Proceedings of the 1988 U.S. Geological Survey Workshop on the Geology and Geohydrology of the Atlantic Coastal Plain

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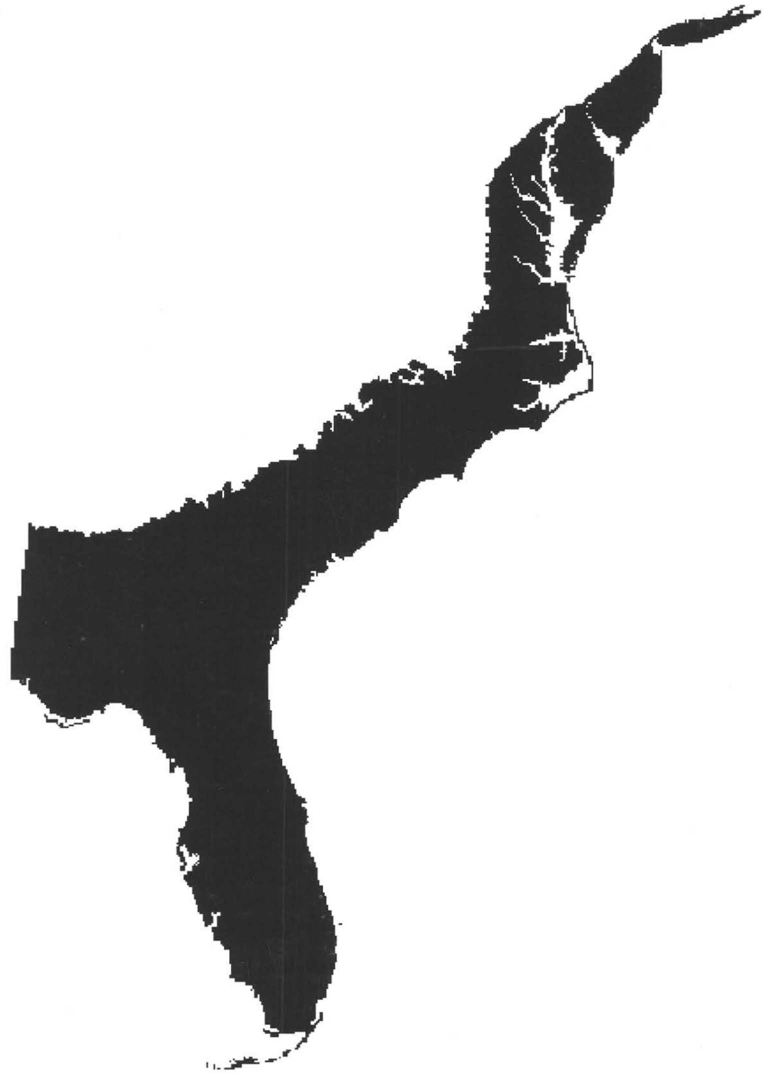

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Proceedings of the 1988 U.S. Geological Survey Workshop on the Geology and Geohydrology of the Atlantic Coastal Plain

Edited by GREGORY S. GOHN

A collection of short scientific papers that discuss research presented at the workshop held September 28–29, 1988, in Reston, Va.

U.S. GEOLOGICAL SURVEY CIRCULAR 1059

U.S. DEPARTMENT OF THE INTERIOR
MANUEL LUJAN, Jr., Secretary

U.S. GEOLOGICAL SURVEY
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PREFACE

This Circular records the proceedings of a workshop sponsored by the U.S. Geological Survey (USGS) on the geology and hydrology of the U.S. Atlantic Coastal Plain. The workshop was held on September 28–29, 1988, in the U.S. Geological Survey's National Center in Reston, Va., and was attended by more than 80 State and federal scientists and scientific managers. Geological surveys and water-resource agencies from seven Coastal Plain States were represented, as were the Geologic Division and Water Resources Division of the USGS. The goal of the workshop was to provide a forum in which Atlantic Coastal Plain researchers representing a variety of agencies, geologic disciplines, and geographic areas could meet and exchange scientific and programmatic information. An abridged version of the workshop agenda is given in the following section.

Scientific presentations at the workshop were grouped into three technical sessions for oral presentations and one poster session. The technical sessions were grouped along geographic lines; the opening session featured results of an ongoing cooperative research effort in New Jersey between the New Jersey Geological Survey and the USGS. The subsequent session concentrated on the northern part of the U.S. Atlantic Coastal Plain exclusive of New Jersey, and a following session concentrated on the southern part of the Atlantic Coastal Plain as well as the eastern Gulf of Mexico Coastal Plain. Seventeen posters on a variety of topics were exhibited at the poster session. In addition to the technical sessions, a panel and audience discussion and three regional discussion groups were organized to discuss existing programs and new directions in Atlantic Coastal Plain research.

All individuals in attendance at the workshop were invited to contribute to this proceedings volume. Twenty-six papers were submitted in response to that invitation. These reports are presented in two following sections: one section is devoted to the deliberations of the workshop's regional discussion groups plus one related article on research directions and needs, and the other section is devoted to technical articles. Collectively, these 26 articles represent much of the spectrum of ongoing research in the U.S. Atlantic Coastal Plain and suggest future research directions that will become increasingly important in the coming decade.

Gregory S. Gohn

Gregory S. Gohn
Editor

WORKSHOP AGENDA (ABRIDGED)

WEDNESDAY, SEPTEMBER 28, 1988

8:00 a.m.	Registration
8:30 a.m.	Introduction and welcome
8:40 a.m.	Opening remarks *John Dragonetti,* USGS Assistant Director for Intergovernmental Affairs
9:00 a.m.	Session 1: New Jersey Coastal Plain *Richard Dalton,* New Jersey Geological Survey, Presiding
1:30 p.m.	Session 2: Northern Atlantic Coastal Plain Exclusive of New Jersey *Lucy McCartan,* USGS, Presiding
6:30 p.m.	Session 3: Poster session *David Soller,* USGS, Coordinator

THURSDAY, SEPTEMBER 29, 1988

9:00 a.m.	Session 4: Southern Atlantic Coastal Plain and Eastern Gulf of Mexico Coastal Plain *Henry Trapp,* USGS, Presiding
10:50 a.m.	Session 5: Atlantic Coastal Plain Research—Panel and Audience Discussion on Existing Programs and New Directions in Atlantic Coastal Plain Research *Juergen Reinhardt,* USGS, Presiding
1:00 p.m.	Session 5—Continued

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New Jersey Coastal Plain

Central Atlantic Coastal Plain

Southeastern Coastal Plains

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CONVERSION FACTORS

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
mile (mi)	1.609	kilometer (km)
nautical mi (nmi)	1.852	kilometer (km)
angstrom (Å)	10^{10}	meter (m)
millimeter (mm)	0.03937	inch (in.)
centimeter (m)	0.3937	inch (in.)
meter (m)	3.28	foot (ft)
kilometer (km)	0.6214	mile (mi)
	0.5400	nautical mile (nmi)
	Area	
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
cubic foot (ft^3)	0.02832	cubic meter (m^3)
liter (L)	1.057	quart (qt)
cubic centimeter (cm^3)	0.061	cubic inch ($in.^3$)
	Mass	
ton (2,000 pounds)	0.9072	metric ton (1,000 kg)
gram (g)	28.35	ounce avoirdupois (ozavdp)
	Pressure	
pound-force per square inch ($lbf/in.^2$)	6,895	pascal (Pa)
	Transmissivity	
square foot per day (ft^2/d)	0.0929	square meter per day (m^2/d)
	Flow	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
	Acceleration	
milligal (mGal)	0.00001	meter per second squared (m/s^2)
	Radioactivity	
picocurie (pCi)	0.037	becquerel (Bq) = 1 disintegration per second

Temperature.—Temperatures can be converted as follows:

$$\text{degree Fahrenheit (°F)} = 1.8 \text{ degree Celsius (°C)} + 32$$
$$\text{degree Celsius} = (\text{degree Fahrenheit} - 32)/1.8$$

Age designations.—The time of a geologic event and the age of an epoch boundary are expressed as Ma (mega-annum), and intervals of time are expressed as m.y. (million years). Both terms mean 1,000,000 years, or years$\times 10^6$. For example, sediments were deposited at 85 Ma (85×10^6 years before 1950 A.D.), and the deposition continued for the next 2 m.y.

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REVIEWS OF REGIONAL RESEARCH GOALS

1. Regional Discussion Group 1—The Northern Atlantic Coastal Plain in New Jersey, Pennsylvania, Delaware, and New York

Compiled by James P. Owens[1] *and* Peter J. Sugarman[2]

The regional discussion group for the northern Atlantic Coastal Plain consisted primarily of scientists from the New Jersey Geological Survey, the Pennsylvania Geological Survey, and the Trenton, N.J., and Reston, Va., offices of the U.S. Geological Survey. The group focused on new research directions in the New Jersey Coastal Plain but also discussed stratigraphic and hydrologic relations among Coastal Plain units of New Jersey, Long Island, Pennsylvania, and Delaware. The following topics were suggested by group members for possible future research programs and cooperative studies.

Saltwater intrusion.—Studies of saltwater intrusion into Coastal Plain aquifers are essential because of the threat this phenomenon poses to ground-water resources. Measurements of saltwater heads are particularly important because only limited information is presently available. Suggested studies related to this topic include the following:

1. Establishment of a regional monitoring-well network for identifying coastal areas where saltwater intrudes and for better understanding the hydrology of saltwater intrusion.
2. Studies of saltwater intrusion into the Potomac-Raritan-Magothy aquifer systems along Delaware Bay. This work would probably require a cooperative, multi-State effort.
3. Studies of saltwater intrusion into the Potomac-Raritan-Magothy aquifer system along Raritan Bay. This investigation could also include study of the movement of lead and cadmium into the aquifer system along with the saline water.
4. Hydrogeochemical studies (including core squeezing) of confining units that contain brackish or saline water. Saltwater intrusion is believed to change the physical properties of confining beds.

Middle Tertiary aquifers.—Relations between the stratigraphy and hydrology of middle Tertiary units in southern New Jersey and northern Delaware need study. The emphasis of this study would be on Eocene, Oligocene, and Miocene aquifers and their stratigraphic and hydrologic relations across Delaware Bay.

Offshore geology of Monmouth and Ocean Counties, N.J.—An obvious data gap highlighted by the conference is the lack of information regarding the transition of geologic and hydrogeologic units from the Coastal Plain to the adjacent offshore areas. In the northern New Jersey Coastal Plain and adjacent offshore areas, a study of the Kirkwood Formation, the Cohansey Sand, and overlying Pleistocene formations could incorporate stratigraphic, isotopic (radiocarbon), mineralogic, and geomorphic investigations. A second study could focus on the deeper aquifer units and their continuation into the offshore area.

Geochemical map of New Jersey.—A geochemical map of New Jersey would present water-chemistry data for the numerous subsurface formations in the Coastal Plain. It could utilize existing water-quality data and NURE

Manuscript approved for publication January 28, 1991.
[1]U.S. Geological Survey, Reston, VA 22092.
[2]New Jersey Geological Survey, Trenton, NJ 08625.

(National Uranium Resource Evaluation) data. This map would be useful in regulating underground storage tanks where ground-water corrosion affects tank life.

Geology of impermeable materials.—A project on impermeable materials would involve varied studies of clay and silt in the Coastal Plain. Geologic studies would involve field mapping to produce a resource map and the use of facies models to predict clay occurrences. Hydrologic studies would include measurements of permeability and other properties of clay and silt layers and their geochemical characteristics as confining units.

Raritan Formation in the northern New Jersey Coastal Plain.—The Woodbridge Clay Member of the Raritan Formation defines the boundary between the upper and middle aquifer systems in the northern New Jersey Coastal Plain. Its surface and subsurface distribution and its physical properties are important for the understanding of these major aquifer systems. This study could involve core drilling, palynologic studies, and seismic-reflection surveys.

Amboy trough between Trenton and New Brunswick, N.J.—Geologic studies of the Amboy trough would require integrated investigations of the stratigraphy and geomorphology of the Cretaceous and Pleistocene sections in that area and structural implications derived from these data. Scattered evidence in the trough suggests some tectonism in this region. Detailed mapping of the surficial sediments in the trough would define the extent and timing of these structural movements.

Pennsylvania Coastal Plain.—Relatively few investigations have been conducted in the Pennsylvania Coastal Plain during the past decade. Consequently, if studies were to be initiated, they should be fundamental studies of the stratigraphic framework. The need for geologic studies related to pollution of Coastal Plain units in the Delaware Valley could provide a focus for such work.

2. Regional Discussion Group 2—Summary of Research Directions in the Central Atlantic Coastal Plain in the Salisbury Embayment and North Carolina

Compiled by C. Wylie Poag[1] *and* Robert Shedlock[2]

Members of the regional discussion group concerned with research needs in the Salisbury embayment and North Carolina segment of the U.S. Atlantic Coastal Plain included representatives from the Geologic Division (GD) and Water Resources Division (WRD) of the U.S. Geological Survey (USGS) and from the Maryland Geological Survey (MGS), the Delaware Geological Survey (DGS), the Virginia Water Control Board (VWCB), and the North Carolina Department of Natural Resources-Groundwater (NC–DNR–G). Active participants included Emory Cleaves (MGS), Kelvin Ramsey (DGS), Scott Bruce (VWCB), Ted Mew (NC–DNR–G), Norm Sohl (GD), Wylie Poag (GD), Bob Mixon (GD), Dave Prowell (GD), Lucy Edwards (GD), Norm Fredericksen (GD), Steve Colman (GD), Ron Litwin (GD), Dave Powars (GD), Helaine Markewich (GD), Bob Shedlock (WRD), and Steve Hiortdahl (WRD).

Although the charge of this discussion group was to agree on new research initiatives for this area, the limited time did not permit the group to develop a consensus on any detailed new initiatives. Instead, the group focused on three general and fundamental targets for further investigation.

1. Nomenclature and Correlation of Depositional Units

Everyone agreed that more communication is needed among geologists and hydrogeologists, especially to iron out problems of stratigraphic nomenclature, complex facies changes, and biostratigraphy of noncalcareous units. Multidiscipline research teams and interdisciplinary workshops, such as the present one, are fruitful means of improving this communication and providing more comprehensive research efforts.

[1]U.S. Geological Survey, Woods Hole, MA 02543.

[2] U.S. Geological Survey, Towson, MD 21204.

2. Centralized Field-Data Library

The group expressed a widely felt need to assess currently available field data, such as core collections and libraries of borehole logs, which are held by various institutions represented in this group. These data are not readily accessible to all because they reside in the files of different agencies, which have somewhat different missions. Ron Litwin volunteered to compile a directory of such data and asked that workers from each agency or division send him brief descriptions of their geologic or hydrogeologic data. Dave Powars agreed to compile a summary of drilling capabilities and resources now supported by different institutions.

3. Geologic Data on the Nearshore Region of the Continental Shelf

The most clearly articulated research direction was to carry out more nearshore test drilling. This drilling is essential to refine conceptual models of geologic and hydrogeologic units, to establish a detailed structural and depositional framework, and to examine the relationship of Coastal Plain deposits to downdip facies in the Baltimore Canyon trough. Such work would also have economic significance as it would provide important information relevant to the intrusion of saline waters into the water supplies of coastal cities. An interagency effort was suggested, but time did not permit discussion of a plan of action.

3. Regional Discussion Group 3—Research Needs and Areas For Cooperative Programs in the Southern Carbonate Province

Compiled by Irwin H. Kantrowitz[1] *and* Henry Trapp, Jr.[2]

INTRODUCTION

Members of the regional discussion group concerned with research needs in the southern carbonate province (Florida and parts of Alabama, Georgia, and South Carolina) represented the Geologic Division (GD) and Water Resources Division (WRD) of the U.S. Geological Survey (USGS) and the Florida Geological Survey (FGS). The discussion emphasized areas of overlapping or complementary interests among scientists of these groups. Present for the discussion were Jonathan Arthur (FGS), Paulette Bond (FGS), Irwin Kantrowitz (WRD), Harley Knebel (GD), Lucy McCartan (GD), Ronald Miller (WRD), Peter Popenoe (GD), Juergen Reinhardt (GD), Walter Schmidt (FGS), Henry Trapp, Jr. (WRD), and Thomas Yorke (WRD). Irwin Kantrowitz served as moderator and prepared the first draft of this report; Henry Trapp served as recorder of the discussion. All members of the group were given an opportunity to review and revise the draft, as was James Miller (WRD), who was unable to attend the workshop. This report summarizes the cumulative thinking of the group regarding fruitful areas of geologic research in the southern carbonate province. It was beyond the scope of this report to document existing literature, active research, and available data; however, this body of knowledge is considerable and would provide the starting point for any future research.

