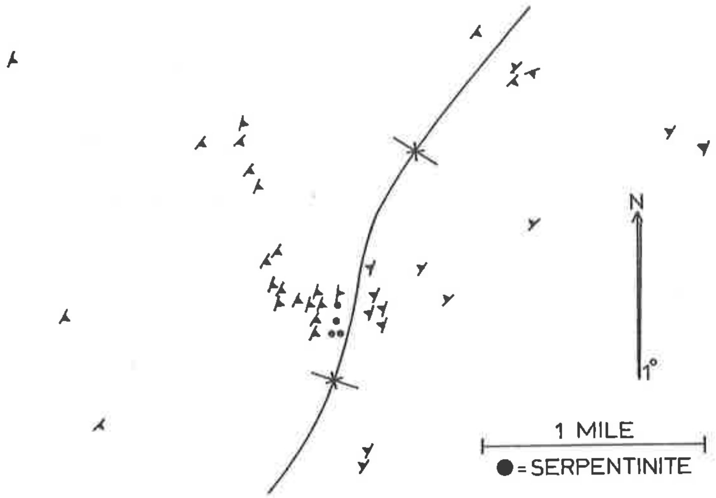


Figure 8. Foliation attitudes.

become hydrated to serpentine. If the structural interpretation by the present investigators is correct, one must hold open the possibility that the mineral assemblage is genetically related and controlled by its structural setting. An extension of the model by Carpenter and an in situ serpentinization might then be called upon as a favorable explanation for the Bank's Creek serpentinite. This model says nothing about the emplacement of the ultramafic body. However, previous studies have suggested that the ultramafics were emplaced prior to the first folding (and regional metamorphic episode) (Astwood, 1970).

In view of the structural and petrographic evidence presented, the in situ serpentinization process combined with the chemical potential hypothesis suggested by Carpenter, is preferred by the investigators to explain the Bank's Creek serpentinite.

REFERENCES

Astwood, P. M., 1970, A petrofabric study of two alpine-type ultramafic bodies: Masters thesis (unpub.), University of South Carolina, Columbia.

Brobst, D. A., 1962, Geology of the Spruce Pine district, Avery, Mitchell and Yancey Counties, North Carolina: U. S. Geol. Survey Bull. 1122-A, 26 p.

Cann, J. R., 1968, Geological process at mid-ocean ridge crests:

Geophys. Journ., v. 5, p. 331-341.

Carpenter, J. R., 1964, Influence of structural deformation on some aspects of metamorphic differentiation: Ph. D. thesis (unpub.), Florida State Univ., Tallahassee.

__________, 1968, Apparent retrograde metamorphism: Another example of the influence of structural deformation on metamorphic differentiation: Contr. Min. Pet., v. 17, p. 173-186.

Chase, R. L. and J. B. Hersey, 1968, Geology of the north slope of the Puerto-Rico Trench: Deep-Sea Res., v. 15, p. 297-318.

Coleman, R. G., 1971, Petrologic and geophysical nature of serpentinites: Bull. Geol. Soc. Am., v. 82, p. 897-918.

Cowan, D. S. and C. F. Mansfield, 1970, Serpentinite flows on Joaquin Ridge, Southern Coast Ranges, California: Bull. Geol. Soc. Am., v. 81, p. 2615-2628.

Dietz, R. S., 1963, Alpine serpentinites as oceanic rind fragments: Bull. Geol. Soc. Am., v. 74, p. 947-952.

Hess, H. H., 1960, Caribbean research project: Progress report, Bull. Geol. Soc. Am., v. 71, p. 235-240.

Hess, H. H., 1966, Caribbean geological investigation: Geol. Soc. Am. Memoir 98, 310 p.

Jahns, R. H., 1967, Serpentinites of the Roxbury District, Vermont, in Ultramafic and Related rocks (ed. Wyllie): John Wiley and Sons, Inc., p. 137-160.

Keith, A., 1905, Geologic atlas of the U. S., Description of Mt. Mitchell Quad. Folio, U. S. Geol. Survey, Washington, D. C.

Lapham, D. M., 1967, The tectonic history of multiply deformed serpentine, in Ultramafic and related rocks (ed. Wyllie): John Wiley & Sons, Inc., p. 174-183.

Le Pichon, X., 1969, Models and structures of the oceanic crust: Tectonophysics, v. 7, p. 385-402.

Madison, J. A., and K. C. Condie, 1969, Compositional and volume changes accompanying progressive serpentinization of dunites from the Webster-Addie ultramafic body, North Carolina: Amer. Min., v. 54, p. 1173-1178.

Moorehouse, W. W., 1959, The study of rocks in thin section: Harper & Row, 514 p.

Page, N. J., 1967, Serpentinization considered as a constant volume metasomatic process: A discussion: Amer. Min., v. 52, p. 545-549.

Rankin, D. W., 1969, Stratigraphy and structure of Precambrian rocks in northwestern North Carolina, in Studies of Appalachian Geology Central and Southern (ed. Rodgers): Wiley & Sons, Inc., p. 227-245.

Thayer, T. P., 1966, Serpentinization considered as a constant metasomatic process: Amer. Min., v. 51, p. 685-710.

__________, 1967, Serpentinization considered as a constant metasomatic process: A reply: Amer. Min., v. 52, p. 549-553.

Turner, F. J., and L. E. Weiss, 1963, Structural analysis of metamorphic tectonites: McGraw-Hill Book Company, Inc., 530 p.
Vogt, P. R., E. D. Schneider and G. L. Johnson, 1969, The earth's crust and upper mantle: Am. Geophys. Union Geophys. Monograph 13, p. 556-618.
Wyllie, P. J., (ed.), 1967, Ultramafic and related rocks: John Wiley & Sons, Inc., 464 p.

CHEMICAL WEATHERING OF THE BISCAYNE AQUIFER, DADE COUNTY, FLORIDA[1]

By

Donald D. Runnells
Department of Geological Sciences
University of Colorado
Boulder, Colorado 80302

ABSTRACT

The average annual rainfall along the coastal ridge in Dade County, Florida is 60 to 65 inches, of which about 20 inches is lost as subsurface drainage. The content of calcium dissolved in the rainwater is about 0.6 parts per million, whereas the shallow groundwater contains an average of 77.2 parts per million dissolved calcium. This difference in dissolved calcium is the result of the chemical weathering within the Biscayne aquifer. Based upon published radiometric dates and newly determined porosities, a theoretical total thickness of 24 feet of limestone could have been removed by chemical weathering since deposition and subaerial exposure of the Miami oolite about 130,000 years ago. The general lack of a karst topography in this area suggests that the actual lowering of the surface has probably been subordinate to the development of secondary porosity in the aquifer. Such porosity is well displayed immediately below the water table in the Miami oolite.

Determinations of the content of insolubles in the Miami oolite show that a thickness of nearly three feet of residual quartz sand could theoretically have been produced by the weathering and lowering of the surface of the Miami oolite.

INTRODUCTION

The discussion that follows is a summary of a study that was conducted during 1966. The results are interesting and point to several additional topics that deserve consideration. However, it is not possible for me to do further work on this subject. It is hoped that others will find the interest to test and pursue these preliminary results.

1/ EPR Publication 588, Shell Development Company, Houston, Texas

Acknowledgments

Robert J. Dunham first suggested this study. As is his way, he also asked the right questions during the course of the investigation. My thanks to him and to Shell Development Company for support and permission to publish. Ronald Perkins of Duke University and Conrad Neumann of the University of Miami offered extremely helpful and constructive criticism of the manuscript.

WATER AND ROCKS

Rainwater in the coastal areas of the eastern United States generally contains about 0.5 ppm total dissolved calcium (Junge and Werby, 1958, p. 421). In contrast, analyses of nine fresh groundwaters from carbonate strata of the eastern United States show from 15 to 124 ppm calcium, with an average value of 62 ppm (White, Hem, and Waring, 1963, p. F22). This difference in content of dissolved calcium indicates that the net effect of one liter of rainwater passing through a limestone in the eastern United States is to dissolve from about 0.0001 to 0.001 moles of the wallrock, corresponding to about 0.01 to 0.1 grams of calcium carbonate. The discussion that follows attempts to assess the significance of this type of chemical weathering in the Biscayne aquifer and, more specifically, in the Miami oolite of southeastern Florida.

The Biscayne aquifer is the most important shallow non-artesian reservoir of groundwater in southeastern Florida (Schroeder, Klein and Hoy, 1958). The approximate areal extent of the Biscayne aquifer is shown in Figure 1. It is a hydrologic unit that cuts across several different formations according to the properties of the rocks (Figure 2). The aquifer extends to an average depth of about 100 feet below the surface along the coast, and it includes all or part of a number of permeable sandstone and carbonate units ranging in age from upper Miocene through Pleistocene (Parker and others, 1955). As shown in Figure 2, the youngest carbonate formation in the Biscayne aquifer is the Miami oolite.