The major area of needed research identified by the group involves a better description and delineation of aquifers and confining beds in the Tertiary sediments comprising and overlying the Floridan aquifer system. In particular, detailed three-dimensional mapping of the material between the surface and a depth of 200 ft is necessary in order to characterize the nature of ground-water recharge and to map shallow aquifers, which commonly are discontinuous. The areal extent of the Floridan aquifer system is approximately coincident with the southern carbonate province (fig. 3.1). Because of the heavy use made of the Floridan and overlying aquifers for water supply and because of the relatively large number of agencies at the Federal, State, and local levels having active programs in hydrogeology (geology and ground-water hydrology), prospects for cooperative efforts are favorable. Within the broad category of hydrogeology, detailed lithostratigraphy, biostratigraphy, geophysical studies, and mineralogy were thought to be the most promising subjects for cooperative programs. Other broad research topics identified were the distribution of radon-producing sediments and vertebrate paleontology.

[1]U.S. Geological Survey, Tallahassee, FL 32301.
[2]U.S. Geological Survey, West Trenton, NJ 08628.

HYDROGEOLOGY

The major controlling factor in obtaining a better understanding of the geology, and, consequently, the hydrogeology, of the southern carbonate province is the flatness of its surface, which restricts exposure of the rocks. Additional geologic understanding will have to come from the interpretation of subsurface data obtained either directly through cores and drill cuttings or indirectly through the use of geophysical methods.

High-resolution seismic-reflection profiling, combined with core drilling, was suggested as a possible means of collecting data in some areas. Offshore seismic lines have been run by oil companies as well as by the Minerals Management Service (U.S. Department of the Interior). Although many of these data may be proprietary, some may be available for purchase or inspection. In addition, some new acoustic profiles should be collected, particularly to correlate the offshore data (commonly obtained beyond State jurisdiction) with data from nearshore and estuarine areas. Onshore Vibroseis exploration and, possibly, ground-penetrating radar offer a means of extending these correlations into areas underlain by aquifers containing freshwater.

Another possible method of obtaining data inland would be to conduct seismic surveys along some rivers or canals. However, applying seismic techniques to these narrow, water-covered areas poses a sizable challenge.

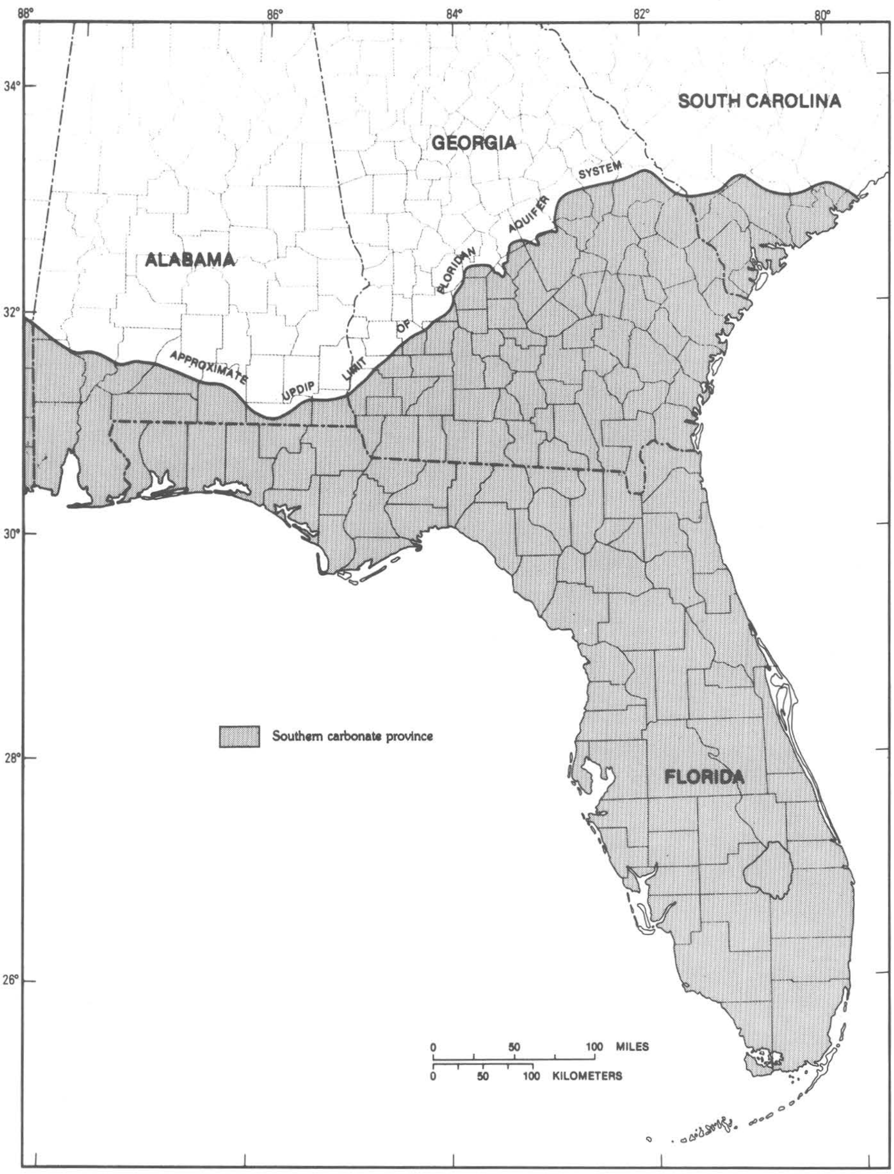

Figure 3.1. Location of the southern carbonate province, which is approximately coincident with the areal extent of the Floridan aquifer system. Modified from Miller (1986).

Difficult acoustic conditions are expected to be caused by (1) extreme shallow water (ringing), (2) hard carbonate rocks (poor signal penetration), (3) irregular karst environments (signal attenuation and sidescattering), and (4) gas in the sediments (poor signal penetration). The data from these environments must be acquired digitally in order to take advantage of processing techniques such as deconvolution, frequency-wave-number filtering, and statics corrections that enhance signal and attenuate noise. In addition, a feasibility study would need to be conducted to determine which equipment could best be used in these environments. This feasibility study would involve using different sound sources (such as water guns, air guns, and Uniboom), different streamers (single channel and multichannel), and different source-streamer offsets. With this suite of comparative data, one could choose and arrange the field equipment in order to obtain the best reflection records.

Geophysical logs run in coreholes are another means of areally extending geologic knowledge of the subsurface. Many water wells have gamma-ray logs, even if little or no other geologic information has been collected. Comparison of these relatively abundant logs with the logs run in carefully studied coreholes, which, in turn, are tied to seismic lines, may permit the correlation of detailed point data over broad areas. Additional coreholes to act as stratigraphic tie points may be needed to verify regional geologic interpretations.

In addition to enhancing the areal interpretation of structure and stratigraphy based on geophysical lines and downhole logs, coreholes could provide material for the detailed study of mineralogy and biostratigraphy. Detailed mineralogic study of core material may prove useful in explaining and predicting the chemical quality of ground water as well as the ability of confining beds to retard the movement of some ions.

Biostratigraphic studies of cores may prove useful in correlating key marker beds, thus permitting the subsurface mapping of hydrogeologic units and the identification of geologic structures affecting ground-water movement. Fossil assemblages, as well as lithology and mineralogy, provide information on depositional environments and the geologic history of the rocks subsequent to deposition. Mapping of depositional lithofacies is essential to accurate prediction of trends in rock permeability. Although recrystallization of the carbonate rocks may have obscured the fossil content, other methods, such as study of insoluble residues and stable isotopes, may prove useful.

RADON

The presence of radon in enclosed areas as a result of outgassing from radon-producing sediments is an emerging national issue. Uranium, the radon parent material, is commonly found in association with phosphate. Radon health hazards may be a particular problem in Florida because of the occurrence of the phosphate-rich Hawthorn Formation throughout much of the State. The FGS has an advisory role in the State's efforts to deal with the radon problem, but it is not yet clear whether there will be research opportunities—that is, geologically oriented problems warranting future study. Elsewhere, however, high radon concentrations have been associated with faults and fissures that have not been previously mapped. Therefore, in addition to providing baseline information on a potential health hazard, radon surveys may provide clues about local stratigraphy and structural geology.

VERTEBRATE PALEONTOLOGY

The Tertiary deposits of Florida contain a rich vertebrate fauna, and the Florida State Museum has the ability to protect fossiliferous areas for study prior to development. The FGS has no program in vertebrate paleontology and the USGS program is extremely limited. Therefore, despite the opportunities, there is little likelihood of research in the near future.

SUMMARY AND CONCLUSIONS

If acquisition and processing problems can be overcome, then seismic profiles can be combined with geophysical logs interpreted on the basis of drill-hole cores and cuttings to augment knowledge of the extent, distribution, and characteristics of aquifers and confining beds within the southern carbonate province. Because the USGS (both the Water Resources and Geologic Divisions) and the FGS have mutual interests and complementary abilities in the broad research field of hydrogeology, prospects for cooperative research appear favorable.

To enhance any formal research efforts, an inventory of available data, including offshore and onshore seismic lines and geophysical logs, needs to be assembled. Other actions, such as requesting access to data from the shallow part of oil or injection wells drilled in the future, can greatly add to our knowledge of the subsurface at a relatively modest cost.

REFERENCE CITED

Miller, J.A., 1986, Hydrogeologic framework of the Floridan aquifer system in Florida and in parts of Georgia, Alabama, and South Carolina: U.S. Geological Survey Professional Paper 1403–B, 91 p.

4. Existing Programs and New Directions in Hydrogeologic Research in the Atlantic Coastal Plain—The West-Central Florida Perspective

By Thomas H. Yorke[1]

This report on existing programs and new directions in Atlantic Coastal Plain research is presented from the perspective of the Florida water managers and is based on day-to-day dealings with many State and local officials. The Tampa Subdistrict office of the Water Resources Division, U.S. Geological Survey (USGS), has a cooperative program with 15 local jurisdictions, including city, county, and State agencies and the Southwest Florida Water Management District. These agencies are responsible for providing an adequate supply of potable water to the ever-increasing population of west-central Florida or are responsible for protecting the water resources.

Numerous discussions among members of the USGS and representatives of these agencies have focused on five major areas of concern about ground water in west-central Florida. These concerns are summarized by the following questions: (1) What are the hydraulic and physical characteristics of water-bearing and confining units that control the horizontal and vertical movement of saltwater into freshwater zones? (2) What are the rate and direction of water movement between the water-bearing units of the intermediate aquifer system (interbedded clays and limestones between the surficial materials and the Upper Floridan aquifer) and the Upper Floridan aquifer, and how will additional pumping affect the rate and direction of movement? (3) What is the effect of ground-water withdrawals on the quantity and quality of water in streams, lakes, and estuaries? (4) What is the potential for contamination of the Floridan aquifer system by toxic agricultural and industrial chemicals? (5) What changes occur in the chemical character of ground water as it moves through and between different aquifer and confining units?

The Tampa Subdistrict and Florida District offices of the Water Resources Division are addressing most of these issues as part of a jointly funded investigative program with the Florida Department of Environmental Regulation, Florida Water Management Districts, counties, and municipalities. Several studies are underway to define the saltwater-freshwater interface and the potential for saltwater intrusion along the west coast of Florida. The purpose of one study is to locate and sample wells on or near the interface; the purpose of another study is to investigate the areal and temporal variation in the quantity and quality of discharges of major springs on the coast in Hernando and Citrus Counties. Digital models are being used to simulate the transport of salt and other chemically conservative constituents in ground water as part of three separate investigations.

Other projects are investigations of the migration of water between different water-bearing units of the intermediate and Floridan aquifer systems. Two studies include analyses of hydraulic characteristics and chemical constituents to determine directions and rates of water movement and changes in the chemical quality of the water. A third study is an assessment of the potential for contamination of the Floridan aquifer; concentrations in water of tritium, oxygen isotopes, and other indicators of relative age are used to evaluate ease of recharge.

Stochastic and deterministic models are being used in several current studies to evaluate the relation between ground water and surface water. Statistical relations have been developed to estimate changes in streamflow and lake levels that would result from additional pumping and concomitant declines in ground-water levels. Ground-water flow models are being used to estimate changes in spring and stream discharges due to declines in ground-water levels caused by droughts and by increased withdrawals for irrigation and municipal supplies. Both statistical and numerical models are being used to evaluate potential water quality and biological changes in estuarine environments.

Although many of the concerns of water managers in west-central Florida are being addressed by current investigations, additional hydrogeologic research is needed in several areas. Answers to many of the questions depend on a better understanding of secondary porosity in the carbonate rocks that form the intermediate and Floridan aquifer systems. The approach being taken by many of the Water Management Districts is to obtain core, geophysical, and aquifer test information from a dense network of observa-

[1]U.S. Geological Survey, Reston, VA 22092.

tion wells and boreholes. Although this information is invaluable, it is site specific and very expensive. Some of this effort could be redirected toward identifying lineations on aerial photographs, acquiring borehole televiewer images, and using surface geophysical techniques to improve our capability for defining fractures, solution features, and flow patterns.

The mineralogy and quality of the pore water in the clay beds of the intermediate and Floridan aquifer systems are among the least understood characteristics of hydrogeology in west-central Florida. This information is important for predicting changes in water quality that would result from vertical migration of water induced by additional pumping. These characteristics also are important for evaluating the migration of contaminants from the surface or deep waste-injection zones to the potable-water zones. The mineralogy of the clays will influence how long the confining units will retard the movement of contaminants and whether there will be differential sorption of certain types of toxic compounds.

Another topic for geologic research is the distribution of sediments in tidal rivers and estuaries. Water managers in west-central Florida and all coastal States are faced with multibillion-dollar decisions on alternatives for restoring and maintaining the quality and productivity of these water bodies. The mechanisms affecting the distribution of sediments within the affected water bodies and the interchange of nutrients and contaminants between the sediments and overlying water must be determined. Research on both the historical depositional patterns of sediments in the estuaries and the current sediment transport characteristics will be useful in assessing alternatives for improving the quality of our estuaries. Ultimately, the research must lead to analytical tools such as digital models for predicting how various land and water development activities will affect the sediment balance, water quality, and biological interactions of tidal rivers and estuaries.

TECHNICAL PAPERS

New Jersey Coastal Plain

5. Calcareous Nannofossils—Their Use in Interpreting Paleocene and Eocene Geologic Events in the New Jersey Coastal Plain

By Laurel M. Bybell[1]

INTRODUCTION

Since 1985, as part of a cooperative project between the New Jersey Geological Survey (NJGS) and the U.S. Geological Survey (USGS), I have examined Paleogene samples from 19 wells in the New Jersey Coastal Plain for their calcareous nannofossil content (fig. 5.1). Calcareous nannofossils are now being used to accurately date many marine sediments in New Jersey. This information provides a data base that is currently being used to develop a model for interpreting the geologic history of this area; unconformities and other features dated by nannofossils aid in modeling the timing of tectonism, eustatic sea-level changes, and depocenter shifts.

Acknowledgments

I thank James P. Owens (USGS) and Peter J. Sugarman (NJGS) for providing all the well samples, for constant cooperation, and for helpful suggestions, without which this paper would not have been possible. I thank Jean Self-Trail (USGS) for processing many of the samples and drafting the figures. I appreciate the thoughtful reviews of James P. Owens and Wayne L. Newell (USGS).

METHODS

Samples for this study are from cores, cuttings, and sidewall cores and even from old wells drilled by percussion tools. Because many of the samples were not from recent coreholes, and it was impossible to determine what precautions, if any, originally had been taken to avoid contamination, I had to be constantly aware of the possibility of downhole contamination, reworking, and drilling-mud contamination. Where necessary, I even performed species counts to ensure the utmost accuracy of the paleontological results.

I used the calcareous nannofossil zones of Martini (1971). No one well contained the entire Paleocene and Eocene sequence from Zone NP 1 to the middle of Zone NP 21, but data from the 19 wells could be pieced together to produce a complete section. In producing the composite section, I was careful to consider geographic location, facies changes, and updip-downdip relations. This report focuses on Paleocene through upper middle Eocene sediments (Zones NP 1–16, fig. 5.2), although upper Eocene, lower Oligocene, and upper Oligocene sediments are also present in the subsurface of the New Jersey Coastal Plain. The ACGS–4 core, which contains the most detailed information from the subsurface of New Jersey, was described by Owens and others (1988) and Poore and Bybell (1988).

Once samples from all the wells were examined and each sample was placed within a nannofossil zone, I measured the thickness of each NP zone in each hole. Zones NP 1–13 each represent about 1 m.y., and Zones NP 14–16 each represent about 3–4 m.y. according to the time scale of Berggren and others (1985). For a nannofossil zone at the top or bottom of a well or next to a barren interval, I assume that only a minimum thickness is represented. However, a barren interval could also indicate changing depositional conditions or a transition zone into the next nannofossil zone, and if it did, little sediment might be missing.

In addition to being used to compile a complete vertical section, the nannofossil data were also used to map variations in sediment thickness across the New Jersey Coastal Plain at a particular time represented by one calcareous nannofossil zone. The first step was to take the surface elevation of each well and correct it to sea level. The base of each nannofossil zone was plotted and corrected

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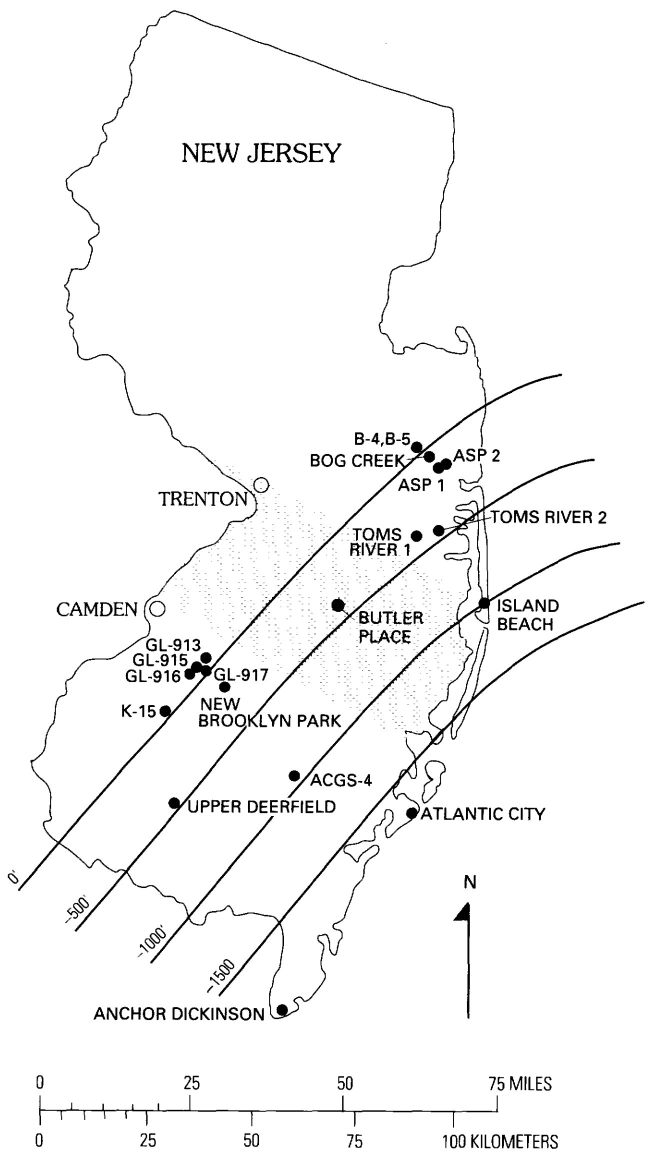

Figure 5.1. Locations of New Jersey wells yielding samples that were examined for calcareous nannofossils. Contour lines indicate feet below present sea level for base of Zone NP 12. Shading indicates the Camden-Trenton band, an area that may have been tectonically active during the Paleocene and Eocene.

to feet below and above current sea level. For each NP zone, contour lines of equal depth were plotted relative to sea level. The contour lines that were plotted across the approximately 40-mi-wide New Jersey Coastal Plain that makes up the study area are at 0, 500, 1,000, and 1,500 ft below present sea level; as an example, figure 5.1 shows the contours for the base of the lower Eocene Zone NP 12. These sediments dip seaward at about 35 ft/mi, and this amount of dip implies that tectonic tilting has probably taken place since deposition. In this report, however, the direction, not the amount, of dip is what is most significant in determining variations in sediment thickness, and thus the positions of depocenters, in the New Jersey Coastal Plain.

RESULTS

In the simplified passive-margin model proposed in this report for the New Jersey Coastal Plain during the Paleocene and early to middle Eocene, individual calcareous nannofossil zones are represented by thin layers of sediment that generally become somewhat thicker, but not significantly so, downdip to the southeast. There is no evidence for the South New Jersey high or the Raritan embayment or any other significant persistent topographic feature in the New Jersey Coastal Plain during the Paleocene or Eocene. The variation in thickness within each zone is minor (usually less than 50 ft) in comparison to the current dip of these deposits to the southeast (1,500 ft in 40 mi). However, these data on thickness variation, used in conjunction with lithologic studies, are invaluable for determining local geologic processes. Using calcareous nannofossil data, I plotted the varying patterns through time that are the result of changes in sedimentation rate, shifts in depocenters, tectonism, and eustatic sea-level changes. I discuss below what appears to have been happening in New Jersey during the time of deposition of each nannofossil zone from the base of the Paleocene up through the early middle Eocene.

Paleocene, Zones NP 1–9

As is described below, the Paleocene strata on the New Jersey Coastal Plain record repeated eustatic sea-level changes, as well as complex lateral and vertical shifts in depocenters. Local basinal forces appear to have acted separately north and south of the region between Camden and Trenton and the present shoreline. This region (shaded in fig. 5.1) appears to have been tectonically active at various times during the Paleocene. It is called the Camden-Trenton band in this report.

Zone NP 1, 66.4–66.0 Ma.—Zone NP 1 is the oldest NP Zone in the Paleocene, and it is not present in any of the 19 holes examined to date in New Jersey. Thus, the unconformity at the Cretaceous-Tertiary boundary in New Jersey represents a minimum of 500,000 yr according to the time scale of Berggren and others (1985).

Hornerstown Sand, Zones NP 2–4 (Lower Paleocene)

Zone NP 2, 66.0–65.0 Ma.—Zone NP 2 was identified only in core ASP 1 (fig. 5.1). For most of the New Jersey Coastal Plain, therefore, the unconformity at the Cretaceous-Tertiary boundary is indicated by the absence of Zone NP 2, as well as Zone NP 1, and represents about 1.4 m.y.

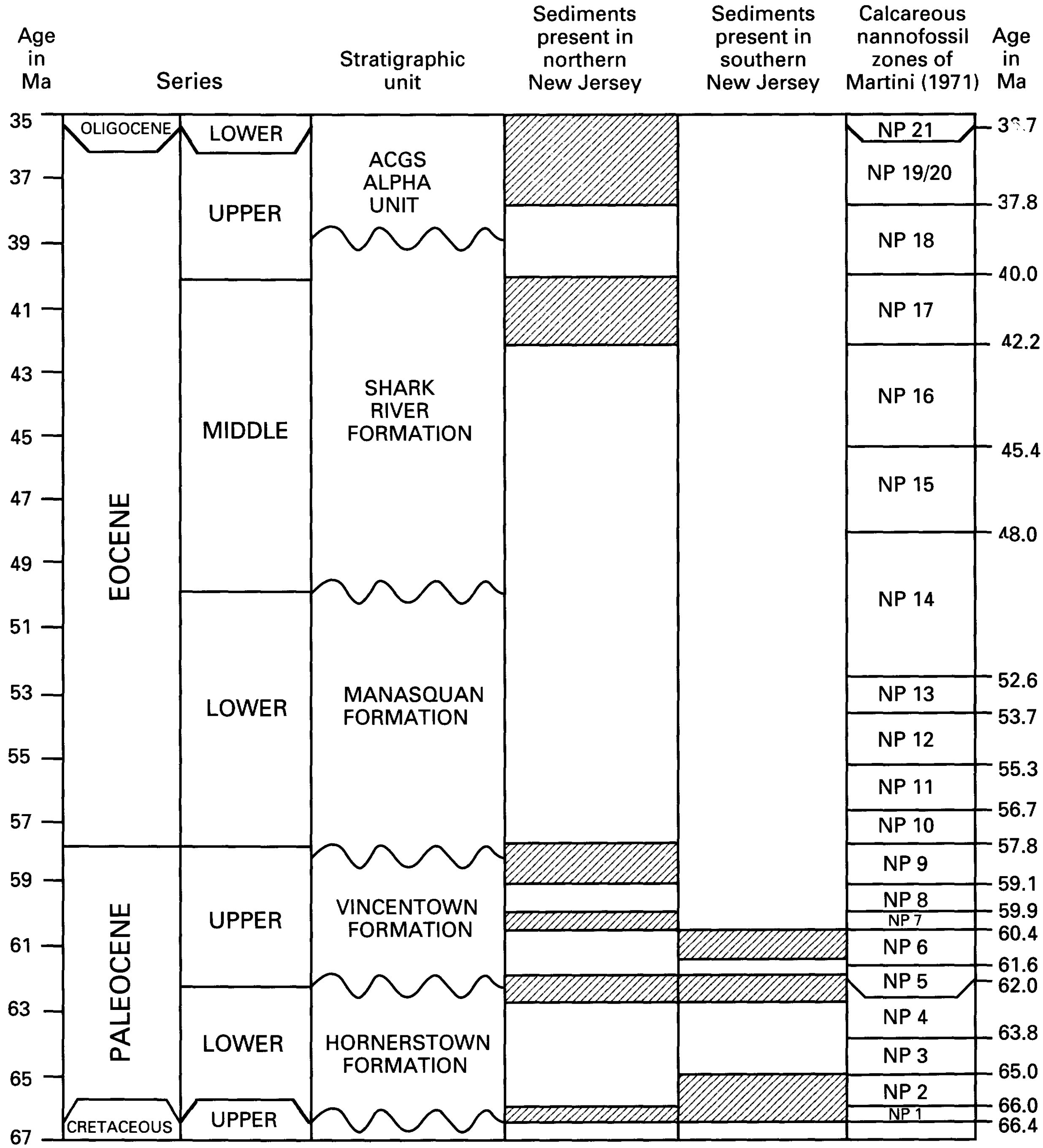

Figure 5.2. New Jersey Paleocene and Eocene stratigraphic units. Diagonal stripes show major unconformities in the regions north and south of the Camden-Trenton band. Time scale is from Berggren and others (1985).

Zones NP 3–4, 65.0–62.0 Ma.—Both outcrop and subsurface sediments of the Hornerstown from Zones NP 3 and NP 4 are preserved only in updip locations (GL and ASP holes, fig. 5.1). The absence of this material from downdip localities may be a result of erosion of the Hornerstown during the subsequent, less extensive transgression recorded in the upper part of Zone NP 5. Zones NP 3 and NP 4 are each about 5–10 ft thick, and the contour lines for these two zones indicate that the sediments dip relatively uniformly to the southeast. These sediments represent a widespread lower Paleocene transgression that occurred not only in New Jersey, but also in Virginia and Maryland, and that probably resulted from a global sea-level rise at about 65 Ma.

Vincentown Formation, Zone NP 5–Lower Part of Zone NP 9 (Upper Paleocene)

Zone NP 5, 62.0–61.6 Ma.—As mentioned above, there is an unconformity at the Hornerstown-Vincentown boundary, which is marked by the absence of the lower part of Zone NP 5, and probably the upper part of Zone NP 4. The thickness of Zone NP 5 sediments increases somewhat downdip to the southeast from about 5 ft in the GL cores to about 20 ft at Island Beach (fig. 5.1). The widespread presence of this zone in wells in New Jersey, as well as in Maryland and Virginia, probably indicates that this again was a time of global sea-level rise. The sediments dip to the southeast.

Zone NP 6, 61.6–60.4 Ma.—During deposition of Zone NP 6, the depocenter in New Jersey appears to have shifted to the north, where the units indicate a significant increase in the sedimentation rate. At localities B–4 and B–5 (fig. 5.1), the sediments are 60 ft thick (these are the thickest deposits observed for any one zone in this study); they thin to the south to 10 ft at Butler Place (fig. 5.1) and are completely absent south of the Camden-Trenton band. This distribution could be explained by downward tectonic movement to the north or upward tectonic movement in southern New Jersey, or both, or an increase in sediments being poured into the area to the north.

Zone NP 7, 60.4–59.9 Ma.—Sediments representing Zone NP 7 are found only south of the Camden-Trenton band in updip areas, and the depocenter appears to have shifted again, this time to the south. There is no downdip information because none of the wells I have examined to date in this part of the State were drilled deeply enough to reach this zone. Sediments of this age are 5–10 ft thick. There appears to have been a reversal at this time of the conditions found during Zone NP 6 time, and the region north of the Camden-Trenton band was higher than the region south of the band.

Zone NP 8, 59.9–59.1 Ma.—Zone NP 8 thickens downdip toward the southeast from 5–10 ft thick at the GL cores to approximately 40 ft thick at Island Beach. The presence of this zone both north and south of Camden indicates a significant transgression, which may have masked the effects of any local basin warping. A corresponding transgression in Virginia and Maryland indicates that we are seeing the effects of a worldwide rise in sea level around 59 Ma.

Lower part of Zone NP 9, 59.1–57.8 Ma.—Zone NP 9 is missing north of the Camden-Trenton band. The thickness of the lower part of Zone 9 ranges from 10 to 30 ft in southern New Jersey, but there is no apparent downdip thickening. Local basin forces again dominated during Zone NP 9 time, and the northern part of the coastal plain apparently was high relative to the southern part.

Upper Paleocene Part of the Manasquan Formation, Upper Part of Zone NP 9

Upper part of Zone NP 9.—The upper part of Zone NP 9 occurs in the Manasquan Formation only in the GL wells (fig. 5.1). This part of the Manasquan has not been found anywhere else to date in New Jersey.

Eocene, Zone NP 10–Lower Part of Zone NP 21

The lower and middle Eocene deposits differ significantly from the Paleocene in New Jersey and appear to indicate more stable conditions. There are indications of small shifts in the depocenter during the Eocene, again in relation to the Camden-Trenton band, but they are of much smaller magnitude than those in the Paleocene. However, the late Eocene may have been a time of renewed tectonism in New Jersey.

Lower Eocene Part of the Manasquan Formation, Zones NP 10–14

Zone NP 10, 57.8–56.7 Ma.—Zone NP 10 is the oldest Eocene zone. The Paleocene-Eocene boundary occurs within the Manasquan, and, at the GL localities, deposition appears to have been continuous across this boundary. New Jersey may well have the most complete marine Paleocene-Eocene boundary section in the world. Zone NP 10 is present both north and south of the Camden-Trenton band. However, this zone is absent at Butler Place within the band. This distribution could be explained if the Camden-Trenton band was up slightly during deposition of Zone NP 10; there also is a possibility that this thin zone was not sampled at Butler Place. Additional data are needed from the Camden-Trenton band to show whether Zone NP 10 is present there. The sediments in Zone NP 10 thicken downdip to the southeast from 5–10 ft in the GL cores to 35 ft at Island Beach.

Zone NP 11, 56.7–55.3 Ma.—Zone NP 11 is missing from updip locations (GL cores, New Brooklyn Park, B–4, B–5, and Bog Creek, fig. 5.1). In fact, the updip locations have no marine Eocene deposits preserved from zones younger than Zone NP 10. This absence may represent less extensive transgressions, or, more likely, these sediments may have been eroded during a widespread Miocene transgression. The thickest Zone NP 11 sediments in the study area occur in the Camden-Trenton band (40 and 46 ft, respectively, for Butler Place and Island Beach, in contrast to 6 ft at ASP 2). The Camden-Trenton band may have been slightly lower than the northern and southern parts of the New Jersey Coastal Plain during deposition of Zone NP 11.

Zone NP 12, 55.3–53.7 Ma.—Zone NP 12 is thickest north and south of the Camden-Trenton band, which appears to have been uplifted after Zone NP 11 time. The sediments dip to the southeast.

Zones NP 13/14, 53.7–48.0 Ma.—Zones NP 14 through NP 16 each represent 3–4 m.y. Zones NP 13 and 14 could not be clearly delineated in many of the wells examined to date, and so they have been combined for this report. The thickness range from approximately 15–20 ft updip to approximately 50 ft downdip at ACGS–4 does not appear to indicate an increase in sedimentation rate. The contours show that sediments dip to the southeast; they thicken downdip and possibly to the south.

Shark River Formation, Zone NP 14–Middle Part of Zone NP 18 (Middle Eocene to Lower Upper Eocene), and Younger Zones

Although the upper part of Zone NP 14 at ACGS–4 is in the Shark River Formation, this zone is not clearly delineated in most of the New Jersey wells, as mentioned above, and the Manasquan and Shark River parts are not discussed separately.

Zones NP 15/16, 48.0–42.2 Ma.—Zones NP 15 and 16 could be differentiated in only a few wells in the study area, and so they are combined in this report. The lack of an obvious pattern in the thickness variations of these zones probably is a result of the erosion of each zone's upper part before deposition of the overlying unit. In more updip areas particularly, these sediments are overlain by Miocene sediments. The dip is to the southeast. We have evidence for an unconformity in Zone NP 15 and possibly in Zone NP 14 (Poore and Bybell, 1988).

Zone NP 17–lower part of Zone NP 21.—After Zone NP 16 time, the conditions in New Jersey changed again. For example, parts of Zones NP 17 through 21 are present only in the most downdip holes (ACGS–4, Island Beach, Atlantic City, Anchor Dickinson), and this distribution suggests some tectonism in New Jersey at this time. However, Zones NP 17–21 are currently too poorly represented in wells examined in the study area to be discussed in this report.

CONCLUSIONS

During the Paleocene through early middle Eocene, New Jersey had nearly continuous marine sedimentation. Although the sediments are thin and of nearly uniform thickness, detailed calcareous nannofossil data were used to (1) map variations in sediment thickness across the New Jersey Coastal Plain at any one time, as well as through time, (2) locate shifting of depocenters that may have been due to local tectonism, and (3) identify global eustatic sea-level changes. As additional wells are examined, data will be added to this New Jersey model and will be compared with data collected in Virginia and Maryland.