The Miami oolite is a pelletoid and oolitic packstone to grainstone of Pleistocene age (rock names based on classification by Dunham, 1962). It lies at or near the surface over most of Dade County, and is particularly well exposed along the coastal ridge which extends southwestward from the city of Miami. Along this ridge the Miami oolite lies as high as 20 feet above sea level and is either exposed at the surface or is covered only by a thin mantle of sandy soil. In the western two-thirds of Dade County the unit is covered by the peat and marl of the Everglades. The maximum thickness of the Miami oolite is about 40 feet (Parker and others, 1955, p. 102). Mineralogically it consists chiefly of calcite with a low content of magnesium, plus

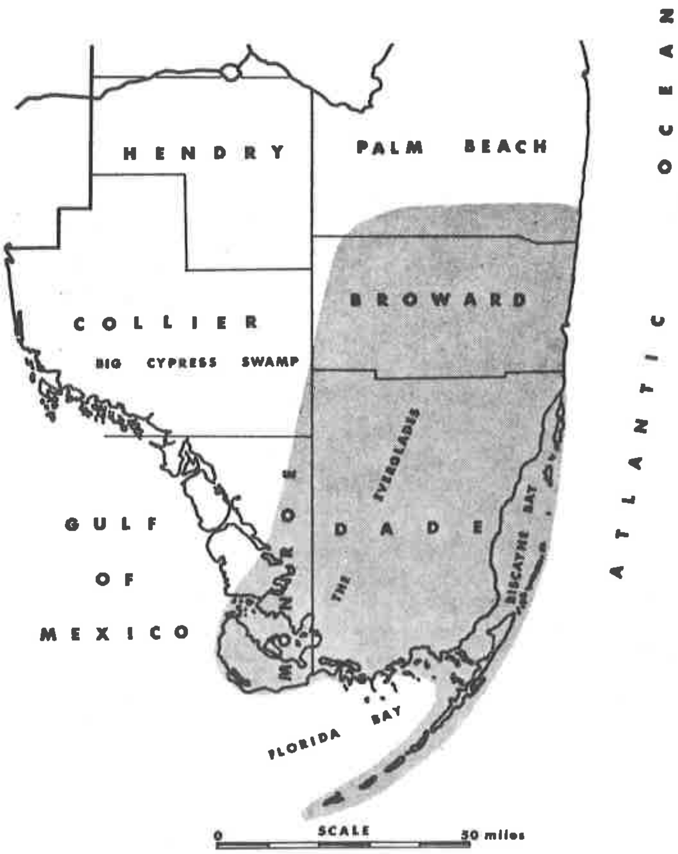


Figure 1. Approximate areal extent of Biscayne aquifer in southern Florida (modified after Schroeder, Klein, and Hoy, 1958).

some quartz sand and residual aragonite. Four samples of the rock, taken over a vertical interval of 7 feet in a trench, contain from 9.4 to 13.0 weight percent insoluble residue, with an average value of 11.4 percent. Most of the insoluble residue is subrounded and well-sorted fine to medium quartz sand.

The porosity of the Miami oolite is quite high. I have obtained approximate values for eight hand-specimens of the rock taken from above the present water table. The porosity was measured by saturating the hand-specimens with water under vacuum, measuring the weight loss upon drying at 110°C, and assuming a grain density of 2.71 for a mixture of 90 percent calcite and 10 percent quartz. The values range from 29.2 to 43.4 percent, with an average of 35.7 percent. The actual porosity is probably slightly greater, because some water was

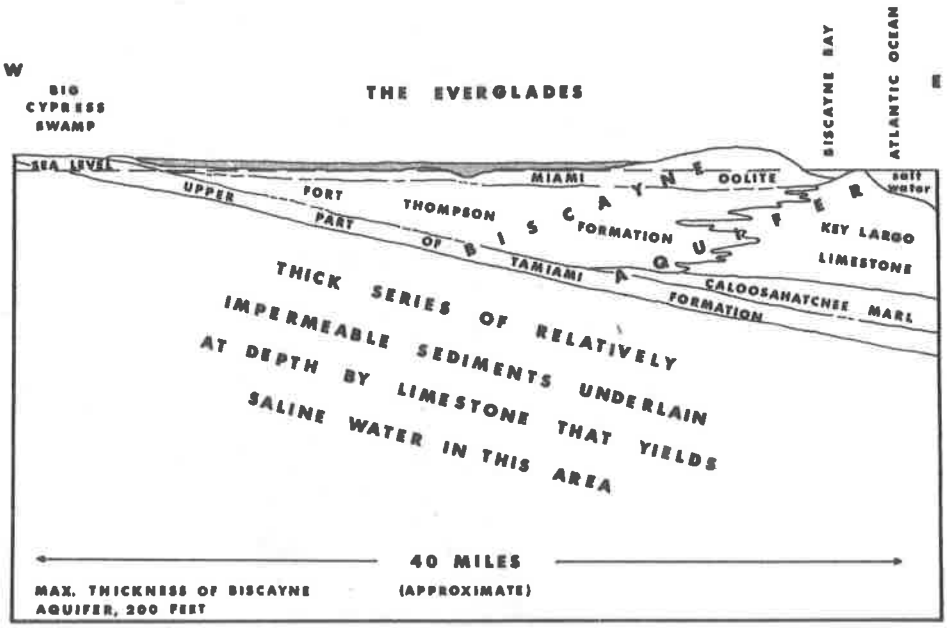


Figure 2. Generalized geologic cross-section showing the composite nature of the Biscayne aquifer (modified after Parker, 1951).

undoubtedly lost from the pores ot the rock during the measurements. An estimate of the in-situ porosity of the Miami oolite as a whole can be gained from certain parameters that have been measured by geologists during the course of studies of groundwater in the region. For example, data from Parker and others (1955, p. 219, 248, 268) indicate that the specific yield of the Miami oolite is about 15 to 20 percent (the specific yield is the ratio, expressed in percent, of the volume of water a material will yield by gravity drainage to the total volume of the material). However, the specific yield of a soil or rock is less than the porosity because not all of the water will drain by gravity. From studies in the area of Los Angeles, California, it was shown by Eckis (1934) that fine to coarse and gravelly sands, with specific yields ranging from about 15 to 30 percent, had porosities from about 35 to 43 percent; these values check fairly well with those mentioned above for hand specimens of the Miami oolite.

CHEMICAL WEATHERING

From available data on the amount and composition of rainfall and groundwater, it is possible to calculate the approximate annual rate of chemical weathering of the Biscayne aquifer and the Miami oolite in eastern Dade County; the requisite data are not available for the swampy Everglades to the west. The average annual rainfall along the coastal ridge near Miami is 60 to 65 inches (Thomas, 1970, Fig. 2). Of the total rainfall, about 40 to 45 inches is lost through evapotranspiration (Parker and others, 1955, p. 231). Surface runoff is insignificant in areas not traversed by drainage canals. Thus, along the coastal ridge only about 15 to 25 inches of the total annual rainfall is discharged to the sea as subsurface drainage. The content of calcium in the rain in this area is about 0.6 ppm (Junge and Werby, 1958, p. 421). In contrast, the average content of calcium in waters from 51 shallow wells in eastern Dade County, as listed by Parker and others (1955, p. 794-795), is 77.2 ppm, with a range from 60 to 102 ppm. The difference of 76.6 ppm dissolved calcium, multiplied by an estimated average 20 inches of groundwater runoff, indicates that along the coastal ridge the net leaching is equivalent to about 0.0036 cubic centimeters of calcite per year per square centimeter (density of calcite = 2.72 g/cc).

Osmond, Carpenter, and Windom (1965) have determined Th^{230}/U^{234} dates of 130,000 years for the Miami oolite. If certain simplifying assumptions are now made, it is possible to calculate the total volume of calcite removed from the Biscayne aquifer since deposition of the Miami oolite. Because the Miami oolite is the topmost unit of the Biscayne aquifer in the area of interest, it seems likely that most of the removal of the calcite has taken place from the Miami oolite itself.

The first assumption necessary for the calculation of the extent of chemical weathering is that the rocks of the upper part of the Biscayne aquifer were permanently exposed to subaerial conditions shortly after deposition of the marine Miami oolite. This assumption seems to be supported by curves of changes of sea level in this region, with initial exposure probably occurring 5000 to 10,000 years after deposition (Neumann, 1968; personal communication, 1971). Because of the flexibility in these dates the period of exposure is assumed to be 130,000 years. The second necessary assumption is that the amount and relative proportions of rainfall have remained essentially constant since leaching began. The weakness of this latter assumption is obvious.

Using the previous figures for the composition and amount of groundwater runoff, plus the two assumptions given above, a total thickness of approximately 15 feet of calcite could have been removed from the area of the coastal ridge since subaerial exposure was initiated. Correcting for the present average porosity of 35.7 percent which I measured in hand-specimens, the thickness of 15 feet of calcite corresponds to a thickness of approximately 24 feet of limestone. This removal of material could be reflected either as a lowering of the surface of the coastal ridge or as the production of additional porosity within the aquifer or as a combination of these two factors. Although sinkholes do occur, the fact that the oolite ridge as a whole generally does not exhibit a karst-type topography argues against extensive lowering

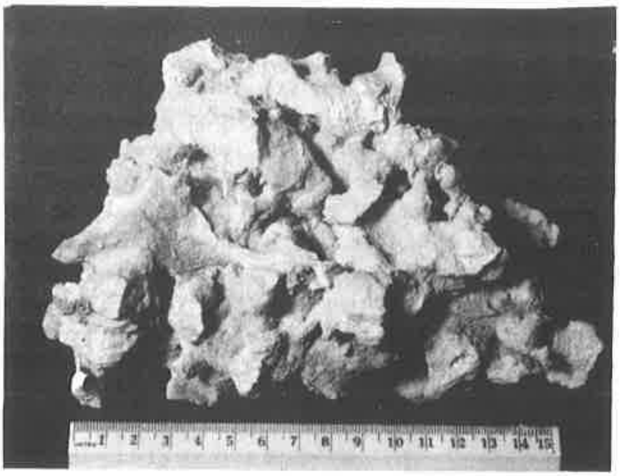


Figure 3. Specimen of Miami oolite from below the present water table in the Miami area, showing the tortuous nature of the secondary porosity. The finger-sized porosity is apparently related to the positions of paleo-burrows. (Specimen and interpretation supplied by Conrad Neumann).

of the surface. Similarly, the average porosity in hand-specimens of the Miami oolite from above the water table is slightly less than would be expected for an unconsolidated sand, thus implying that the production of additional porosity through leaching above the water table has not been an important net process. In contrast, it has been pointed out to me by Neumann (personal communication, 1971) that there is extensive secondary porosity just below the water table in the Miami oolite. In some samples the porosity is obviously greater than 50 percent. As shown in Figure 3, these large pore spaces have the shape of sinuous channels, with a sub-circular cross-section and a diameter about the same as that of a man's finger. The tortuous paths of these channels through the rock are apparently controlled by fossil burrow structures (Neumann, personal communication, 1971). These facts seem to suggest that most of the leaching of the Biscayne aquifer is taking place below the present water table, but this cannot be verified until better values become available for the in-situ porosity of the various rock units.