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6. Preliminary Ostracode Biostratigraphy of Subsurface Campanian and Maastrichtian Sections of the New Jersey Coastal Plain[1]

By Gregory S. Gohn[2]

INTRODUCTION

Ostracodes have rarely been considered in stratigraphic investigations of the Cretaceous section of the New Jersey Coastal Plain. Although ostracodes have proven to be of considerable stratigraphic utility in studying Cretaceous sections elsewhere in the Atlantic and Gulf of Mexico Coastal Plains (Alexander, 1929; Brown, 1957; Hazel and Paulson, 1964; Crane, 1965; Brouwers and Hazel, 1978; Hazel and Brouwers, 1982), only the early paper by Jennings (1936) and the unpublished thesis by Nine (1954) comprehensively treat Cretaceous ostracode faunas of the New Jersey Coastal Plain. Probably the most extensive stratigraphic use of ostracodes in New Jersey was by Brown and others (1972), who used a relatively few marker species to identify their regional chronostratigraphic units. Brouwers and Hazel (1978) gave partial lists of species from four outcrop samples from northern New Jersey as part of their study of Maastrichtian ostracodes in Maryland. Adams (1960) listed 26 species from the Red Bank Formation of New Jersey.

This report is intended to illustrate that stratigraphically important Campanian and Maastrichtian ostracode faunas are common in New Jersey and that they hold considerable promise for biostratigraphic zonation of that section. A short discussion of the relationships between subsurface lithostratigraphic units used in this study and the well-known outcrop units of the Campanian and Maastrichtian Stages in New Jersey also is included.

For this report, ostracode faunas were studied from Campanian and Maastrichtian sections in four coreholes in the New Jersey Coastal Plain (fig. 6.1). Two of these coreholes, USGS-Clayton 1 in Gloucester County and USGS-Freehold 1 in Monmouth County, were drilled in recent years by the U.S. Geological Survey (USGS) in cooperation with the New Jersey Geological Survey. Both of these holes were continuously cored. Prepared ostracode slides (5- to 10-ft sample interval) also were available from a 150-ft-thick cored interval at the base of the Ocean County Water Department corehole 6, drilled at Mantoloking in the 1950's. In addition, a few short cores were available from the Howell Township corehole 5, southeastern Monmouth County, of which three were productive. In all, 73 ostracode slides were examined from the four coreholes. Although not reported herein, about 30 slides were also examined from other localities including the Woodstown,

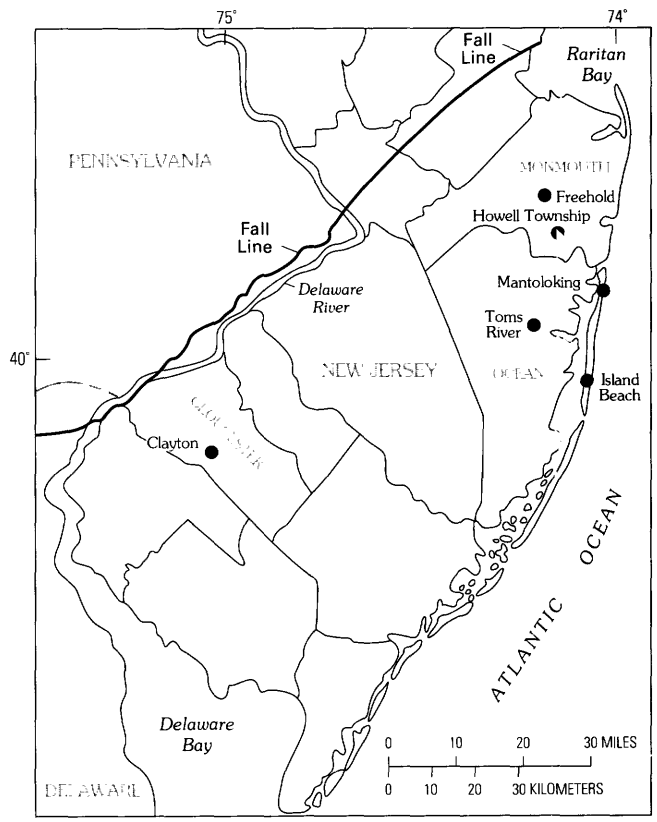

Figure 6.1. County boundaries on the New Jersey Coastal Plain; Monmouth, Ocean, and Gloucester Counties are labeled. Locations of drill holes mentioned in this report are also shown.

[1]Prepared in cooperation with the New Jersey Geological Survey.
[2]U.S. Geological Survey, Reston, VA 22092.

Outcrop Section: Traditional nomenclature and sediment cycles of Owens and Sohl (1969)		Nomenclature of subsurface cycles: this report
Tinton Sand	6	S6
Red Bank Formation	5	S5
Navesink Formation		
Mount Laurel Sand		
Wenonah Formation	4	S4
Marshalltown Formation		
Englishtown Formation		S3
		?
Woodbury Clay	3	S2
		?
Merchantville Formation		S1

Figure 6.2. Correlation chart for the Campanian and Maastrichtian cycles of the New Jersey Coastal Plain.

Berlin, Mantua, Clayton, Island Beach, and Poricy Brook areas in New Jersey and the Chesapeake and Delaware Canal in Delaware.

Acknowledgments

I am indebted to Joseph E. Hazel (Louisiana State University) for his patience and endurance during the initial stages of my ostracode studies. Samples from the Mantoloking and USGS-Island Beach drill holes were originally prepared and examined by J.E. Hazel. The manuscript benefited greatly from reviews by N.F. Sohl (USGS) and J.E. Hazel.

LITHOSTRATIGRAPHY

Stratigraphic nomenclature for the outcropping Campanian and Maastrichtian sections of the New Jersey Coastal Plain has remained almost unchanged since publication of the stratigraphic report by Owens and Sohl (1969). In their report and in nearly all subsequent reports, nine formations are recognized between the Santonian Magothy Formation and the Tertiary section (fig. 6.2; see Olsson, 1964, and Olsson and others, 1988, for an alternative stratigraphy for the Navesink and Red Bank Formations). Owens and Sohl (1969) also recognized the strongly cyclic nature of the New Jersey Cretaceous section and assigned the Campanian and Maastrichtian formations to their cycles 3, 4, 5, and 6 (fig. 6.2). Each cycle consists of a basal glauconite sand (Merchantville, Marshalltown, and Navesink Formations), a middle clay-silt unit (Woodbury Clay, Wenonah Formation, and lower part of the Red Bank Formation), and an upper quartz sand (Englishtown Formation, Mount Laurel Sand, and upper part of the Red Bank Formation). The Tinton Sand is a lithologically and areally restricted cycle that may be erosionally truncated. The cycle boundaries are sharp and disconformable, whereas formation contacts within the cycles are gradational.

The formation nomenclature used for the outcrop section is also used in the subsurface of New Jersey (for example, see Petters, 1976, 1977; Zapecza, 1989, this volume). This usage implies that the four Campanian–Maastrichtian cycles of Owens and Sohl (1969) are also present in the subsurface. However, the drill-hole sections studied for this report suggest a more complex stratigraphy.

The thickest Campanian–Maastrichtian sections in New Jersey occur along the coast in southeastern Monmouth and eastern Ocean Counties (Brown and others, 1972; Perry and others, 1975; Zapecza, 1989, this volume). These sections thin from this area northwestward toward the Fall Line and also southwestward along strike toward the Delmarva Peninsula (Owens and others, 1977; Olsson and others, 1988). Much of the southwestward thinning is accomplished through the thinning or loss of the quartz-sand units at the tops of cycles 3 and 5 of Owens and Sohl.

Data from the deep drill holes in the Ocean County-Monmouth County area of thick Campanian–Maastrichtian sediments indicate that five sedimentary cycles (fig. 6.2, cycles S1 to S5) are present in that area. The Tinton Sand is absent (eroded?) from most of this area although it is present in the subsurface further updip in the Freehold corehole, where it represents cycle S6. In contrast, cycles S4 and S5 are readily correlated from the Howell Township corehole (fig. 6.3) and the frequently studied Toms River Chemical Company well (Perry and others, 1975; Petters, 1976; Zapecza, 1989, this volume) northward to the updip Freehold corehole and the outcrop belt, where they correlate with outcrop cycles 4 and 5.

The thick sand of cycle S3 in the Howell Township corehole (fig. 6.3, 525–410 ft) is also readily correlated with a similar section in the Freehold corehole, and this unit is equivalent to the outcropping Englishtown Formation of the northern New Jersey Coastal Plain. However, the thinner sands at the tops of cycles S1 and S2 at Howell Township are not present updip at Freehold, where S1 and S2 consist only of micaceous clay-silts and glauconite sands. Apparently, the upper parts of cycles S1 and S2 are truncated in the updip direction. As a result, a marine section that consists of the amalgamated fine-grained parts of cycles S1 and S2 and the lower part of S3 in the Freehold corehole probably is equivalent to the Merchantville Formation and Woodbury Clay of the outcrop belt.

Similar trends occur to the southwest, where the Clayton corehole (fig. 6.1) and nearby wells in southern New Jersey lack sands correlative with the sands at the tops of cycles S1, S2, and S3 in the Toms River, Howell Township, and other nearby drill holes. In southern New

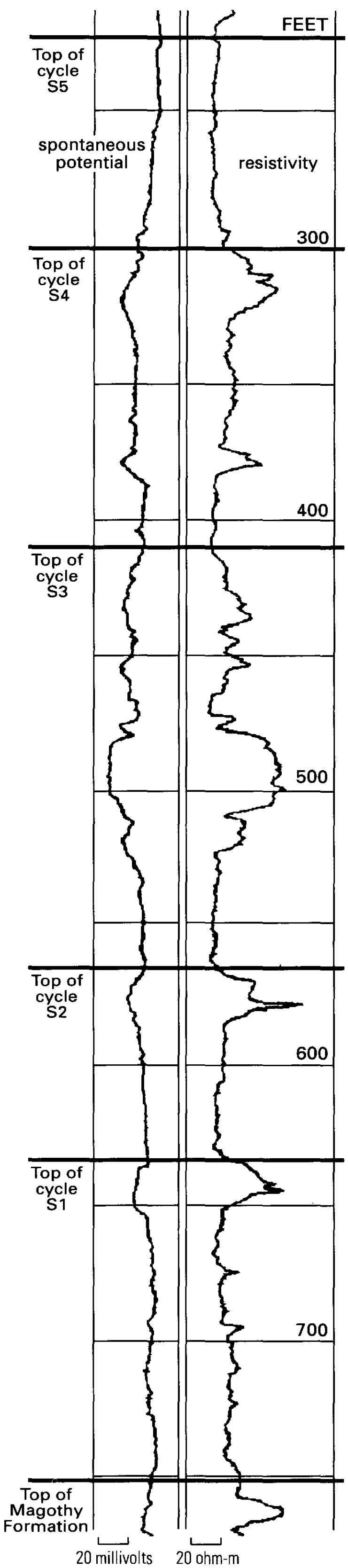

Figure 6.3. Electric log for the Campanian and Maastrichtian sections in the Howell Township corehole 5. Boundaries for the subsurface cycles of this report are indicated.

Jersey, the amalgamated fine-grained marine sections of cycles S1, S2, and S3 are also assigned to the Merchantville and Woodbury, as well as to an atypically fine-grained Englishtown Formation, in the outcrop (Minard, 1965) and the subsurface (Zapecza, 1989, this volume).

OSTRACODE BIOSTRATIGRAPHY

Species selected for inclusion in this report were restricted to forms having relatively simple (or complex but stable) synonymies that have been discussed and (or) illustrated in recent years. Most of these species have been re-illustrated by Hazel and Brouwers (1982). The species list was further restricted to include primarily species having well-established ranges and known stratigraphic utility, principally species used in the zonation of Hazel and Brouwers (1982). Many of the included species were defined or discussed in papers by Hazel and Paulson (1964) and Brouwers and Hazel (1978), and concepts of most of these species have changed little since publication of those reports. My concepts of the species included in this report are drawn in large part from these three articles. Two species, *"Cythereis"* n. sp. 1 (=*Cythereis magnifica* of Nine, 1954) and *Planileberis*? *pulchra*, described only by Jennings (1936) and Nine (1954, unpublished thesis), also are included, as is one distinctive undescribed species (*Veenia*? n. sp. 1, which is a large, thick-valved species that may be assignable to *Veenia* Butler and Jones, 1957). The stratigraphic positions of the studied sections in the four cores are shown in figure 6.4.

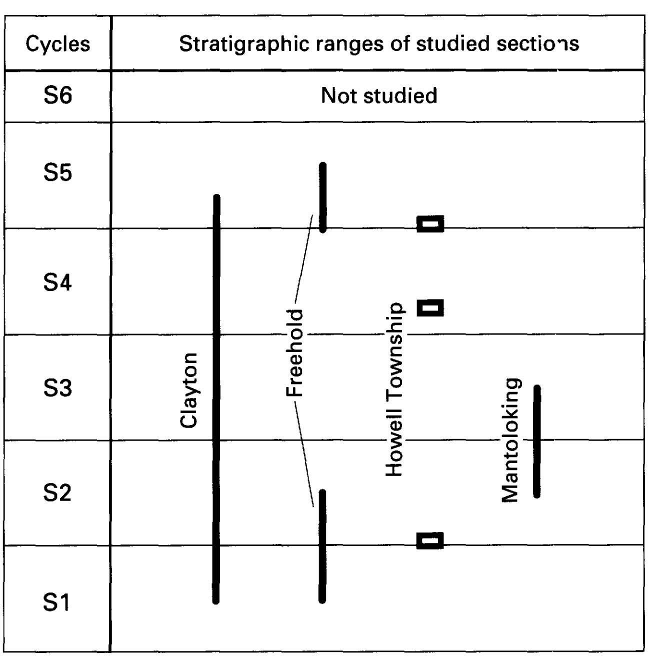

Figure 6.4. Stratigraphic positions of the intervals studied from the four coreholes used in this report.

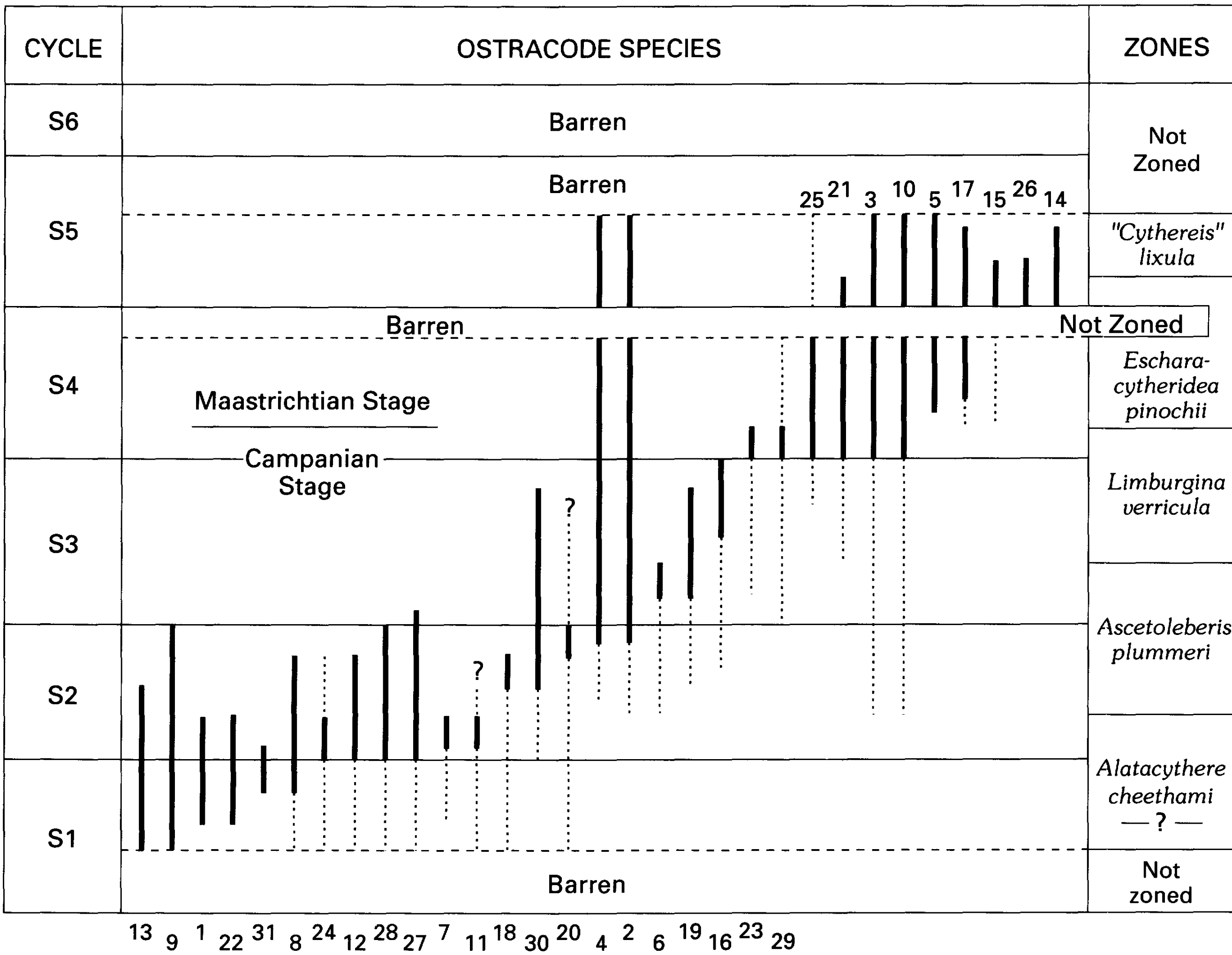

Figure 6.5. Distribution of selected ostracode species in the subsurface stratigraphic cycles used in this report. Identifying numbers for individual species are from table 6.1. The ranges shown on the figure are composites of the ranges seen in the four studied coreholes. Ranges of species in New Jersey are shown by heavy lines; dotted lines are extended ranges of species given by Hazel and Brouwers (1982). Zones are the ostracode interval zones of Hazel and Brouwers (1982).

The ranges shown for the species found in the New Jersey cores (fig. 6.5, table 6.1) are augmented by the ranges given for those species by Hazel and Brouwers (1982) in their provincewide study of Coniacian through Maastrichtian ostracodes of the Atlantic and Gulf of Mexico Coastal Plains. Many New Jersey ranges are shorter than those given by Hazel and Brouwers. These truncated ranges may simply be an artifact of the relatively modest number of samples used in this report. However, they may also result, in part, from the cyclic nature of the studied sedimentary section. Ostracodes are predominantly a benthic group whose distribution is controlled by a variety of ecologic factors. It is likely, therefore, that the factors that produced the cyclic lithofacies changes, such as water depth, sedimentation rate, energy regime, and sediment type, also produced shifts in the distribution of biofacies. Certain species may, therefore, be absent from a particular formation or facies for paleoenvironmental reasons. Finally, the shorter New Jersey ranges could result from the erosional truncation of cycles discussed above. In this case, parts of species ranges could be encompassed within the disconformities between cycles.

Cycle S1

Ostracodes are relatively sparse in S1, and all the species found in this unit also occur in cycle S2 (fig. 6.5). *Alatacythere cheethami* is the stratigraphically most important species in unit S1. Its total range in the lower Campanian (upper Austinian) sections of the western Gulf of Mexico Coastal Plain is nearly coincident with the *Alatacythere cheethami* Interval Zone of Hazel and Brouw-

Table 6.1. Selected ostracode species from the studied New Jersey cores
[Numbers are used in figure 6.5]

1. *Alatacythere cheethami* (Hazel and Paulson, 1964)
2. *Antibythocypris fabaformis* (Berry, 1925)
3. *Antibythocypris gooberi* Jennings, 1936
4. *Antibythocypris minuta* (Berry, 1925)
5. *Anticythereis reticulata* (Jennings, 1936)
6. *Ascetoleberis rugosissima* (Alexander, 1929)
7. *Brachycythere acuminata* Hazel and Paulson, 1964
8. *Brachycythere crenulata* Crane, 1965
9. *Brachycythere pyriforma* Hazel and Paulson, 1964
10. *Curfsina communis* (Israelsky, 1929)
11. "*Cythereis*" *bicornis* Israelsky, 1929
12. "*Cythereis*" *hannai* Israelsky, 1929
13. "*Cythereis*" *veclitella* Crane, 1965
14. "*Cythereis*" n. sp. 1
15. *Escharacytheridea pinochii* (Jennings, 1936)
16. *Fissocarinocythere gapensis* (Alexander, 1929)
17. *Fissocarinocythere huntensis* (Alexander, 1929)
18. *Fissocarinocythere pittensis* (Swain and Brown, 1964)
19. *Haplocytheridea insolita* (Alexander and Alexander, 1933)
20. *Haplocytheridea plummeri* (Alexander, 1929)
21. *Limburgina verricula* (Butler and Jones, 1957)
22. *Mosaleberis*? *reesidei* (Swain, 1948)
23. *Phacorhabdotus* cf. *P. texanus* Howe and Laurencich, 1958
24. *Physocythere annulospinata* (Hazel and Paulson, 1964)
25. *Planileberis*? *costatana* (Israelsky, 1929)
26. *Planileberis*? *pulchra* (Jennings, 1936)
27. *Schizoptocythere*? *compressa* (Hazel and Paulson, 1964)
28. *Schuleridea parvasulcata* (Swain, 1948)
29. *Veenia ponderosana* (Israelsky, 1929), late form of Hazel and Brouwers, 1982
30. *Veenia spoori* (Israelsky, 1929)
31. *Veenia*? n. sp. 1

ers (1982). The lowest occurrence of this species in the western Gulf Coast sections is in the short interval between the first occurrence of *Globotruncanita elevata* and the last occurrence of *Dicarinella concavata* (Hazel and Brouwers, 1982; Bryant, 1984; J.E. Hazel, written commun., 1988). Therefore, the lowest occurrence of *A. cheethami* is just slightly higher than the Santonian-Campanian Stage boundary as defined by the first appearance of *G. elevata*. Because the lower part of S1 is barren of ostracodes, the lowest occurrence of *A. cheethami* is not sufficient to indicate whether that unit is of Santonian and Campanian, or just Campanian, age. Petters (1976, 1977) and Olsson and Sikora (1989) previously have recognized marine Santonian beds (Merchantville lithology) above the Magothy Formation in New Jersey. The Santonian age was based, however, on the presence of *D. concavata*, part of whose range is placed in the Campanian by some authors. It is likely that cycle S1 of this report is largely equivalent to the Santonian sequence of Olsson and Sikora (1989).

Cycle S2

Ostracodes are relatively abundant in cycle S2. *Alatacythere cheethami* occurs in the lower part of S2 along with *Brachycythere acuminata* and numerous other species including *Brachycythere pyriforma* and "*Cythereis*" *bicornis*. On the basis of the observed species, the lower part of S2 is assignable to the *Alatacythere cheethami* Interval Zone of Hazel and Brouwers (1982). The upper part of S2, which contains the highest occurrences of the genus *Schuleridea*, is assignable to the lower part of the overlying *Ascetoleberis plummeri* Interval Zone of those authors. These zonal assignments indicate an early Campanian age for S2 and probable equivalence to the lower Campanian sequence of Olsson and Sikora (1989). In addition to its occurrence in the Clayton, Freehold, and Howell Township coreholes, *Alatacythere cheethami* also occurs at 1,616 ft in the USGS-Island Beach drill hole.

Cycle S3

Stratigraphically, *Fissocarinocythere gapensis*, *Haplocytheridea insolita*, and *Ascetoleberis rugosissima* are the most important species in S3, which contains locally sparse to common ostracodes. To date, *A. rugosissima* is known only from the *Ascetoleberis plummeri* Interval Zone of Hazel and Brouwers (1982), whereas the highest occurrences of *F. gapensis* and *H. insolita* are in the overlying *Limburgina verricula* Interval Zone of those authors. It is likely that both zones are represented in cycle S3 and that this unit is of middle to late Campanian age. Cycle S3 probably equates with the middle Campanian sequence of Olsson and Sikora (1989).

Cycle S4

Cycle S4 is characterized by the presence of *Planileberis*? *costatana* with *Limburgina verricula*, *Fissocarinocythere huntensis*, and *Anticythereis reticulata*. Although Hazel and Brouwers (1982) showed a much longer range for *P.*? *costatana*, I have not observed it outside of cycle S4 in New Jersey. The presence of specimens that may be *Phacorhabdotus texanus* suggests that the lower part of S4 (Marshalltown Formation) may be of latest Campanian age (Hazel and Brouwers, 1982); this age is compatible with the presence of *Globotruncanita calcarata* in the Marshalltown as reported by Petters (1977) and Olsson and Sikora (1989). Slightly higher in S4, the lowest occurrences of *F. huntensis* and *A. reticulata* approximate the Campanian-Maastrichtian Stage boundary. The presence of *F. huntensis* in cycle S4, and the presence of *L. verricula* with *Escharacytheridea pinochii* in the lower part of cycle S5 (see next section) require assignment of most of cycle S4 to the *Escharacytheridea pinochii* Interval Zone of Hazel and Brouwers (1982). Coarse quartz sands of the upper part of the Mount Laurel Sand are barren of ostracodes although clayey, finer grained sands of the lower part of the formation contain abundant specimens. Ostracodes are common in the Wenonah Formation and sparse to common in the Marshalltown.

Cycle S5

Ostracodes are abundant in the Navesink Formation and the lower part of the Red Bank Formation but are absent from the sands of the upper part of the Red Bank (and from the Tinton Sand). Characteristic species of cycle S5 include *"Cythereis"* n. sp. 1 (=*Cythereis magnifica* of Nine, 1954) in the Navesink and Red Bank and *Planileberis*? *pulchra*, which is restricted to the Navesink. *Limburgina verricula* last appears within the lower part of the Navesink.

The presence of *Limburgina verricula* with *Escharacytheridea pinochii* in the lower part of the Navesink places this section within the *Escharacytheridea pinochii* Interval Zone of Hazel and Brouwers (1982). Cretaceous sediments above the last appearance of *L. verricula* are necessarily within the chronozone of the "*Cythereis*" *lixula* Interval Zone of Hazel and Brouwers (1982). The lowest occurrence of *Planileberis*? *pulchra* and "*Cythereis*" n. sp. 1 in the lower part of the Navesink serve as local markers for the cycle S4-cycle S5 (Mount Laurel-Navesink) boundary in New Jersey.

SUMMARY

In summary, it is apparent that ostracode faunas can be used successfully to provide local correlations within the New Jersey Campanian and Maastrichtian sections. These faunas can be related to the regional zonation of Hazel and Brouwers (1982), and they can be used to relate the chronozones of certain planktic Foraminifera to wells in which planktic faunas are unstudied or lacking. Several ostracode datums, among others, that appear to be of stratigraphic utility include the following:

1. The highest occurrence of *Planileberis*? *pulchra* near the Navesink-Red Bank contact (within cycle S5).
2. The highest occurrence of *Limburgina verricula* in the lower part of the Navesink Formation (cycle S5).
3. The restriction in New Jersey of *Planileberis*? *costatana* to cycle S4.
4. The lowest occurrence of *Fissocarinocythere huntensis* and *Anticythereis reticulata* near the Campanian-Maastrichtian boundary (within cycle S4).
5. The highest occurrence of *Fissocarinocythere gapensis* and *Haplocytheridea insolita* near the top of cycle S3.
6. The highest occurrence of *Brachycythere pyriforma* and species of *Schuleridea* at the top of cycle S2.
7. The highest occurrence of *Alatacythere cheethami* in the lower part of cycle S2.

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7. Paleohydrology of Four Watersheds in the New Jersey Coastal Plain[1]

By Wayne L. Newell[2] *and* John S. Wyckoff[2]

INTRODUCTION

We studied four watersheds of the southern New Jersey Coastal Plain (fig. 7.1): the Maurice, Great Egg Harbor, Mullica, and Rancocas watersheds. The Maurice River enters Delaware Bay after flowing south across the dip slope of surficial and bedrock deposits, which include the Kirkwood Formation (early Miocene), Cohansey Sand (middle Miocene), and Bridgeton Formation (late Miocene). The Great Egg Harbor and Mullica Rivers flow southeastward into the Atlantic Ocean. Rancocas Creek and its branches flow westward into the Delaware River; they are entrenched into the Cohansey and Kirkwood and also drain across successively older formations of early Tertiary and Late Cretaceous age. All four watersheds contain thick extensive surficial deposits that cannot be produced under present-day hydrologic conditions. In this report, we describe the present hydrogeology of the southern New Jersey Coastal Plain and evidence for former hydrologic conditions that were capable of eroding the landscape and producing a cover of surficial deposits.

PHYSICAL GEOGRAPHY

The uplands of the southern New Jersey Coastal Plain form a plateau that dips gently seaward; the altitude ranges from more than 200 ft on the hills in the west in and around Berlin to 30 ft on the plains in the southeast around Woodbine. Broad drainage basins within the area descend to the Atlantic Ocean and Delaware Bay. The valley bottoms commonly are 30–100 ft below the upland surfaces, are 1–2 mi wide, and have gently sloping, swampy surfaces.

The upland surfaces are gently rolling. Locally subdued linear ridges are the surface expression of gravel bars within the underlying Bridgeton Formation. Along the western margin of the uplands, windblown sand and silt from the nearby Delaware Valley cover topographic irregularities. Farms and hardwood forests exist where the upland soils retain water.

The uplands south of the Mullica River are underlain by the Bridgeton Formation, which is deeply weathered, arkosic sand and gravel. The uplands are generally well drained except where locally discontinuous clay beds cause a perched water table. Thick stands of white cedar and swamp maple grow in the poorly drained areas. The Bridgeton Formation ranges from 10 to 65 ft in thickness.

Underlying the Bridgeton Formation is the Cohansey Sand, which is typically a well-sorted, orthoquartzitic, marine-shelf sand. Locally, the Cohansey includes a full range of nearshore marine and deltaic facies. Although these facies include some laterally continuous clay beds, the

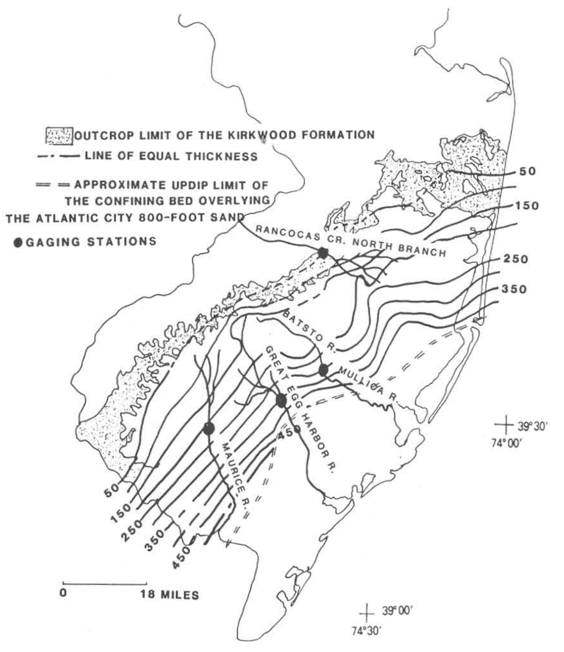

Figure 7.1. Locations of the four studied watersheds plotted on an isopach map of the Kirkwood-Cohansey aquifer system in New Jersey. Isopachs in feet. Modified from Zapecza (1989).

[1]Prepared in cooperation with the New Jersey Geological Survey.
[2]U.S. Geological Survey, Reston, VA 22092.

Cohansey and the surficial deposits and soils overlying it where the Bridgeton is not present are generally excessively well drained. Vegetation on these deposits is characterized by scrub pines and scattered oak trees.

The Kirkwood Formation underlies the Cohansey and crops out along the western slopes of the scarp separating the uplands from the Delaware River valley bottom (fig. 7.1). In outcrop, it is a very fine, well-sorted, micaceous, quartzitic, marine-shelf sand. In the subsurface, changes in facies occur, and the Kirkwood becomes progressively darker and finer grained, and it is locally fossiliferous.

Most of the entrenched Holocene fluvial systems draining the uplands dissect the Bridgeton Formation and flow along broad valley bottoms floored with the Cohansey Sand. Surficial deposits in the valley bottoms are a combination of reworked Cohansey and Bridgeton sediments. Locally, blankets of windblown Cohansey Sand occur in the valley bottoms and on the downwind valley slopes. These deposits have the same range of permeability characteristics as their parent materials.

HYDROLOGIC CHARACTER OF FOUR NEW JERSEY COASTAL PLAIN WATERSHEDS

The present southern New Jersey climate is humid temperate and is moderated by Delaware Bay and the Atlantic Ocean. The average annual precipitation is 40–44 in., and temperature extremes are 5 and 100 °F (Powley, 1978).

Four gauged rivers that drain the surficial and bedrock deposits of the New Jersey Coastal Plain were chosen to characterize the hydrologic response of their watersheds to present climatic conditions. These watersheds are the North Fork of Rancocas Creek, the Batsto River (tributary to the Mullica), the Great Egg Harbor River, and the Maurice River (fig. 7.1). The basins of Rancocas Creek and the Mullica share a common divide. Rancocas Creek flows westward across the strike of the bedrock formations into the Delaware River, whereas the Mullica River flows southeastward down the dip slope of the Cohansey Sand into the Atlantic. The Great Egg Harbor River is also a dip-slope-draining river discharging to the Atlantic. The Maurice River flows obliquely down the dip of the Cohansey and into Delaware Bay.

Continuous discharge records (U.S. Geological Survey, 1979–86) from the four watersheds for 7 yr, from 1979 to 1986, were examined, and their base flows and maximum flood peaks were compared. The normalized base flows for the Maurice River and Great Egg Harbor River are each just less than 1 cubic foot per second per square mile ($(ft^3/s)/mi^2$). The Batsto River's base flow is about 1.3 $(ft^3/s)/mi^2$, and the Rancocas basin contributes about 0.75 $(ft^3/s)/mi^2$ to base flow. The large base flow for the Batsto River reflects the contribution of the southeast-dipping surficial aquifer, which produces discharge collected not only from the Batsto watershed but also poached from the headwaters of the Rancocas watershed. The surficial aquifer is the zone containing permeable sediments at or near the surface that contribute discharge to the surface hydrology (that is, lakes, streams, and ponds) or receive recharge from it. Base flow for each of the rivers is similar through the seasons, and this uniformity indicates major continuous ground-water discharge into the surface-water system.