RESIDUAL SAND

Correcting for the slight difference between weight percent and volume percent calcite and quartz, and noting that the residual sand has a porosity about the same as the underlying bedrock, a thickness of about 2.8 feet of quartz sand could have theoretically been produced by the removal of 24 feet of Miami oolite. Depending upon the degree to which the surface has been lowered, this substantial figure indicates that some portion of the quartz sand that now forms a thin and discontinuous mantle over the bedrock in southern Florida may represent the insoluble residue left from the weathering of the Miami oolite.

CEMENTATION

As noted previously, approximately 40 to 45 inches of the annual rainfall is lost from the coastal ridge through evapotranspiration. Of this total amount, about 20 to 25 inches is removed from beneath the present water table (Parker and others, 1955, p. 231). That is, 20 to 25 inches of water are lost only after the water has percolated downward through the vadose zone and has become part of the groundwater reservoir. The removal of this large quantity of water, after it has had time to become at least partially saturated with calcium carbonate, could be important in the production of patchy calcite cement in both the vadose and shallow phreatic zones.

SUGGESTIONS FOR FURTHER STUDY

The conclusions reached in this reconnaissance study obviously involve a number of assumptions that could be seriously in error. However, the results do seem reasonable and point to at least three specific questions which deserve additional study. First, what is the magnitude and distribution of the present porosity of the various units in the Biscayne aquifer, and do these factors point to the relative importance of the production of additional porosity through weathering versus the lowering of the surface? Second, do the quantity and characteristics of the quartz sand now present on the surface in southern Florida suggest a substantial contribution from the insoluble residue in the Miami oolite? And third, are some of the patterns of patchy cementation that we observe in carbonate rocks related to evapotranspiration of partially saturated water from below the water table?

REFERENCES

Dunham, R. J., 1962, Classification of carbonate rocks according to depositional textures, in W. E. Ham (ed.), Classification of carbonate rocks: Amer. Assoc. Petrol. Geologists, Memoir 1, Tulsa, p. 108-121.

Eckis, R., 1934, South coastal basin investigation, geology and ground-

water storage capacity of valley fill: Calif. Div. Water Resources, Bull. 45, Sacramento, 279 pp.
Junge, C. E. and Werby, R. T., 1958, The concentration of chloride, sodium, potassium, calcium, and sulfate in rain water over the United States: Jour. of Meterology, v. 15, no. 5, p. 417-425.
Neumann, A. C., 1968, Elevated Pleistocene sea level features in the Bahamas: Fifth Caribbean Geological Conference, St. Thomas, V. I., (abstr.).
Osmond, J. K., Carpenter, J. R., and Windom, H. L., 1965, Th^{230}/U^{234} age of the Pleistocene corals and oolites of Florida: Jour. Geophys. Research, v. 70, no. 8, p. 1843-1847.
Parker, G. G., 1951, Geologic and hydrologic factors in the perennial yield of the Biscayne aquifer, Miami area, Florida: Amer. Water Works Assoc. Jour., v. 43, no. 10, p. 821.
Parker, G. C., Ferguson, G. E., and Love, S. K., and others, 1955, Water resources of southeastern Florida: U. S. Geol. Survey Water Supply Paper 1255, 965 pp.
Schroeder, M. C., Klein, Howard, and Hoy, N. D., 1958, Biscayne aquifer of Dade and Broward Counties, Florida: Fla. Geol. Survey Rept. Invest. 17, 56 pp.
Thomas, T. M., 1970, A detailed analysis of climatological and hydrological records of south Florida with reference to man's influence on ecosystem evolution: Univ. of Miami, Rosenstiel School of Marine and Atmospheric Science, Tech. Report 70-2, 89 pp. plus tables.
White, D. E., Hem, J. D., and Waring, G. A., 1963, Data of Geochemistry, 6th edition, Chapter F, Chemical composition of subsurface waters: U. S. Geol. Survey Prof. Paper 440-F, 67 pp.

SEDIMENTATION IN ST. LOUIS BAY, MISSISSIPPI

By

Charles M. Hoskin
Department of Geology
Box 5345, Southern Station
University of Southern Mississippi
Hattiesburg, Mississippi 39401

ABSTRACT

Sediments now accumulating in St. Louis Bay are detrital silicates with trace amounts of particulate organic matter and shell fragments. Twenty-seven percent of the Bay surface sediments are sands, the remainder being sandy silt-clay mixtures. Average grainsize composition, by weight, for 62 Bay sediments is a trace of gravel, 34 percent sand, 30 percent silt and 36 percent clay.

Sand is supplied by westward longshore drift from the man-made north shore beach of Mississippi Sound. There may be two sources of sand as evidenced by the common occurrence of angular and well-rounded quartz grains in the same size class. Sand accumulates near the Bay mouth, in places only inches thick, overlying the continental Citronelle Formation. Sand modes are coarsest at the Bay mouth (0.28 mm, 1.6 Φ) and become finer to the northwest Bay (0.09 mm, 3.5 Φ), suggestive of selective size transport from the south.

Silt and clay are supplied by the Wolf and Jourdan Rivers and on the flood tide from Mississippi Sound. Salinity of the Bay water, 7 to 8 °/oo is sufficient to cause mud flocculation. Silt size modes are coarsest in the northwest Bay (22.5 microns, 5.50 Φ) and become finer southward toward the Bay mouth (11.2 microns, 6.5 Φ). Mud is now accumulating near the river mouths and in the northwest Bay as interpreted from high water content of mud (74 percent) in those places in contrast to lower water content of surface mud at the Bay mouth (53 percent).

INTRODUCTION

Major objectives of this study were to (1) determine the grain-size characteristics and spatial distribution of sediment in the Bay, (2) identify important sources of sediment, and (3) identify sites of significant sediment accumulation.

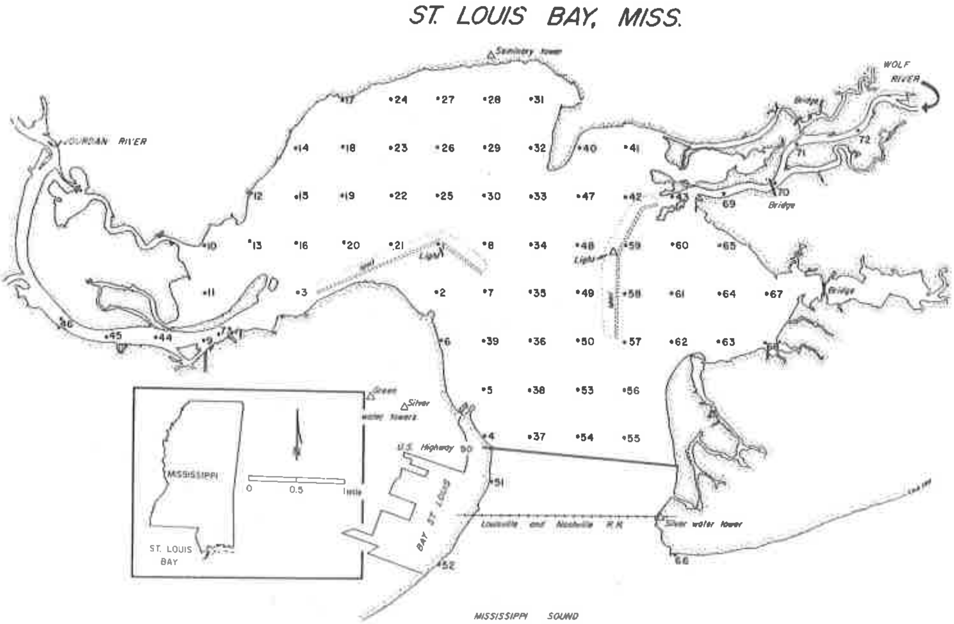


Figure 1. Index and sample station map for St. Louis Bay, Mississippi. Base from U. S. Department of Commerce 7.5 minute Quadrangle maps for Bay St. Louis 1956; Pass Cristian 1955; Vidalia 1956; and Waveland 1956.

Environmental Setting

St. Louis Bay is shaped like an inverted "L" and measures about 3 miles along each leg of the "L" (Figure 1). No contoured bathymetric chart is available; soundings on C & GS charts 1268 and 876-SC and fieldwork by the author suggest the average Bay depth to be about 6 feet. The eye of Hurricane CAMILLE, the most powerful hurricane known for the United States, crossed the Mississippi coast and traversed St. Louis Bay to the northwest on August 17, 1969. Winds up to 200 mph and 22-foot storm surge probably produced significant, but as yet uncharted changes in the bathymetry (U. S. Corps of Engineers 1970). Prior to CAMILLE, the deepest Bay water was a 10-foot channel at the Bay-mouth center, and with the exception of the 7-foot dredged channels shown in Figure 1, the Bay shoaled gradually to the east, north, and west. By observation in the field, correlation of increased turbidity with wind-driven waves nearly every summer afternoon suggests that the entire Bay floor is subject to disturbance by waves.

Mean annual precipitation is 58 inches for the western Bay, and this is time-distributed with the monthly mean being 7.3 inches for July and 2.3 inches for October (U. S. Dept. Interior 1970a). No discharge data are available for the Jourdan River, but the average minimum flow for the Wolf River near Landon is 18 mgd (Newcome et al., 1968). Seasonably variable rainfall probably has a strong influence on Bay water salinity as data for June 1971 were 7.29 to 8.29 $^{o}/oo$ (this paper) whereas salinity for August 1954 was 19.16 to 21.38 $^{o}/oo$ (Priddy et al., 1955, Table 6). Some of this variation may be caused by seasonal evaporation as the average water temperatures in the adjacent Mississippi Sound are 64°F for the winter and 83°F for summer (U. S. Dept. Interior, 1970b). No synoptic salinity data are known to be available for the Bay.

Winds blow from the south in summer at an average velocity of 8 mph and blow from the northeast at about 17 mph in October (U. S. Dept. Interior, 1970b). Wave heights in adjacent Mississippi Sound exceed five feet more than 20 percent of the time in winter but only five percent of the time in summer (U. S. Dept. Interior, 1970b). The tide is diurnal with a range of 1.6 feet (Jones, 1970). Water circulation in time and pattern for St. Louis Bay is not known. Surface currents in adjacent Mississippi Sound flow to the west at a mean velocity of 0.7 knots in summer and 1.1 knots in winter (U. S. Dept. Interior, 1970b). Judging by sand pile-up at groins and bar orientation, longshore drift is to the west near the mouth of St. Louis Bay.

Soil and vegetation maps are given by the U. S. Department of the Interior (1970a) for the western Bay shores. The surrounding land is conifer forest and a tidal marsh is building out into the Bay from the north. Population density is 10.0 to 14.9 persons mi^2 for the north shore, and 25.0 to 99.9 persons mi^2 for the south shore; the southern increase is due to Bay St. Louis townsite. Shellfish harvesting is man's chief use of the Bay (U. S. Dept. Interior, 1970a).

Acknowledgments

The work on which this report is based was supported in part by funds provided by the United States Department of the Interior as authorized under the Water Resources Research Act of 1964, Public Law 88-379, and administered through the Water Resources Institute for Mississippi. Laboratory facilities were provided by the University of Southern Mississippi, Department of Geology.

The author thanks C. Dana Hoskin and John McCarty for field assistance, Bill Marchant and John McCarty for lab work; and Mrs. A. Alexander of Bay St. Louis for generous logistic support.

SEDIMENTS

Previous Work

Local bedrock is the red and yellow gravel and sand river terrace deposits of the ?Pliocene - ?Pleistocene Citronelle Formation.

These deposits crop out along the 50-foot ridges about 1.5 miles north of the Bay (Fisk, 1944); the Citronelle Formation dips to the south under the Bay and spoil banks from dredged channels along the south Bay shore commonly contain these red and yellow materials.

Except for Priddy et al., (1955), earlier studies of Recent sediments for southwest Mississippi provide no data for St. Louis Bay (Brooks, 1962; Lynch, 1954; Shepard et al., 1960; and Upshaw, et al., 1966).

Field Methods

Sediment was collected by Phleger corer; each core was immediately extruded, cut into 3-inch segments unless grainsize changes were present in a given interval, and stored wet, unpreserved in polyethylene bags. Seventy-two stations were occupied on a one-half mile grid; navigation was done by resection with sextant-measured angles between prominent shore features (Figure 1). Water was collected with a one liter Van Dorn sampler and was stored unpreserved in glass bottles.

Laboratory Analysis

Grainsize analyses were made using screens and pipets following the methods of Folk (1968). Particulate organic matter and fecal pellets were destroyed with H_2O_2 prior to size analysis. Detailed analyses for the entire length of each core are not given; data presented here represent only the increment between the water-sediment interface and three inches down. The only notable changes with depth for these cores (maximum length two feet, average length for 55 cores was 1.5 feet) is that the surface material is oxidized (brown) except for stations 9 and 37 which were black throughout; below the surface interval sediments are gray; compaction tends to increase with depth; and cores in the river were mostly sand and contain mud only near the surface.

Salinity was determined by titration with $AgNO_3$ using IAPO sea water as a standard (U. S. Hydrographic Office, 1959). Suspended sediment concentration was determined by Millepore filtering of a known volume of water with before-and-after weighing of the 0.45 micron filter.

Grainsize Distribution

Particle size-distributions show that, by weight, the average composition of 62 Bay surface sediments is a trace of gravel (> 2.00 mm) as one sample in 62 contained gravel; 34 percent sand (2.00 to 0.0625 mm); 30 percent silt (0.0625 to 0.0039 mm); and 36 percent clay (0.0039 to 0.00006 mm). Sediment from the Bay mouth beaches was all sand and river channel sediment averaged 82 percent sand,

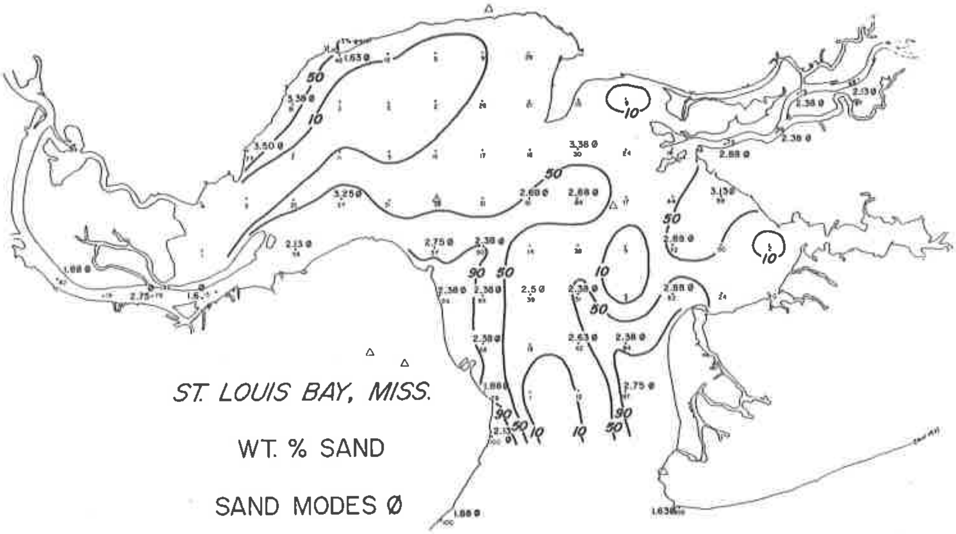


Figure 2. Sand abundance map. Smallest numbers represent weight percent sand in the top three inches of sediment at the station; intermediate-size numbers represent the sand modes expressed in Phi; and largest numbers are contour-line labels. Data for rivers not contoured.

except for the deep places that contain sandy mud. Histograms of number of samples plotted against weight percent showed near-symmetrical distributions about the mean for the silt and clay size-classes, but the sand distribution was strongly skewed with the most frequently-occurring increment containing between 0. and 9.9 percent sand.

Spatial distribution of sand, silt and clay in the Bay is given in map format in Figures 2, 3, and 4, respectively. Sand is most abundant along the Bay shores at its mouth, in a tongue extending two miles to the northeast and near the mouth of the Jourdan River. Sand is least abundant in the center of the Bay mouth and in a southwest-northeast-trending lobe in the northwest Bay. Silt was nearly ubiquitous at about 30 percent by weight except in the northwest Bay where it exceeds 50 percent, and along the northeast-trending sand tongue and Bay shore mouth where silt forms less than 10 percent of the Bay sediment. Similarly, clay occurs in low abundance in the northeast-trending sand tongue, and is most abundant in the northwest Bay, and with silt, is very abundant in the center of the Bay mouth. Several areas of clay abundance (> 50 percent) occur in the eastern Bay.

Sediment Sources

Bay sediments are virtually 100 percent detrital silicates; biogenic sediment as wood, leaf and marsh grass fragments and carbonate

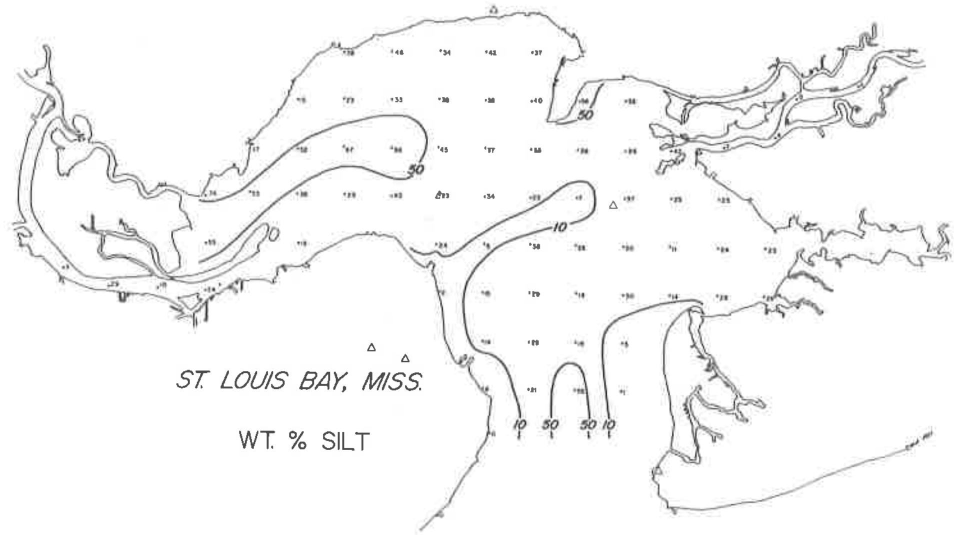


Figure 3. Silt abundance map. Smallest numbers represent weight percent silt at that station. Data for rivers not contoured.