The largest storm during the recording period generated a discharge less than 1,310 ft^3/s for the Batsto River. Examination of slopes, flood plains, and channel sediments indicates that sediment stripping, transportation, and deposition are minimal. During the most extreme precipitation events, when inches fell per hour, there was no surface runoff. All the water infiltrated into the surface aquifer. Stream discharge levels rose as a combination of (1) direct input into the channel and adjacent swamps and (2) rising

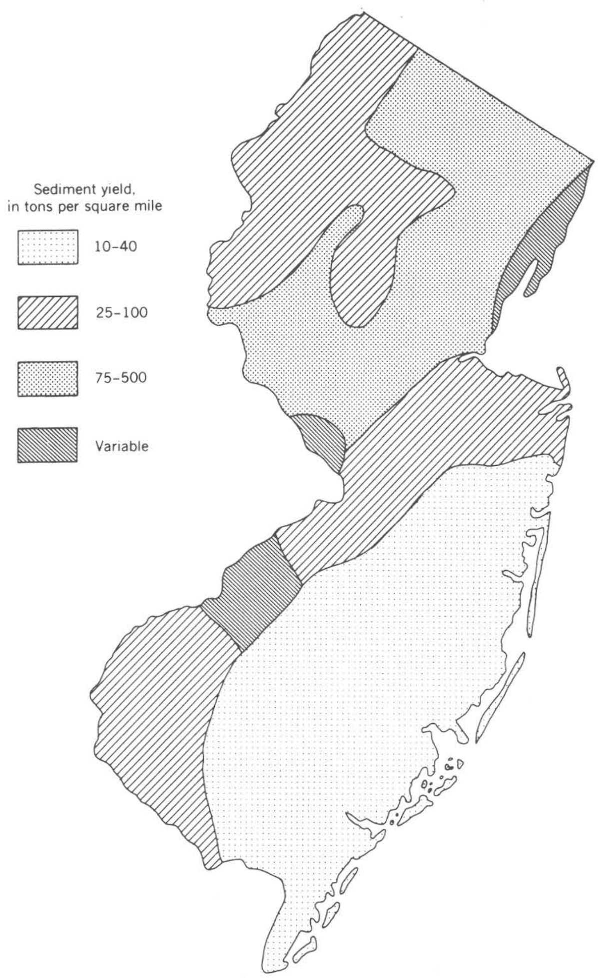

Figure 7.2. Sediment yield in New Jersey. Modified from U.S. Geological Survey (1968).

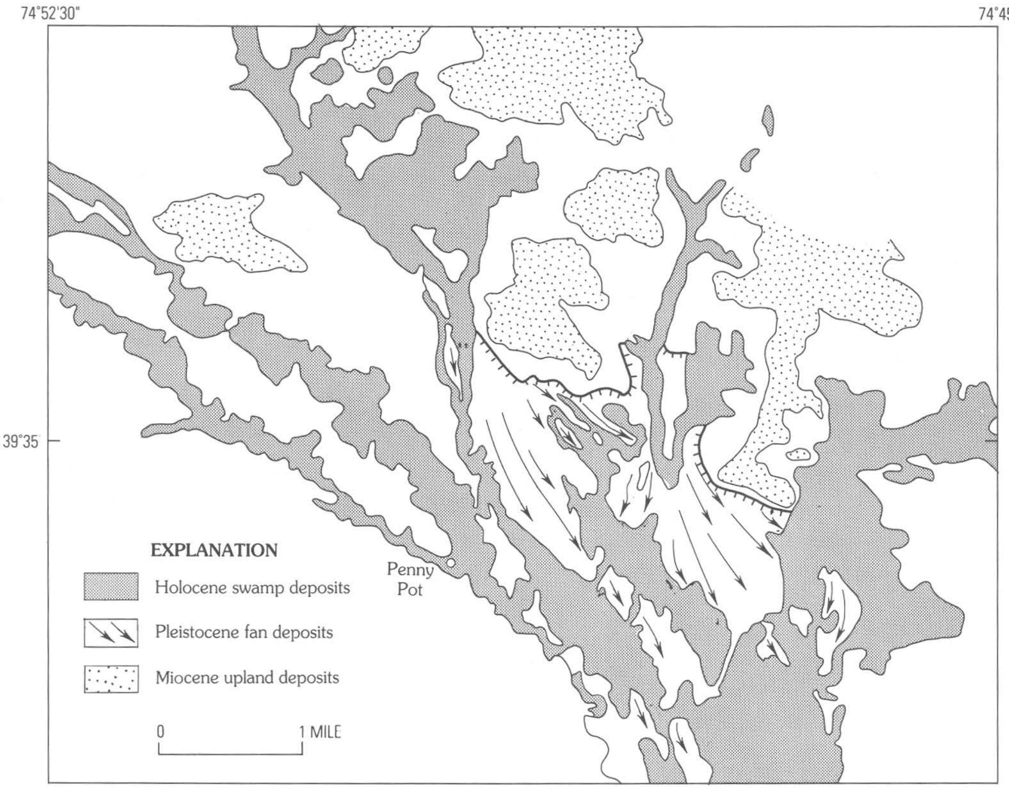

Figure 7.3. Surficial deposits and geomorphology in the Great Egg Harbor River valley in part of the Newtonville quadrangle, New Jersey. Large alluvial fans grade from an upland underlain by the Bridgeton Formation indicated by the scarp shown by the hachured line. Directions of sediment transport are shown by arrows.

ground-water discharge. Sediment yield from the watersheds studied is probably less than 10 tons/mi^2, and the yield for the whole region averages 10–40 tons/mi^2 (fig. 7.2). Such small sediment yields suggest that present fluvial processes are ineffective in modifying the southern New Jersey landscape.

GEOMORPHOLOGY AND SURFICIAL GEOLOGY

The broad bottomlands of the Great Egg Harbor, Maurice, and Mullica River valleys are bordered by low, gently sloping terraces that are parallel to the valley bottoms. Locally, they merge almost imperceptibly with very low gradient valley side slopes. Exposures indicate that the low terraces are underlain by a layer of alluvium 3–6 ft thick. The side slopes are covered with an equally thin but extensive cover of colluvium. A thin layer of windblown sand has a variable thickness and commonly obscures the transition from slope deposits to valley bottom deposits. Where exposed, the terrace deposits commonly are thin-bedded sheets of very fine to coarse sand containing sparse gravel. Stone lines mark the unconformable contacts. Holocene channel and flood-plain alluvium, by comparison, is a complex of point-bar deposits and trough cross-beds, which have abundant plant fragments and local deposits of woody peat. The thin colluvium is separated from underlying materials by a stone line that truncates older bedforms and weathering profiles. The material above the unconformity is homogenized and massive. Typically, weathered, rounded quartz and chert pebbles and ironstone clasts are observed randomly floating in a sandy matrix. Sparse ventifacts are present in the stone line and the colluvium. Young, thin, chromic soil profiles are common in such deposits.

The windblown sand consists of very well rounded, frosted or polished fine to medium grains of quartz and heavy minerals in structureless sheets. Locally, low dunes (6–9 ft high) enclose depressions. Ventifact pavements present at the bases of some dunes indicate a period of deflation prior to deposition of the overlying blanket of windblown sand.

Extensive debris fans debouch from tributaries into the main valleys, especially in the middle reaches of the Great Egg Harbor River (fig. 7.3). The fans merge with the

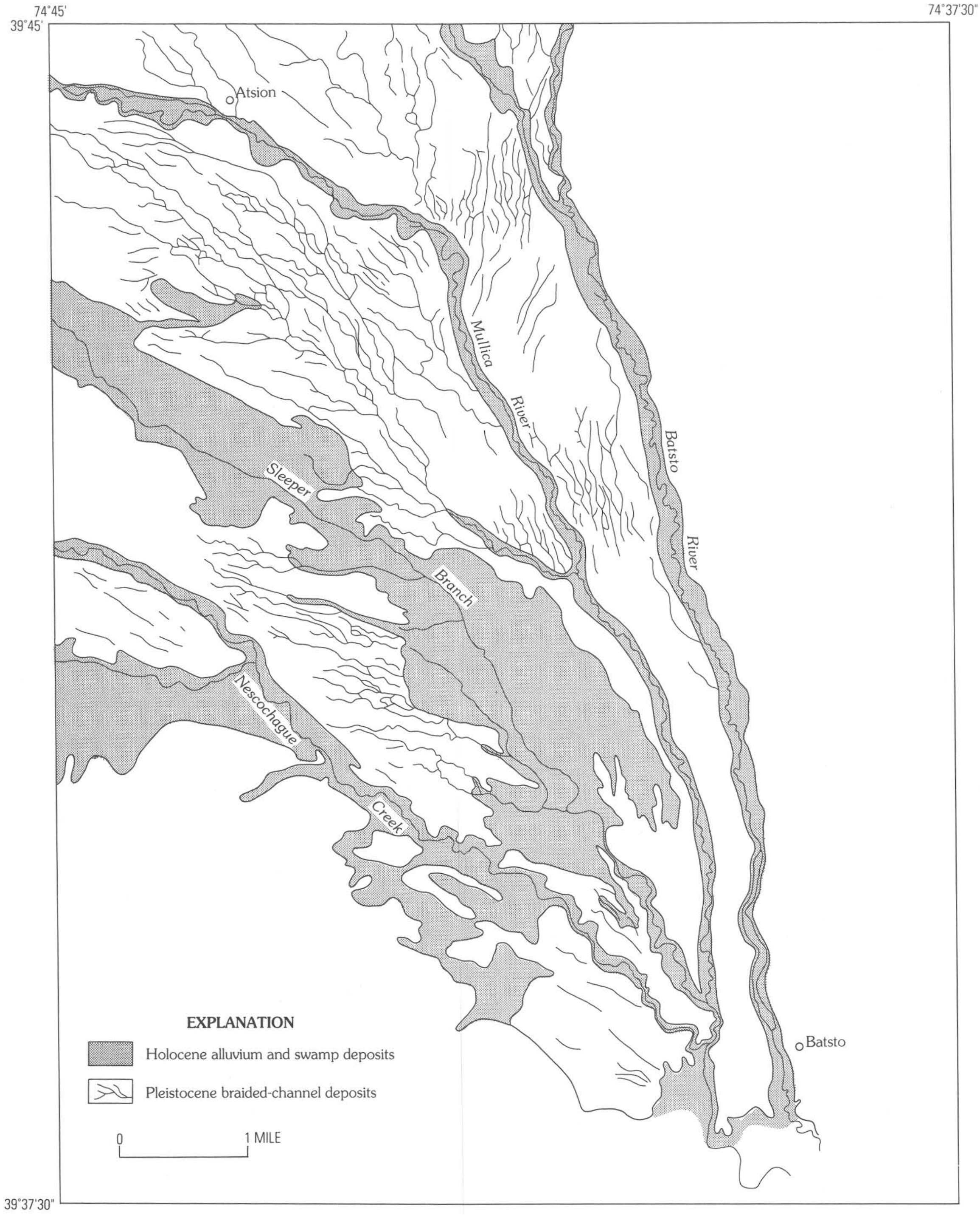

Figure 7.4. Surficial deposits and geomorphology in the Atsion quadrangle, New Jersey. Holocene drainage is incised into the braided-stream fabric that resulted from permafrost meltout cycles.

alluvial terraces at some localities and include the same extensive veneer of thin-bedded deposits.

The Mullica River arises in a broad plain that is also drained by the Batsto River, Sleeper Branch, and Nescochague Creek. Each of these streams is entrenched 6–9 ft below the plain and has a small, swampy flood plain and tightly meandering narrow channel that is scaled to present discharge from each subbasin. The intervening low, sloping

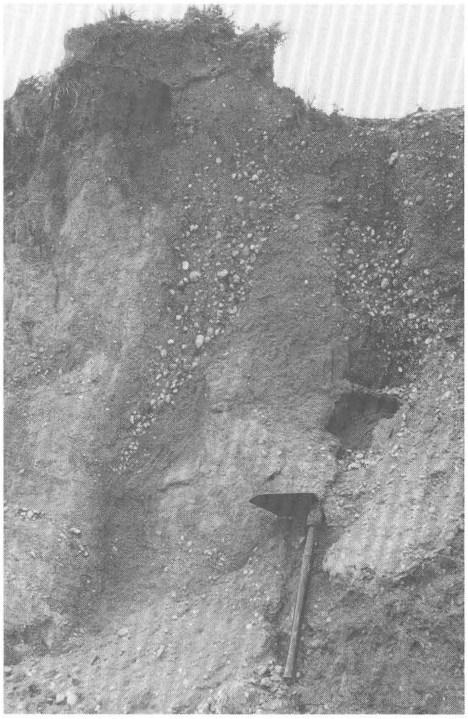

Figure 7.5. Frost wedge within the Bridgeton Formation near Hammonton, N.J.. Note the vertical alignment of the coarse clasts and the chaotic fabric within the wedge and the color transition between the adjacent sediments and the wedge. A thin layer of congeliturbate can be seen at the top of the profile; it is homogenized, poorly sorted, coarse to fine sand containing scattered pebbles and floating fine gravel. Shovel is 2.0 ft long.

plains that parallel the dip slope of the Cohansey Sand are etched with an intricate pattern of braided, small-scale stream channels (fig. 7.4). Each channel is about 10 ft wide and 3 ft deep. Excavations during dry spells indicate that the low sloping plain and channel complex is underlain by complex sets of festoon crossbeds (Farrell, 1983). The deposits themselves are entirely indigenous to the Cohansey Sand within the watershed; no distinctive Delaware Valley provenance clasts or sediments (notably, glauconite sands) have been found. This lack of exotic material indicates that there was no great debouching of the Delaware River across the New Jersey Coastal Plain uplands to deposit sediments along the outer edge of the Coastal Plain of New Jersey.

PALEOCLIMATE—SIGNATURE OF FROZEN GROUND

In the shallow subsurface across the uplands, on low terraces, and on marine terraces, almost all exposures reveal evidence of a frozen landscape (figs. 7.5 and 7.6). This

Figure 7.6. An involution in the Bridgeton Formation near Hammonton, N.J., displaying characteristic alignment of clasts indicative of deformation and reorientation of clasts.

evidence includes intersecting patterns of fossil frost wedges, involutions, congeliturbated layers containing vertically aligned clasts, closed depressions underlain by involutions, and debris flows overlying formerly frozen sand dunes. Throughout much of the area, topmost horizons of old, deeply weathered soils have been destroyed by congeliturbation in the active layer (the top 3–6 ft). At some localities, frost-wedge structures and weathering profiles suggest multiple periods of frost activity separated by periods of weathering and erosion without freeze-thaw events. A general picture emerges that during maximum glacial climatic events, southern New Jersey was a high plateau 300 ft or more above the sea level at that time; the shoreline then was approximately 100 mi east of the present shoreline. These conditions prevailed as far south as the latitude of Washington, D.C. (E.J. Ciolkosz, Pennsylvania State University, oral commun., Nov. 1987).

PALEOHYDROLOGY

During glacial events, the combination of frozen surface horizons and steeper channel gradients out onto the exposed continental shelf would have caused rates and volumes of water movement through the fluvial system to be drastically different from present conditions. If modern permafrost environments are a guide, then melting of the top 3 ft during summer would have provided a saturated active layer capable of large-scale sediment movement on very low gradient slopes. Hydrographs of streams draining thawed Arctic Coastal Plain watersheds indicated that summer rainstorms can provide an overprint of catastrophic erosional and depositional events (Brown and others, 1968). The impermeable boundary of frozen sediments restricts erosion and sediment transport to the thawed, thin surficial layer. Conversely, recharge of the surface aquifer is minimized. Under such conditions, overland flow and the entrainment of sediment in the fluvial system was possible in sediments that presently constitute a super-permeable medium. Such erosion and deposition, coupled with widespread downslope creep and debris flows, resulted in the surficial deposits that exist today.

The broad plain where the Mullica River arises is the one locality where the surficial deposits and underlying Cohansey Sand have not displayed any of the frozen-ground characteristics common to all the other upland areas. The braided channels and festoon crossbeds of the deposits suggest massive discharge events. Peak discharges for these streams currently run in the hundreds of cubic feet per second. An order of magnitude increase in discharge (to thousands of cubic feet per second) could have been generated by thawing and dewatering of the top 3 ft of the regolith. Large storms may have further overloaded the Mullica watershed. The melting event would have moved one last mass of debris to the mouths of tributaries. The braided channels are in the area of converging flow where discharge accumulates exponentially. Downstream, major erosion cut the channels deeper and a final veneer of alluvial debris was shed out on the exposed continental shelf. These deposits are now found under Holocene swamp, bay, and beach deposits along the coast (Ferland, 1981). Following warming and melting of permafrost, the surface aquifer and present hydrologic conditions were established.