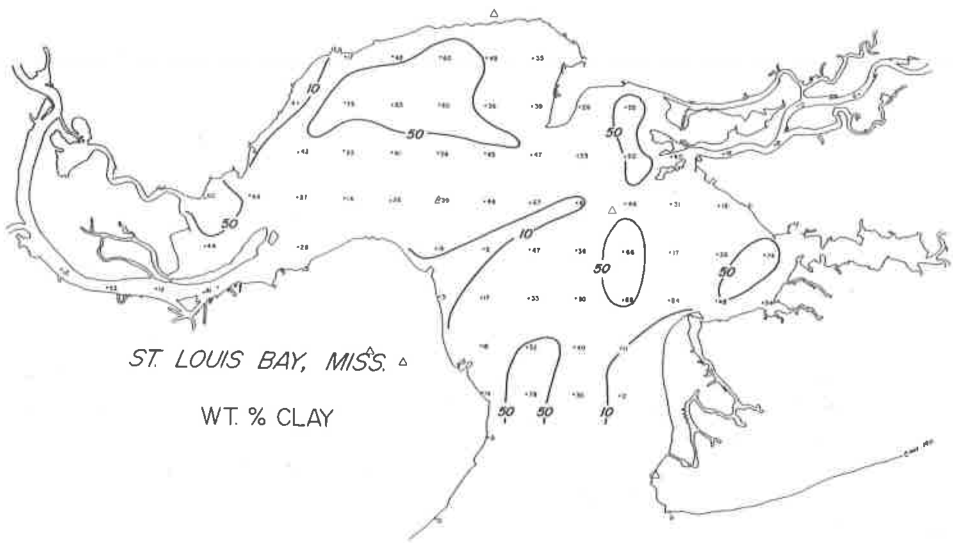


Figure 4. Clay abundance map. Smallest numbers represent weight percent clay at that station. Data for rivers not contoured.

shells were found in small-to-trace amounts, and no chemical sediment was found.

Sand is most probably supplied to the Bay by westward longshore drift of sand on the man-made beach of the north shore of Mississippi

Sound. Sand drift is probably not volumetrically significant as the beach was built in 1951 (Upshaw et al., 1966), endured many storms without major sand loss and has not required replenishing since its emplacement (personal communication, H. E. Walker, U. S. Corps of Engineers, 1971). The borrow channel from which the sand was dredged can be clearly seen in U. S. Department of Agriculture photomosaics (1958).

Sand in the northwest Bay is interpreted to have been entrained from sand beaches in the mouth area and transported in suspension during the building phase of the storm surge of Hurricane CAMILLE. Regretably, the pre-CAMILLE sediment distribution is not known. Widespread distribution in the Bay of ?asphalt from roads or ?roof-material from houses plus the known track of CAMILLE to the northwest support this interpretation. Four out of five samples from both the Wolf and Jourdan Rivers do not contain the asphaltic material.

Sand is the dominant sediment in the Wolf and Jourdan River channels near, and for several miles upstream of their juncture with the Bay. Silt and clay become increasingly more abundant in the river channels going downstream toward the Bay, and for this reason, it is believed that the rivers do not now supply sand to the Bay, except possibly during river flood stages.

Evidence for at least two sources of sand comes from the observation that angular and well-rounded quartz sand grains are found together in the same size-class in the coarse fraction of most sand-rich Bay sediments. As the rounding process is believed to be physical abrasion, all grains of the same size should have equal roundness if they are abraded at the same time. A mixture of angular and well-rounded grains of the same size therefore indicates one source for the angular grains and another source for the well-rounded grains. Presently, the location and time of mixing of the two sources is not known.

It has been suggested that a given source of sediment particles could be identified by grainsize modes - the peak on the size-frequency distribution or the steepest segment of the cumulative curve (Folk and Ward, 1957). Curray (1961) used this idea to trace sediments on the continental shelf west of the Mississippi delta. Grainsize modes for 32 sands are plotted in Figure 2; and the frequency distribution of these modes (Figure 5) is unimodal with the most frequently-occurring sand mode at about 0.2 mm (2.3Φ). Inspection of these data affords no evidence for two sand sources. Instead, there is an apparent decrease in sand mode size from 0.28 mm (1.6Φ) to 0.09 mm (3.5Φ) in a south-to-north direction (Figure 2); one exception being sediment from station 17, which contains three percent gravel (one pebble) and has a sand mode of 0.28 mm (1.6Φ). This suggests that if the major supply of sand is the sand beaches of the north shore of Mississippi Sound, then selective size-sorting occurred during transport, and that the idea of size mode identifying source does not apply in this place.

Potential sources of fine-grained sediment are the Wolf and

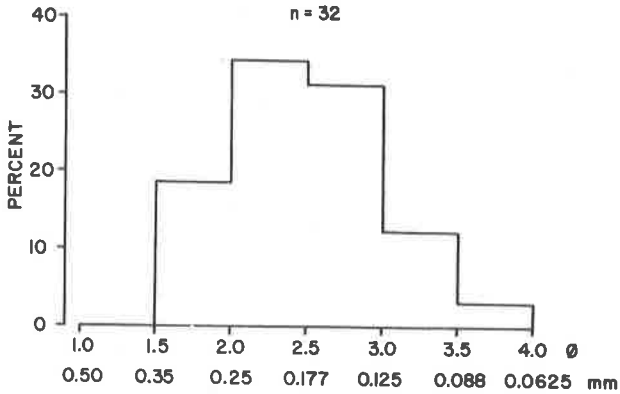


Figure 5. Histogram, showing size-frequency distribution of sand modes for 32 samples containing at least 15 grams of sand in the top three inches of core.

Jourdan Rivers and Mississippi Sound. Mud (silt plus clay) is transported as suspended material by the Wolf and Jourdan Rivers to St. Louis Bay. Data for amounts of suspended sediment at various stages of the Wolf River are presented by Newcome et al., (1968, Table 8, Fig. 9). No published data are available for the Jourdan River. Measurements of total suspended sediment concentration in the Wolf ($\overline{X}$ 6.8 mg/L mid-channel) and Jourdan ($\overline{X}$ 10.1 mg/L mid-channel) Rivers and the Bay were made on June 2-3, 1971; particulate organic matter and detrital silicate particles were not differentiated. To quantitatively evaluate mud supply, such measurements should be made on a repeating basis for at least one year to document river- and tide-stage effects, but that is clearly beyond the scope of this project. Data in this report (Millepore filtering) appear low compared to Newcome et al., 1968- evaporation method) perhaps due to the different methods used.

Although the data are sparse, one tentative interpretation is that bottom sediment in the Bay is resuspended by waves. Suspended sediment increased from station 53 ($\overline{X}$ 10.0 mg/L, Bay mouth) to station 30 ($\overline{X}$ 24.5 mg/L), two miles downwind and in mid-Bay. Data for stations 16 ($\overline{X}$ 14.1 mg/L) and 18 ($\overline{X}$ 11.1 mg/L), although an additional 1.5 miles further downwind from station 53, were obtained the following morning under near-calm wind conditions. Wind-driven waves erode mud, particularly from northern Bay shores and previously-accumulated mud is recycled almost daily. Three groups of four stations each were chosen to represent mud-rich areas in the Bay; Bay mouth, stations 37, 38, 53 and 54; mid-Bay, stations 8, 30, 33 and 34; and northwest Bay, stations 18, 19, 22, and 23. Size modes for the silt size-class were determined by pipet analyses made at one-half Phi intervals. Although complicated by overlapping data, results suggest a decrease in the average silt size mode from the northwest Bay (22.5 microns, 5.50 Φ) to mid-Bay (19 microns, 5.89 Φ) to Bay mouth (11.2 microns, 6.50 Φ). Silt grains appear to be larger to the north, and sand grains appear to be larger to the south; does this indicate separate sources for silt and sand? Unfortunately, these data are too few to permit a meaningful interpretation.

Salinity of the Bay water 7 to 8 ‰) and of mid-depth and bottom water in the Wolf River (to about six miles upstream, Newcome et al., 1968, Plate 2) and Jourdan River (this report) is sufficient to cause mud flocculation which may result in subsequent gravity-settling due to increased "particle" size.

Sediment Accumulation

Sand accumulation appears to be confined to the near-shore areas of the Bay mouth. Sediment thickness is generally not known, but at station 62 the coring tube penetrated two inches of sand and bottomed in compact red and yellow sandy clay, presumably the Citronelle Formation. This suggests that sediment thickness near the Bay mouth is not great, perhaps due to wave and tidal scour.

Originally suspended silt and clay which accumulates by gravity-settling usually traps a large amount of water. Compaction after settling squeezes out the water. This should permit identification of recently-accumulated sediment, as opposed to older sediment, by water content. Cores were taken for this purpose and water content determined for Bay muds from stations 53- Bay mouth ($\overline{X}$ 52.5 % H_2O by wt.), 30- upper Bay ($\overline{X}$ 57.0 % H_2O), and 18- northwest Bay ($\overline{X}$ 73.7 % H_2O). These data show increasing water content from Bay mouth to northwest Bay, suggesting that the northwest Bay is a site of mud accumulation. Water content for cores from the river mouths is also high ($\overline{X}$ Jourdan River 75.5% H_2O; Wolf River 64.7% H_2O), suggesting that mud is now accumulating there. Sampling sites of Priddy et al., (1955) are not close enough to those of this study for confident comparison of similar data (their Table 6).

CONCLUSIONS

1. Sediment now accumulating in St. Louis Bay is mostly of detrital silicate origin.

2. Grainsize analyses show 27 percent of the Bay sediments are sand, and 73 percent are mud. Average grainsize composition for 62 Bay sediments is 34 percent sand, 30 percent silt and 36 percent clay.

3. The major sand source is probably the north shore beach of Mississippi Sound. Sand accumulates mostly near the Bay-mouth shores. Sand size modes decrease from south-to-north, interpreted to

be the direction of sand transport. Sand is not transported by the Wolf and Jourdan Rivers into St. Louis Bay in non-flood conditions. Small amounts of sand in the northwest Bay were probably transported by Hurricane CAMILLE.

4. The Wolf and Jourdan Rivers and the flood tide from Mississippi Sound transport mud in suspension into St. Louis Bay. Salinity of the Bay water is sufficient to cause mud flocculation. Mud accumulates near the river mouths and in the northwest Bay. Size modes for silt appear to decrease from north-to-south in the Bay.

REFERENCES CITED

Brooks, F. L., 1962, A study of the substratum of a portion of the continental shelf in the northeastern Gulf of Mexico: Unpublished MS thesis in Geology, Mississippi State University, 65 p.

Curray, J. R., 1961, Tracing sediment masses by grain size modes: 21st International Geol. Congress, Copenhagen, Denmark, Reports, v. 23, p. 119-130.

Fisk, H. N., 1944, Geologic investigation of the alluvial valley of the Lower Mississippi River: Mississippi River Commission, Vicksburg, Mississippi, 78 p.

Folk, R. L., 1968, Petrology of sedimentary rocks: Hemphill's Book Store, Austin, Texas, 170 p.

Folk, R. L., and W. C. Ward, 1957, Brazos River bar: a study in the significance of grain size parameters: Jour. Sedimentary Petrology, v. 27, p. 3-26.

Jones, D. A., 1970, Tide tables, high and low water predictions 1971, East Coast of North and South America: U. S. Dept. of Commerce, ESSA, Coast and Geodetic Survey, 290 p.

Lynch, S. A., 1954, Geology of the Gulf of Mexico, p. 67-86, In: Galtsoff, P. S. (editor), Gulf of Mexico, its origin, waters and marine life: Fishery Bull. of the Fish and Wildlife Service, v. 55, 604 p.

Newcome, Roy Jr., D. E. Shattles, and C. P. Humphreys, Jr., 1968, Water for the growing needs of Harrison County, Mississippi: U. S. Geological Survey Water-Supply Paper 1856, 106 p.

Priddy, R. R., R. M. Crisler, Jr., C. P. Sebren, J. D. Powell, and Hugh Burford, 1955, Sediments of Mississippi Sound and inshore waters: Miss. State Geological Survey, Bull. 82, 54 p.

Shepard, F. P., F. B. Phleger, and T. H. Van Andel, 1960, Recent sediments, northwest Gulf of Mexico: American Assoc. Petroleum Geologists, Tulsa, Oklahoma, 394 p.

United States Corps of Engineers, 1970, Hurricane CAMILLE, 14-22 August 1969: U. S. Army Engineer District, Mobile, Alabama, 80 p.

United States Dept. of Agriculture, 1958, Aerial photo-mosaic of

Hancock County, Mississippi, County symbol CZI: Commodity Stabilization Service, The Aeroflex Corporation, Robinson Aerial Surveys, Div., Newark, N. J.

United States Dept. of Commerce, 1965, Lake Borgne and approaches: Coast and Geodetic Survey Chart 1268, 8th edition, scale 1:80,000.

United States Dept. of Commerce, 1970, Dog Keys Pass to Waveland, Mississippi-Louisiana, Nautical Chart 876-SC, Intracoastal Waterway: National Oceanic and Atmospheric Administration, 6th edition, scale 1:40,000.

United States Dept. of the Interior, 1970a, Environmental conditions and resources of southwestern Mississippi: U. S. Geological Survey, Geographic Applications Program, 58 p.

United States Dept. of the Interior, 1970b, The national atlas of the United States of America: U. S. Geological Survey, Washington, D. C., 417 p.

United States Hydrographic Office, 1959, Instruction manual for oceanographic observations: U. S. Hydrographic Office Publication, 607 p.

Upshaw, C. F., W. B. Creath, and F. L. Brooks, 1966, Sediments and microfauna off the coasts of Mississippi and adjacent states: Miss. Geological, Economic and Topographical Surv., Bull. 106, 127 p.

HEAVY MINERALS OF NORTHERN SAND KEY

PINELLAS COUNTY, FLORIDA

By

William C. Hood
Sandra D. Hood
Department of Geology
Southern Illinois University
Carbondale, Illinois

and

S. Melodie Oleson
Southwest Florida Water Management District
Post Office Box 457
Brooksville, Florida

ABSTRACT

Beach sands of Sand Key, Florida, contain a suite of fourteen heavy minerals, concentrations of which reach thirty percent in a half-mile long area near the northern end of the beach. Ilmenite, staurolite, garnet and zircon are the most common detrital heavy minerals along most of the beach with zircon reaching nearly ten percent in the zone of concentration. Weight percentage data indicate heavy mineral abundances in the zone of concentration are mainly a function of specific gravity with some influence shown by grain shape. Large concentrations of heavy minerals are associated with small grain diameters of quartz. It is postulated that wave erosion has selectively removed quartz, especially the larger grain sizes, and left a residue of the more dense heavy minerals and smaller grain sizes of quartz.

INTRODUCTION

Sand Key, a barrier island on the west-central coast of Florida, has a substantial concentration of heavy minerals along the northern portion of its western beach. This paper reports the results of an investigation to determine the identity and quantity of the heavy minerals along a portion of the Sand Key beach and attempts to explain the cause of the variation in concentration.

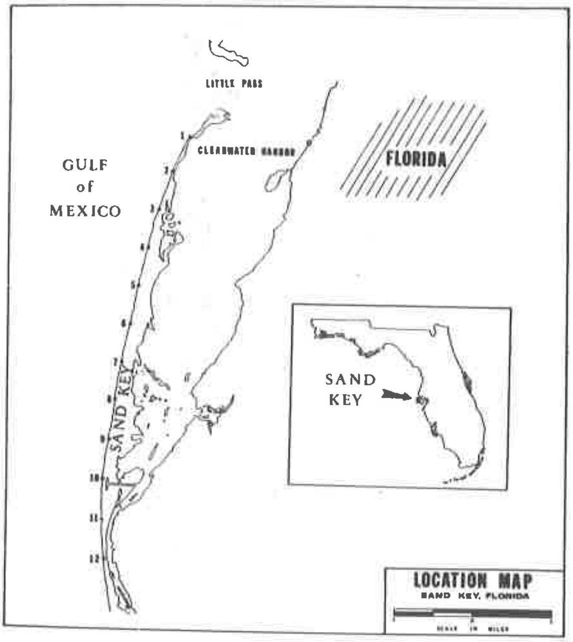


Figure 1. Sample location map, Sand Key, Florida.

Acknowledgments

The authors would like to express their appreciation to John Utgaard and Frank Ethridge, both of whom read the original manuscript and offered many helpful suggestions.

METHODS

Twelve pairs of samples, each consisting of several hundred grams of sand from the uppermost two inches of the beach, were collected at one-half mile intervals (Figure 1) along Sand Key beach northward from the town of Indian Rocks Beach to Little Pass at the north end of the island. In an attempt to obtain the samples from a similar setting, each was collected just above the demarcation of the previous high tide. Two samples were taken from each location to serve as an internal check on heavy mineral concentrations.

A thirty gram representative split of each of the duplicate samples was subjected to bromoform separation to isolate the heavy minerals. After separation, both the heavy and light mineral fractions were thoroughly washed with acetone, dried, weighed, and the weight percentage of heavy minerals calculated from each split. One set of heavies was used in identification studies; the other set of heavies and the associated lights were used in the grain size studies.

To identify and determine the frequency of the heavy minerals a split of heavy minerals from each of the twelve collecting stations was mounted in Canada balsam. Identification of minerals present was by optical techniques supplemented where necessary by mass magnetic susceptibility measurements. Frequency of occurrence was obtained by counting three hundred grains on each slide. During the course of the investigation it was observed that the large quantity of aragonitic shell fragments appeared to reduce the reliability of the frequency figures for the other minerals. Therefore, a separate portion of each heavy mineral sample was acidified to remove the aragonite, after which the insoluble residue was mounted and identified as before. This procedure gave much more meaningful data on the relative frequency of the less common heavy minerals.

One set of heavy and light minerals from each of the sampling stations was sieved to determine grain size distribution, using screens with openings of 2.00, 1.00, 0.500, 0.250, 0.125, and 0.062 mm. Each size fraction was weighed on a Mettler balance and statistical parameters were calculated from the weights thus obtained. The statistical parameters used are those described by Folk (1968): median, graphic mean, inclusive graphic standard deviation, inclusive graphic skewness, and graphic kurtosis.

Examination of the results of the grain size analyses (Table 1) shows that the sediment of northern Sand Key is almost entirely between 0.50 and 0.125 mm in grain diameter, with most of it being between 0.250 and 0.125 mm in both the light and heavy sand fractions. The average mean diameter of the lights is 1.05Φ units, with the heavies averaging 0.07Φ units higher at 1.12Φ (Table 2).

The light fraction is well sorted at all localities sampled; inclusive graphic standard deviations range from 0.32 to 0.58Φ with the arithmetic mean for all collecting localities having a value of 0.44Φ . The heavy mineral fraction is moderately well sorted, with the range in standard deviations being from 0.33 to 0.76Φ . The average is 0.59Φ , which is significantly higher than that of the light fraction. This is to be expected, because the light fraction is almost all quartz with minor calcite, so that in the light fraction the specific gravity of each grain is nearly identical. However, in the heavy mineral fraction, up to fourteen minerals, having a wide range in specific gravity and hence a wide range in equivalent hydraulic diameters, are present.

The inclusive graphic skewness for the light minerals ranges from 0.03 to -0.39 with an average value of -0.16 for the twelve samples. The heavies range in skewness from 0.03 to -0.46, slightly less than the lights, but the average value, -0.14, is nearly identical. Negative skewness values mean coarser particles predominate over finer particles.

Kurtosis values for the light fraction range from 0.97 to 1.76,

Table 2. Sediment Descriptive Parameters, in Φ Units.