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8. Geochemical Variation in Pore-Water Samples from the Freehold, New Jersey, Core

By Amleto A. Pucci, Jr.,[1] *and* James P. Owens [2]

INTRODUCTION

The pore water of confining units can be shown to significantly affect the solute chemistry in an aquifer, although few field studies have directly collected hydrochemical data on pore water from confining units (Back, 1985). This report describes the geology and solute chemistry from core samples of confining-unit and aquifer sediments at a test corehole (25–566) drilled near Freehold, Monmouth County, N.J. (fig. 8.1), and identifies several geochemical variations. The test corehole was cored continuously to bedrock (1,320 ft) and penetrated seven aquifers and eight confining units. Nearly total core recovery was achieved for silts and clays; recovery of sands was more difficult. Therefore, it was possible to sample pore water within confining units, where the geochemical controls are different from those in aquifers.

Acknowledgments

The authors thank the New Jersey Geological Survey, which supported the drilling, and Frank Manheim, U.S. Geological Survey, who provided the core-squeezing equipment and advice on the sampling methods.

GEOLOGY AND MINERALOGY

The corehole is in the northern part of the New Jersey Coastal Plain, which is underlain by a wedge-shaped body of unconsolidated sediments that contains scattered, thin indurated units. These sediments range in age from Cretaceous to Holocene. Cretaceous and Tertiary sediments generally strike northeast and dip southeast (Owens and Minard, 1975).

The core provided samples of Tertiary and Cretaceous strata and samples of crystalline basement rock (fig. 8.2). The Tertiary formations in the upper 95 ft of the core are mainly intensively weathered marine shelf deposits that include glauconitic sand but consist mostly of quartz sands or silty sands (fig. 8.2).

Cretaceous marine sediments were found at depths ranging from 95 to 625 ft. These sediments, which include significant amounts of glauconite, calcium carbonate shell

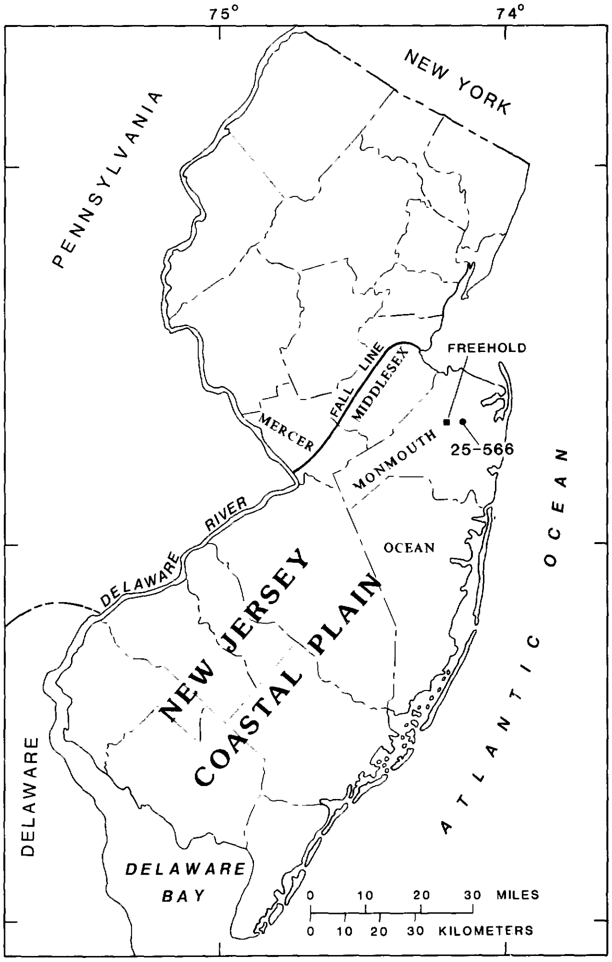

Figure 8.1. Location of the test corehole (25–566) near Freehold, N.J. County boundaries in New Jersey are shown.

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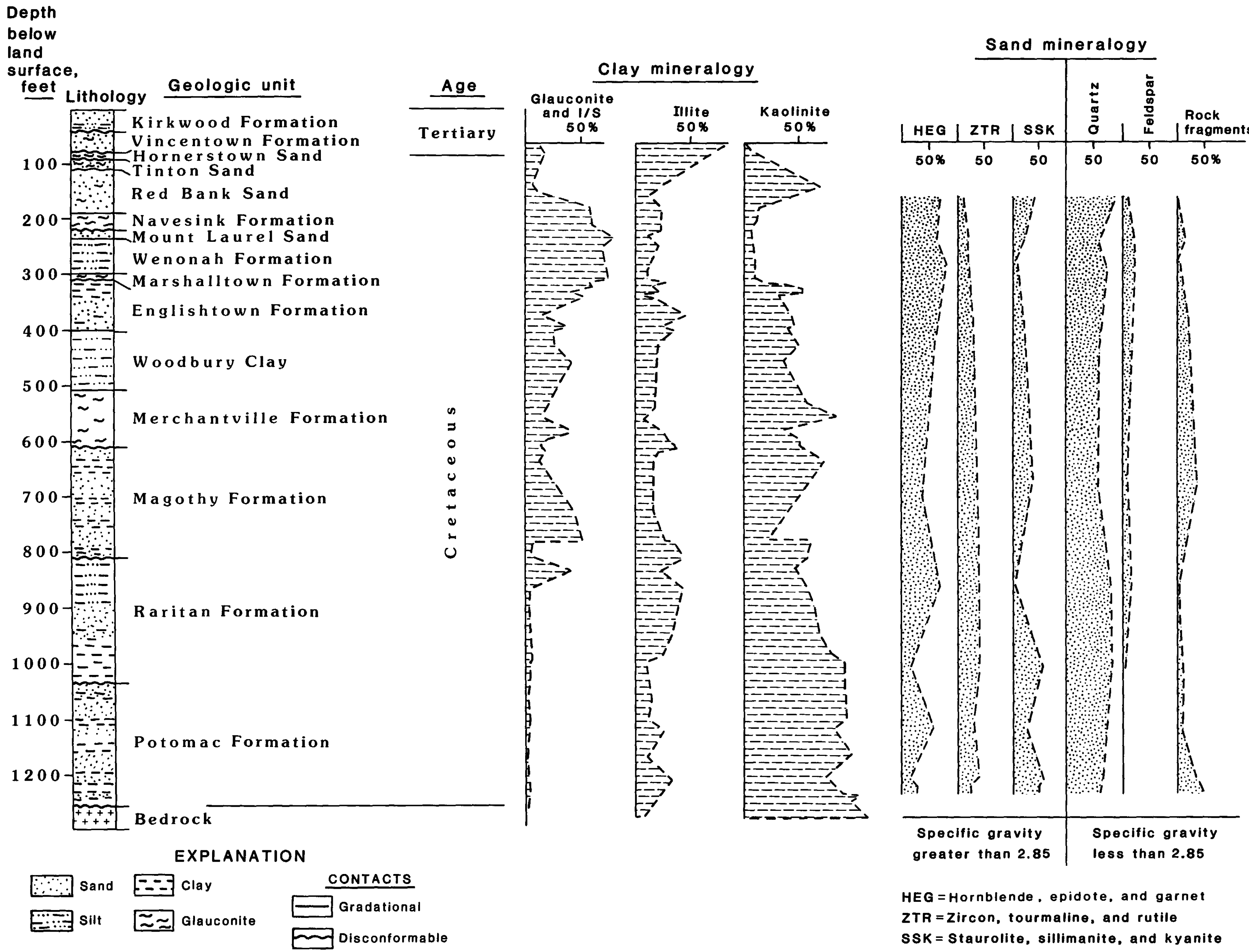


Figure 8.2. Geologic units, lithology, and mineralogy in the core from near Freehold, N.J. Clay and sand mineralogies are reported in weight percentages of the total clay or sand fraction, respectively. I/S, illite/smectite.

material, siderite, and pyrite and minor amounts of goethite, show little evidence of subaerial oxidation. The marine sediments in this section are cyclic deposits, which form thick and regionally continuous confining units and aquifers (Zapecza, 1989).

Fluviodeltaic and marine Cretaceous sediments were found at depths of 625 to 1,277 ft. The fluviodeltaic sediments are complexly interfingered clastic sands and silty clays. Most of the clays and silts in the interval are red, white, or yellow; their colors indicate that they are highly oxidized. As is typical of fluviodeltaic deposits, lithofacies changes are rapid and less predictable than those in the overlying marine section.

In the clay-mineral fraction (fig. 8.2), the content of glauconite and mixed-layer illite-smectite (I/S) generally varies inversely with kaolinite content in the core. I/S is most abundant in the marine and marginal-marine sediments in the upper core (depth interval 160 to 800 ft), whereas kaolinite is most abundant in the fluviodeltaic sediments in the lower part of the core (depth interval 800 to 1,277 ft). Illite, which is less abundant than either I/S or kaolinite in the clay fraction, makes up 10 to 20 percent of the clay fraction throughout most of the core.

Sand minerals were separated into two groups: heavy (specific gravity greater than 2.85) and light (specific gravity less than 2.85) (fig. 8.2). Heavy minerals, which typically compose less than 1 percent of the sand fraction, have been divided into three groups: (1) hornblende, epidote, and garnet (HEG); (2) zircon, tourmaline, and rutile (ZTR); and (3) staurolite, sillimanite, and kyanite (SSK). The HEG group is relatively the most abundant, and it generally increases in proportion from the bottom to the top of the core, mainly because of increases in hornblende. The proportion of the ZTR group is relatively constant over the core length. The SSK group is most abundant in the base of the core and generally decreases in proportion from the bottom to the top of the core.

In the light-sand group, quartz is the most abundant mineral (greater than 50 percent) throughout the core. Rock fragments, principally mica gneiss or nonfeldspathic rocks, constitute the second most abundant fraction of this group throughout the core; the maximum amount is found in the Magothy Formation. Feldspars, mostly potassium feldspar (microcline), compose less than 10 percent of the light-sand group throughout most of the core. The low feldspar and high quartz contents of the light-sand group in the lower core indicate that these formations consist mainly of mature sands.

GEOHYDROLOGY

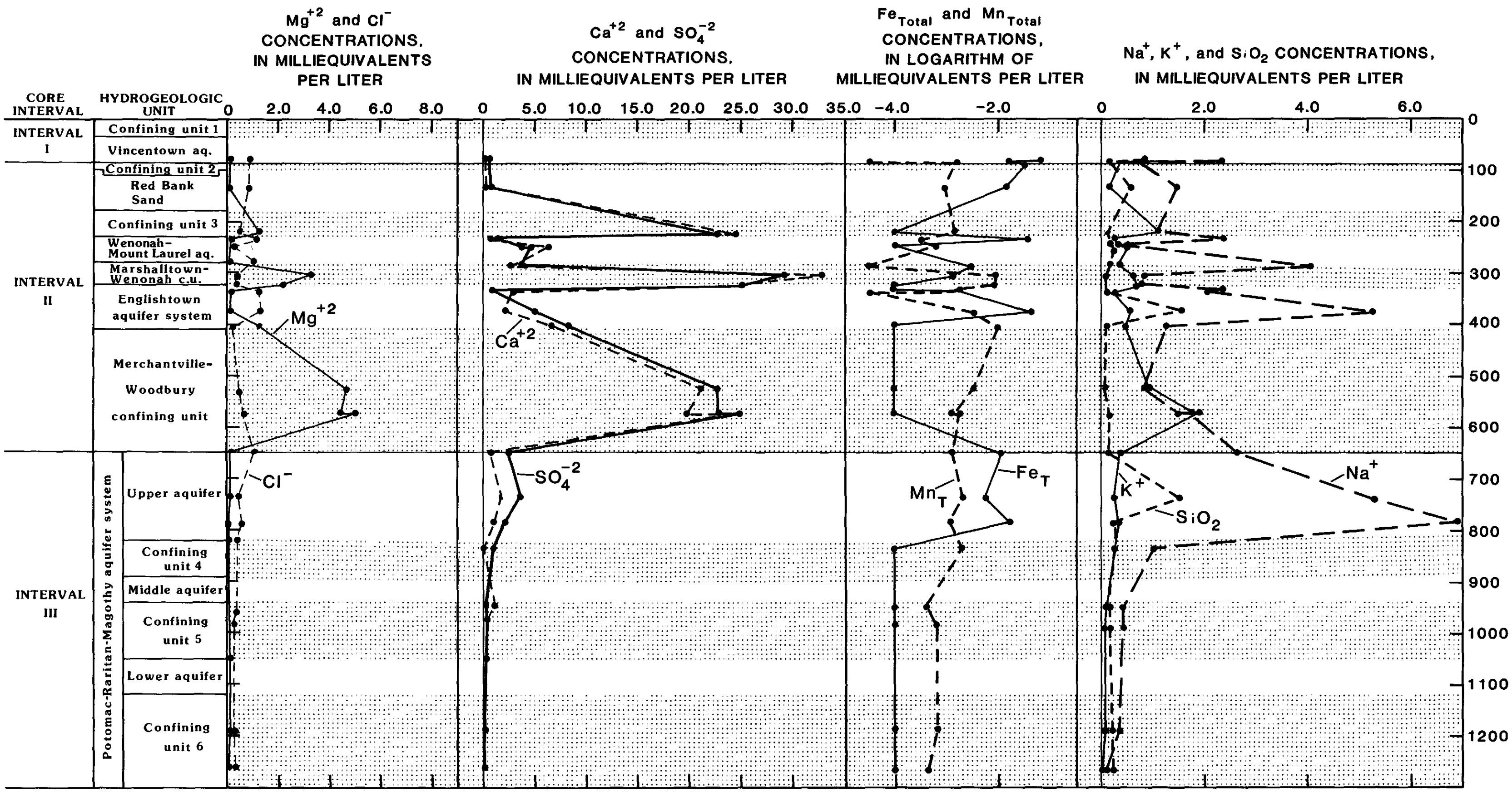
Regional ground-water movement generally is to the east and southeast. The aquifers are recharged by infiltration in outcrop areas northwest of Freehold and by vertical leakage through overlying confining units beneath topographically high areas. A topographic high at the drill site suggests that the local flow could be downward. Simulated predevelopment flow (Martin, in press) had a downward Darcy velocity of approximately 1.0 in. per year into the Wenonah-Mount Laurel aquifer (depth interval, 228 to 298 ft). Downward vertical velocities into lower aquifers from confining units are progressively smaller with depth, ranging from 0.5 in. per year into the Englishtown aquifer system (depth interval, 331 to 416 ft), to less than 0.1 in. per year into the lower aquifer (depth interval, 1,042 to 1,110 ft) of the underlying Potomac-Raritan-Magothy aquifer system. Winograd and Farlekas (1974) concluded that cross-formational flow downward from a confining unit was a major source of dissolved constituents in the upper aquifer of the Potomac-Raritan-Magothy aquifer system near the site of the present study.

METHOD OF PORE-WATER SAMPLING AND ANALYSIS

Pore-water samples were obtained by using the methods and the stainless-steel hydraulic squeezer described by Lusczynski (1961) and Trapp and others (1984). Some central core material (approximately 100 mL volume) was removed immediately after recovery to lessen the possibility of contamination from drilling fluid. The core material was placed in plastic containers, chilled, and stored for several days before squeezing. Pore water was squeezed from the core samples by pressures of 4,000–6,000 pounds per square inch; it was collected in a plastic tube, and the tube was sealed and chilled. The volume of pore water extracted from each of the samples ranged from 5 to 10 mL, and the extraction time ranged from several minutes to several hours. Ion chromatography was used to determine concentrations of major anions, and inductively coupled emission-spectra analysis was used to determine concentrations of the major cations and silica. In all, 26 samples, including three split-spoon samples, were obtained at various depths. Because sample volumes were limited, alkalinity, pH, and aluminum content were not determined for most samples. Manheim (1966) reported that the composition of extracted pore water is not appreciably affected when the mechanical squeezer is used properly. The possibility that the extraction procedure significantly influenced the water chemistry of sediments seemed minimal (Bischoff and others, 1970; Troup and others, 1974; Patterson and others, 1978).

RESULTS AND DISCUSSION

For purposes of discussion, the core is divided into three intervals (fig. 8.3): the shallow, weathered interval (Interval I); the intermediate-depth, unweathered marine



EXPLANATION

aq. = aquifer

c.u. = confining unit = [dotted pattern]

GEOLOGIC UNIT

Confining unit 1 = Kirkwood Formation

Confining unit 2 = Hornerstown Sand-Tinton Sand

Confining unit 3 = Navesink Formation

Confining unit 4 = } Raritan Formation
Confining unit 5 = }

Confining unit 6 = Undifferentiated sediments within the Potomac Formation

Figure 8.3. Variation of selected chemical constituents with depth in the core from near Freehold, N.J.

interval (Interval II); and the deepest fluviodeltaic-silicate interval (Interval III). These intervals are based on pore-water chemistry as well as lithology and depositional environments of the sediments.

Interval I, the weathered interval, is the uppermost interval in the core, ranging from 0 to 95 ft below the land surface. It consists of sediments of the Kirkwood and Vincentown Formations and the Hornerstown Sand. The most abundant minerals are quartz, glauconite, illite, and kaolinite. This interval also contains calcite, siderite, and small amounts of pyrite (mainly in the lowest part of the section). Iron staining (red) and ferric-iron concretions were found in this core interval.