Sample	Median	Graphic Mean	Standard Deviation	Graphic Skewness	Graphic Kurtosis
1 (Light)	1.05	1.01	0.50	-0.17	1.04
2 (Light)	0.89	0.92	0.48	0.03	0.97
3 (Light)	1.29	1.27	0.39	-0.13	1.20
4 (Light)	1.20	1.17	0.47	-0.10	1.06
5 (Light)	1.17	1.13	0.40	-0.17	1.12
6 (Light)	1.03	0.91	0.48	-0.27	1.02
7 (Light)	1.01	0.97	0.43	-0.19	1.05
8 (Light)	1.05	1.02	0.40	-0.13	0.97
9 (Light)	1.23	1.14	0.58	-0.39	1.76
10 (Light)	0.95	0.92	0.41	-0.09	1.00
11 (Light)	1.39	1.15	0.36	-0.14	1.14
12 (Light)	1.08	1.03	0.32	-0.22	0.99
1 (Heavy)	0.87	0.84	0.66	0.03	1.12
2 (Heavy)	1.00	0.94	0.64	-0.17	1.05
3 (Heavy)	1.61	1.61	0.51	-0.01	1.09
4 (Heavy)	1.40	1.38	0.55	-0.08	0.98
5 (Heavy)	1.38	1.35	0.55	-0.10	1.09
6 (Heavy)	0.99	0.90	0.74	-0.22	1.06
7 (Heavy)	1.29	1.09	0.72	-0.46	1.64
8 (Heavy)	1.10	1.06	0.53	-0.10	1.67
9 (Heavy)	1.40	1.32	0.76	-0.27	1.66
10 (Heavy)	1.05	0.67	0.53	-0.13	1.01
11 (Heavy)	1.25	1.21	0.60	-0.12	1.03
12 (Heavy)	1.08	1.08	0.33	-0.01	0.99

will be discussed later. The kurtosis is the only other parameter that shows any noticeable change. In the heavy mineral fraction of samples 7, 8, and 9, and the light mineral fraction of sample 9, kurtosis is significantly higher than in other samples. No cause for this marked change in kurtosis is apparent.

Heavy Minerals Identified

The heavy minerals identified in this investigation were: apatite, aragonite, epidote, garnet, ilmenite, kyanite, magnetite, monazite, pleonaste, rutile, spinel, staurolite, tourmaline, and zircon. Tourmaline was observed in brown, green, blue, and rarely, pink color varieties. Apatite, epidote, pleonaste, spinel, rutile, and monazite occurred

Table 1. Results of Grain Size Analyses.

Sample Number	Size in millimeters 2.00-1.00	1.00-0.50	0.50-0.25	0.25-0.125	0.125-0.06	-0.06
1 (Light)	1.63%*	1.78%	39.95%	55.34%	1.30%	0.00%
2 (Light)	0.63	3.57	53.59	40.95	1.25	0.01
3 (Light)	0.17	0.45	20.74	76.38	2.25	0.01
4 (Light)	0.15	1.05	30.01	65.53	3.26	0.00
5 (Light)	0.14	0.78	30.32	68.09	0.67	0.00
6 (Light)	1.12	3.64	41.03	53.95	0.25	0.01
7 (Light)	0.40	3.42	38.81	57.09	0.25	0.02
8 (Light)	0.10	0.85	45.24	53.40	0.40	0.00
9 (Light)	3.62	3.03	19.71	72.61	1.03	0.00
10 (Light)	0.26	1.53	50.57	47.36	0.27	0.00
11 (Light)	0.06	0.51	28.45	70.47	0.51	0.00
12 (Light)	0.01	0.20	40.78	50.82	0.19	0.00
1 (Heavy)	2.19	6.95	47.17	40.38	3.25	0.06
2 (Heavy)	1.34	6.56	40.36	48.33	3.34	0.06
3 (Heavy)	0.01	0.50	10.12	68.36	20.98	0.03
4 (Heavy)	0.10	0.88	22.42	63.63	12.96	0.01
5 (Heavy)	0.08	1.42	21.88	65.05	11.53	0.04
6 (Heavy)	2.55	9.55	39.79	43.99	4.10	0.03
7 (Heavy)	2.10	12.14	11.96	70.18	3.48	0.14
8 (Heavy)	0.79	2.16	38.66	55.59	2.75	0.05
9 (Heavy)	3.08	4.72	15.60	62.18	14.37	0.05
10 (Heavy)	0.69	3.87	40.79	52.44	2.15	0.06
11 (Heavy)	0.06	3.30	28.90	59.53	8.18	0.02
12 (Heavy)	0.00	0.04	39.40	58.33	2.14	0.10

*Values are in weight percent.

with an average value of 1.11. The heavy fraction is quite similar, with a kurtosis ranging from 0.98 to 1.67 and averaging 1.20. Although the average values for kurtosis indicate these sands are leptokurtic, eight of the twelve samples in both the light and heavy fractions are mesokurtic. Thus, the sands show a nearly normal distribution with a slight tendency for the frequency curves to be "peaked", i.e., the central portion is slightly better sorted than the tails.

Little systematic change in the various sedimentary parameters can be observed from station to station along the beach. The mean, as well as the median, grain sizes tend to be noticeably smaller than normal in the area represented by stations 3 and 4. This is true for both the heavy and light mineral fractions. A possible cause for this

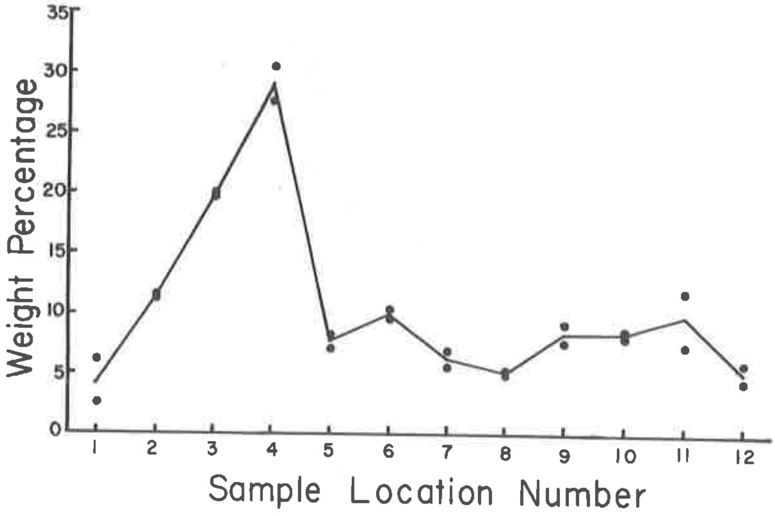


Figure 2. Weight percentage of total heavy minerals at the various collecting localities.

very erratically and in low concentrations; thus, the frequency values are not considered to be very reliable and are not considered further. In the following discussions magnetite is included with ilmenite; only a few grains of magnetite were observed in the entire study. This suite is similar, but slightly more complex, than suites reported for other beaches on the western coast of Florida (Phelps, 1941).

Percentage of Heavy Minerals

Total weight percentage of heavy minerals was determined for both sets of samples with almost identical results in each case. This indicates that very local fluctuations in heavy mineral content are not large and should not obscure the overall trend of heavy mineral occurrence along the beach. The average of the two determinations is used in the following discussions.

The amount of heavy minerals in the beach sands is generally between five and ten percent by weight (Figure 2) except at localities three and four, near the northern end of the beach, where the concentration of heavy minerals increases to as much as 29 percent. This concentration is noticeable in the color of the beach sand, which is considerably darker in this area than to either the north or to the south. The heavy mineral grains are easily visible in the sand, even upon casual observation. The range in amounts of heavy minerals found in this study corresponds with other values reported for beach sands in western Florida (Phelps, 1941).

Excluding the aragonite shell fragments from consideration, ilmenite is usually the most abundant heavy mineral in the sediment of Sand Key (Table 3), a situation that appears to be true in most Florida beach sands (Phelps, 1941; Tyler, 1934). Among the non-opaques, staurolite is usually the most abundant mineral except in the area of concentration (stations three and four). In this respect Sand Key seems to differ from other Florida beach sands; Phelps (1941) reports zircon to be the most abundant non-opaque in the areas he studied on both coasts of Florida and Tyler (1934) reports epidote to be the most abundant non-opaque mineral on the eastern coast.

DISCUSSION

The usual method of presenting heavy mineral data is to give the relative frequency of each mineral species as is done in Table 3. However, it is difficult to make interpretations based on relative abundance data alone. To illustrate this point, consider the relative abundance figures for aragonite at the 12 sampling locations. The relative abundance of this mineral at localities three and four is only about a third of its relative abundance along most of the beach. One might easily be tempted to think shell-producing organisms were scarcer in this area than along the rest of the beach or that aragonite shell fragments were being removed from the area or something equally complicated. Interpretation of the relative abundance of aragonite at stations three and four, as well as many other interpretations based on heavy minerals, can be improved if the weight percentage of a given mineral in the entire sample is used. To obtain the weight percentage, the relative frequency of each mineral in a given sample is multiplied by the specific gravity of the mineral. The products are summed and the sum is divided into each product to give the weight percentage of each mineral within the heavy mineral fraction. These percentages are then multiplied by the total weight percentage of heavy minerals within the sample to get the weight percent of each individual mineral within the sample. Although this method incorporates an error inasmuch as no correction is made for the differences in mean volumes of the various minerals, the resulting numbers are much more easily obtained and are probably more accurate than figures obtained by trying to separate and weigh the various individual heavy minerals (Hunter, 1967).

Using the aragonite example cited above, the usefulness of weight percentage in heavy mineral studies can now be demonstrated. Rather than decreasing sharply at stations three and four as indicated by the relative abundance data, the weight percent of aragonite in the beach sediment shows a gradual but erratic decrease northward from location twelve to location one (Figure 3). The amounts of aragonite at locations three and four are not significantly different from the trend. In fact, the amount of aragonite in sample four is actually slightly higher than anywhere else on the northern half of the beach, but this fact is masked

Table 3. Relative Amounts of Heavy Minerals in Beach Sands from Sand Key, Florida, in percent.

Mineral	Sample Number 1	2	3	4	5	6	7	8	9	10	11	12
Apatite	2.73	0.95	1.21	0.96	1.87	2.37	0.83	1.10	0.55	1.08	1.35	0.96
Aragonite	67.84	45.11	14.58	20.13	42.06	54.77	75.51	69.36	55.66	80.51	67.17	85.69
Epidote	0.00	0.00	0.00	0.80	0.00	0.00	0.15	0.21	0.56	0.33	0.11	0.10
Garnet	1.35	11.31	8.01	9.37	5.54	3.90	2.03	2.71	3.64	2.36	3.79	0.96
Ilmenite	7.88	16.67	17.17	17.72	17.86	11.15	8.62	7.03	13.21	3.77	10.00	3.55
Kyanite	1.87	1.09	2.11	1.34	3.01	3.71	1.40	1.37	4.23	1.89	1.85	1.34
Monazite	0.55	0.66	0.65	0.00	0.93	1.52	0.31	0.25	0.53	0.18	0.41	0.12
Pleonaste	0.11	0.00	0.00	0.26	0.19	0.00	0.08	0.00	0.00	0.00	0.22	0.00
Rutile	0.94	0.18	1.39	3.42	0.40	0.81	0.35	0.11	0.15	0.00	0.35	0.00
Spinel	0.21	0.00	0.00	0.00	0.00	0.15	0.00	0.00	0.00	0.00	0.00	0.05
Staurolite	8.77	13.95	10.84	12.85	12.36	11.15	6.47	8.23	9.10	5.48	8.69	3.44
Tourmaline	3.52	1.26	2.69	0.42	3.12	5.24	2.37	2.60	2.74	4.24	2.43	2.23
Zircon	4.54	8.83	41.33	33.79	12.88	5.38	2.12	7.24	10.20	0.49	3.96	1.70

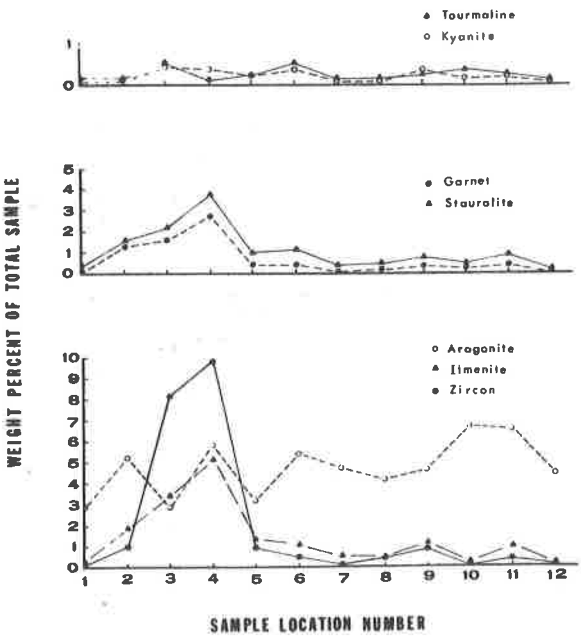


Figure 3. Weight percentage of tourmaline, kyanite, garnet, staurolite, ilmenite, and zircon.

in the relative abundance data by the increase in most other heavy minerals at this location. The other minerals show greater increases in concentration at this location than aragonite does; therefore, the relative abundance of aragonite decreases although its weight percentage doesn't. This example shows that weight percentage data, although harder to obtain, can lead to appreciably more realistic interpretations than can be obtained using relative abundance data.

The weight percentage of several of the more abundant heavy minerals as a function of sampling location is shown in Figure 3. The minerals have been divided into three groups to show the similarities in behavior of some of the minerals. Most show a sharp increase in concentration in samples three and four, although not all of the minerals are concentrated to the same degree. Zircon is the most abundant heavy mineral (excluding shell fragments), and is relatively constant along the beach except in the area of concentration at locations three and four. The average amount of each heavy mineral in the remaining ten samples was taken as the typical weight percentage of each mineral. The concentration factor is defined as the weight percentage of a particular mineral in a given sample divided by its typical weight percentage. At any particular place where heavy minerals are being concentrated, the concentration factors are mainly a direct function of the specific gravity

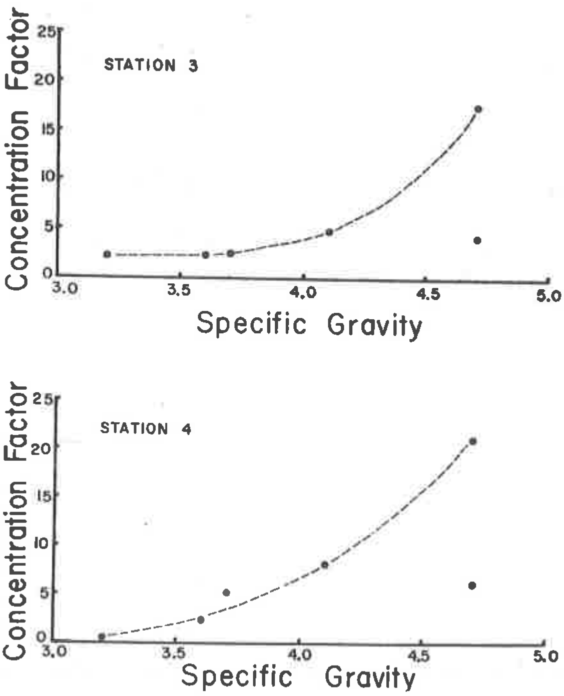


Figure 4. Concentration factors as a function of specific gravity at sampling stations three and four.

(Figure 4). In the Sand Key sediments, the only exception to this rule is ilmenite, which is concentrated to only about one-fourth the amount indicated by its specific gravity. This is attributed to the platy shape of the ilmenite grains, which decreases their settling velocities and allows them to be moved about more freely in the currents than they could be if they were spherical.

The concentration factors and the behavior of the minerals along Sand Key beach (Figure 3) are thus both functions of specific gravity and shape. Tourmaline and kyanite, both of which have relatively low specific gravities and are rod or blade shaped, show little tendency to be concentrated. Staurolite and garnet, intermediate in specific gravity and both roughly equidimensional in shape, behave almost identically. The high density zircon and ilmenite also behave in a similar fashion to one another except the platy shape of the ilmenite prevents its concentration to as great an extent as the zircon.

A rather poor correlation exists between the mean grain size of the quartz fraction in phi units and the percentage of heavy minerals in the sand. Most of the heavy minerals are concentrated at stations three and four, which also have the smallest mean diameters of quartz. A similar correlation, although not as well defined, exists between the median size of quartz and percentage of heavy minerals. No correlation was found between the amount of heavy minerals and skewness, kurtosis, or sorting of the light fraction.

The association of heavy minerals with the smaller grain diameters of quartz can possibly be explained using Hjulstrom's (1939) diagram, which relates grain diameter to the current velocity that must be attained to erode grains of a specific diameter. Hjulstrom found there is a grain diameter, approximately 0.5 mm, that requires lower current velocities for erosion than grains either larger or smaller. The beach sand of Sand Key is almost entirely smaller in diameter than this critical size. According to Hjulstrom's work, the larger grains on Sand Key beach, having a size corresponding to the diameter easiest to erode, ought to be removed preferentially to the smaller grains, which require progressively higher current velocities in order to be eroded. Hjulstrom's diagram is based on quartz; minerals with higher specific gravities are more difficult to erode. Thus, currents generated by wave action on the beach should preferentially set into motion the lighter minerals, especially the coarser fraction, and leave behind the smaller diameter quartz grains and the heavy minerals as a residual concentration.

The observed association of heavies with smaller diameter quartz grains is similar to a situation observed by McIntyre (1959), who used Ruby's (1933) method to compare ideal hydraulic diameters of heavy minerals associated with a given diameter of quartz to those actually observed on the beach. He found the observed hydraulic equivalent sizes of quartz for layers of beach sand that had been reworked by waves were too low in comparison with the grain diameters of the heavy minerals present. He suggested this might be due to selective removal of coarser quartz grains and further speculated that the selection process might be by the small quartz grains "hiding" among the heavy minerals and remaining behind while the larger quartz grains were removed.

The grain size data and heavy mineral distribution along the Sand Key beach seem to be best explained as being caused by wave erosion selectively removing quartz, especially the larger grain sizes, as well as some of the less dense heavy minerals, leaving behind a residual deposit enriched in high density heavy minerals and smaller diameter quartz grains.

SUMMARY

The fine-grained sand of the beach of northern Sand Key, Florida, contains a suite of at least fourteen heavy minerals, the most abundant of which are aragonite (shell fragments), ilmenite, zircon, staurolite and garnet. The total amount of heavy minerals is generally

between five and ten percent, except near the northern end of the beach where concentrations reach as much as 30 percent.

Weight percentage is more useful than relative frequency in making interpretations concerning heavy mineral distributions in these sands. Using weight percentage data, it can be shown that heavy mineral abundance in the area of concentration is a direct function of specific gravity with some modification for grain shape.

Large concentartions of heavy minerals on Sand Key show a slight tendency to be associated with smaller diameter quartz grains. This is thought to be due to greater ease of erosion of the larger diameters of quartz grains on Sand Key.

REFERENCES CITED

Folk, R. L., 1968, The petrology of sedimentary rocks: Hemphill's, Austin, Texas, 170 p.

Hjulstrom, F., 1939, Transportation of detritus by moving water: in Trask, P. D., editor, Recent marine sediments: Soc. Econ. Paleont. and Mineral, Tulsa, Okla., p. 5-31.

Hunter, R. E., 1967, A rapid method for determining weight percentages of unsieved heavy minerals: Jour. Sed. Pet., v. 37, p. 521-529.

McIntyre, D. D., 1959, The hydraulic equivalence and size distribution of some mineral grains from a beach: Jour. Geol., v. 67, p. 278-301.

Phelps, W. B., 1941, Heavy minerals in the beach sands of Florida: Proc. Fla. Acad. Sci., v. 5, p. 168-171.

Rubey, W. W., 1933, The size distribution of heavy minerals within a water-laid sandstone: Jour. Sed. Pet., v. 3, p. 3-29.

Tyler, S. A., 1934, A study of sediments from the North Carolina and Florida coasts: Jour. Sed. Pet., v. 4, p. 3-11.