All three pore-water samples from this interval were collected at depths of 81 to 86 ft (near the contact between the Vincentown Formation and the Hornerstown Sand). For these samples, sodium (Na^+) and potassium (K^+) are the dominant cations, and chloride (Cl^-) is a dominant anion, although all concentrations are low. Elevated sodium concentrations in Coastal Plain sediments probably are caused by ion exchange of sodium for divalent calcium and magnesium ions on exchangeable minerals such as glauconite or I/S (Foster, 1950). Dissolved potassium can be formed from degradation of organic material and (or) from glauconite weathering, which was visible in the core (Wolff, 1967). Iron concentrations are higher than those in the underlying Interval II.

Interval II, the unweathered marine interval, is the intermediate part of the core from 95 to 625 ft below the land surface. This interval consists of massive beds of unweathered marine deposits from the Tinton Sand through the Merchantville Formation. Glauconite, pyrite, and organic silt are distributed through the interval. Siderite deposits were found primarily in the confining units.

In Interval II, the pore-water chemistry of the confining units differs from that of the aquifers in several ways: (1) maximum concentrations of magnesium (5 milliequivalents per liter (meq/L)), calcium (greater than 30 meq/L), and sulfate (greater than 30 meq/L) occur in the confining units and minimum concentrations occur in the aquifers; (2) maximum concentrations of sodium (5.25 meq/L) are present in aquifers; (3) total ion concentrations are greater in the confining units than in the aquifers; (4) solute concentrations do not appear to be related to proportions of minerals in the light- or heavy-sand fractions (fig. 8.2) and probably relate to less abundant, but more reactive, solid phases; (5) concentration gradients between the hydrogeologic units are large.

The confining-unit pore water in Interval II is a calcium-magnesium-sulfate type. The distributions of calcium, magnesium, and sulfate in all the confining units are very similar and suggest that a coupled mechanism is occurring. Dissolution of gypsum ($CaSO_4 \cdot 2H_2O$) cannot be one of the mechanisms because it was not found anywhere in the core, nor were these sediments deposited in an evaporite environment.

Aquifer pore water in Interval II ranges from a calcium-sulfate type to a sodium-potassium-carbonate type. The aquifer pore-water samples from this interval had higher sodium concentrations than samples from the confining units and lower calcium and magnesium concentrations. Exchange minerals, such as glauconite and I/S, are abundant in this core interval and can generate dissolved sodium by ion exchange. Sodium-rich minerals such as albite, which can supply sodium by dissolution, are not found in this core interval. Sulfate concentrations in aquifer pore water are significant (as much as 6.5 meq/L) but are much less than those in the confining units.

Interval III, the fluviodeltaic-silicate interval, is the deepest core interval; it extends from 625 to 1,277 ft. The interval consists of sediments from the Magothy, Raritan, and Potomac Formations. The sediments are fluviodeltaic and marginal marine in origin. They are interfingered quartz sands and clays (mainly kaolinite). Pyrite and lignite are found in all three formations. Some glauconite and much siderite are found at the very base of the Potomac Formation and in the upper part of the Raritan Formation.

Sodium is the dominant cation in the pore-water samples and represents more than 70 percent of the total cations for the three aquifer samples from the Potomac-Raritan-Magothy aquifer system. Maximum sodium concentrations for the core (6.9 meq/L) are in the pore-water samples from the upper aquifer of the Potomac-Raritan-Magothy aquifer system (fig. 8.3). No single anion was dominant in the samples from Interval III.

Lower concentrations of dissolved solids below a depth of 800 ft indicate little interaction between sediments and pore water. The three deepest pore-water samples, which are all from the confining units of the Potomac-Raritan-Magothy aquifer system (depths of 1,050, 1,190, and 1,266 ft), are a sodium-bicarbonate-chloride type. The shallower samples, at depths of 834 ft (within the upper part of the Raritan Formation) and 950 ft (in the middle Raritan Formation, near the contact with the middle aquifer of the Potomac-Raritan-Magothy aquifer system), are less chloride rich than the deepest samples.

SUMMARY AND CONCLUSIONS

The pore-water chemistry in the corehole is consistent with the following observations:

1. The pore-water constituents and the observed diagenesis in the core define three distinct intervals in the core: the weathered marine sediments (Interval I), the unweathered marine sediments (Interval II), and the fluviodeltaic-silicate sediments (Interval III).
2. In the unweathered marine interval, the chemistry of pore water in the confining units is very different from

that in the aquifers, and large vertical gradients for concentrations of chemical constituents may prevail between hydrogeologic units.

Additional investigation of pore-water chemistry and the dynamics within the pores is needed to understand the geochemical interactions between aquifers and confining units in this region of the Coastal Plain.

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9. Geologic Controls on Aquifer Distribution in the Coastal Plain of Northern New Jersey

By Peter J. Sugarman[1]

INTRODUCTION

Since 1983, the New Jersey Geological Survey and the U.S. Geological Survey (USGS) have been working on a cooperative project to produce a new geologic map of the State of New Jersey. Work in the Coastal Plain has concentrated on detailed outcrop and subsurface mapping to produce a three-dimensional map integrating lithostratigraphy, biostratigraphy, sedimentary facies analysis, and subsurface geophysical logs. Use of this basin-analysis methodology also has shed significant light on the distribution and stratigraphic framework of some of the aquifer systems in the New Jersey Coastal Plain. This paper gives some examples from Monmouth County that highlight geologic differences between major and minor aquifer systems determined on the basis of interpretive models derived from previous and current regional geologic studies.

METHODS

Detailed geologic mapping was conducted in the Farmingdale, Adelphia, Freehold, and Marlboro 7.5-minute quadrangles (fig. 9.1). Downdip subsurface control was obtained in five wells (fig. 9.1): Marlboro MUA (split spoon), USGS Freehold (continuous core), Farmingdale Boro (cable tool), USGS Allaire State Park C (split spoon), and New Jersey Geological Survey Allaire 2 (continuous core).

Stratigraphic sections were constructed by defining lithologic assemblages as part of a time-stratigraphic framework. This process enables horizontal facies changes to be recognized and allows these lithologic changes to be understood from a paleogeographic perspective. Cross sections constructed by using this approach elucidate the distribution and correlation of aquifer sands and their relations to geologic controls including source, type, and volume of sediment; environment of deposition; and facies relations.

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SEDIMENTARY MODELS

Aquifers in the Coastal Plain can be classified as major and minor. Major aquifers are capable of yielding large quantities of water; minor ones are used locally and may yield small to moderate quantities of water. The cross sections highlighted two basic geologic differences between the major and minor aquifers:

1. The major aquifers in the Potomac, Raritan, Magothy, and Englishtown Formations are parts of large Cretaceous delta systems and display characteristics and styles of deltaic sedimentation. Their sediments are predominantly nonmarine or marginal marine in origin and include upper-delta-plain, lower-delta-plain, and delta-front facies. The aquifers are commonly of uniform thickness and extend long distances downdip.

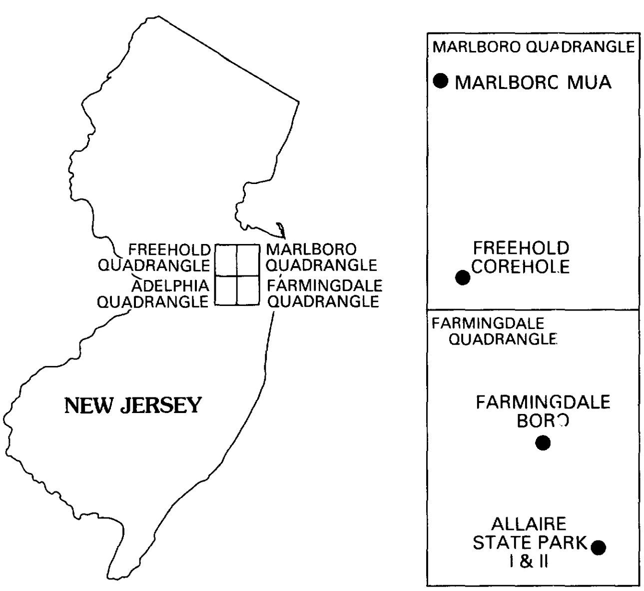

Figure 9.1. Locations of studied coreholes and the four quadrangles composing the study area in the northern New Jersey Coastal Plain. Allaire State Park I, U.S. Geological Survey Allaire State Park C; Allaire State Park II, New Jersey Geological Survey Allaire 2.

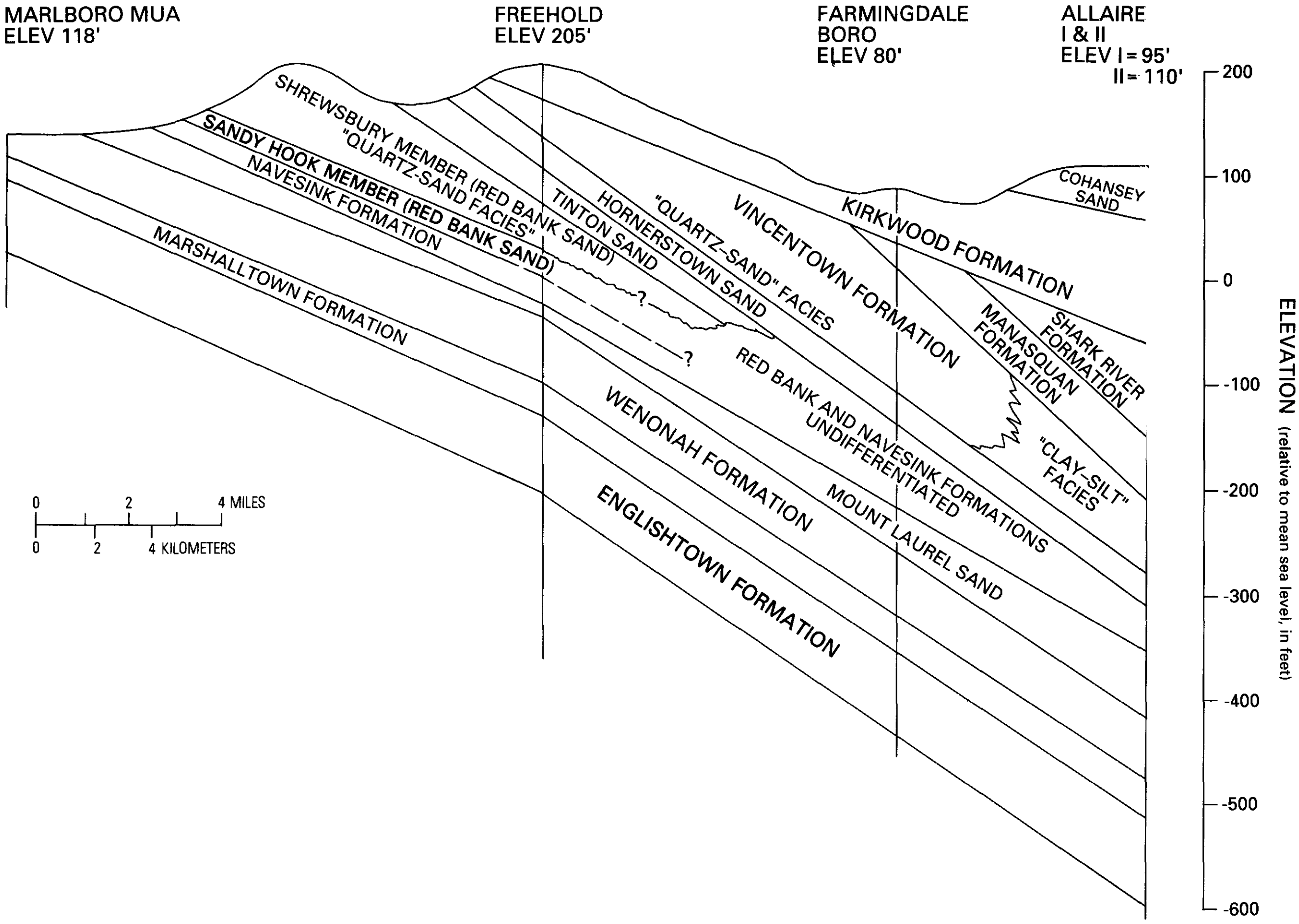

Figure 9.2. Cross section based on five wells from Marlboro to Allaire, N.J. Allaire I, U.S. Geological Survey Allaire State Park C; Allaire II, New Jersey Geological Survey Allaire 2.

2. The minor aquifers in the Vincentown Formation, Red Bank Sand, and Mount Laurel Sand are the products of nondeltaic or interdeltaic deposition (Owens and Gohn, 1985). Sand bodies consist typically of medium-grained quartz, are thick bedded to massive, are slightly glauconitic, and contain marine invertebrate fossils. They were deposited on a shallow shelf near the strandline. Sand aquifers in these units are discontinuous and form localized cells, or depocenters, which pinch out downdip because of facies changes.

A cross section from Marlboro to Allaire is shown in figure 9.2. The transitions from outcrop to the shallow subsurface are summarized below for three minor aquifers and one major aquifer. The Magothy, Raritan, and Potomac Formations (not shown on the cross section) are discussed only from the USGS Freehold corehole. They do not crop out in the study area and were sampled only in this well.

VINCENTOWN FORMATION

The Vincentown Formation is a minor aquifer in the study area. Wells tapping it yield 10 to 15 gallons per minute (gpm) (Jablonski, 1968). Its maximum thickness of 135 ft is present just northeast of Farmingdale. In outcrop along the Manasquan River, the Vincentowr is an olive-brown to gray, massive, medium to coarse, glauconitic quartz sand containing broken shells, bryozoans, and sponge spicules. In the Farmingdale Boro well, about 1 mi downdip from the outcrop belt, the Vincentown remains predominantly a glauconitic quartz sand, but it is finer grained (fine to medium sand) and has considerable silt. The formation thickness remains relatively uniform at 130 ft. At the Allaire State Park well sites, the Vincentown is only 75 to 80 ft thick and is predominantly a noncalcareous micaceous clay-silt to silt. The aquifer sands terminate 12 mi downdip from the outcrop at a facies change (fig. 9.2). This type of facies model is consistent with a nearshore gulf (Shepard and Moore, 1955) or delta-marginal plain as defined by Friedman and Sanders (1978, p. 293). In the second scenario, sand deposition is marginal to a large delta primarily depositing fine-grained sediment. The sands represent thin strandline deposits transported and reworked by waves and longshore currents.

RED BANK SAND

The Red Bank Sand is another local aquifer in Monmouth County and typically produces 3–30 gpm (Jablonski, 1968). It contains two members: the lower Sandy Hook Member consists of dark-gray, glauconitic, micaceous, silty, fine quartz sand, and the upper Shrewsbury Member consists of micaceous, fine to medium quartz sand, which is slightly glauconitic. The Red Bank Sand is conformable with the underlying Navesink Formation, which is a dark-gray, clayey, glauconitic sand. These two formations are an excellent example of a coarsening-upward sedimentary sequence, in which the aquifer sand (Shrewsbury Member) represents the regressive, progradational part of the cycle.

In outcrop in the northwest section of the Marlboro quadrangle, the cyclic package is about 150 ft thick, of which the upper 80 ft is the Shrewsbury aquifer sand. In the Freehold corehole, approximately 5 mi downdip from the area of maximum outcrop thickness, the Shrewsbury Member is thinner (approximately 60 ft), and the underlying Sandy Hook Member is indistinguishable from the Navesink; both are glauconitic sands. At the Farmingdale Boro well, the Shrewsbury Member is absent, and the Navesink and Red Bank constitute an undifferentiated glauconitic sand sequence.

The environment of deposition for the Red Bank is believed to be similar to models discussed for the Vincentown. However, the transition from inner to middle shelf is more evident for the Red Bank, as its quartz sand facies (Shrewsbury Member) pinches out downdip and is replaced by a glauconitic sand.

MOUNT LAUREL SAND

The Mount Laurel Sand is a minor aquifer in the study area, where it is typically combined with the Wenonah Formation into a single aquifer. In this paper, the Mount Laurel Sand is considered as a separate mappable unit.

The Mount Laurel Sand is variable in thickness. In outcrop, as in the subsurface, it may be as much as 60 ft thick. It is predominantly a thin-bedded, fine to medium, glauconitic quartz sand containing thin-bedded clay-silt layers. The sand beds are generally very micaceous and contain lignite. Large *Callianassa* burrows are present. The upper part of the formation is commonly a massive, pebbly, coarse quartz sand.

The cross section (fig. 9.2) shows that its thickness is 10 ft at the Freehold corehole, 40 ft at Farmingdale Boro, and 60 ft at the Allaire State Park wells. In the Marlboro quadrangle, thicknesses range from 5 to 60 ft.

The environment of deposition for the Mount Laurel corresponds closely to the chenier plain model discussed by Kraft (1978). Deposition was centered along the strandline distant from the main deltaic lobe and was dominated by fine-grained sediments. However, instead of one massive nearshore sand body (like those in the Vincentown Formation and Red Bank Sand), a sequence of thin nearshore sand bodies surrounded by silt-clay layers is present. This sand distribution suggests more short-term fluctuations of sea level or more variation in the energy regime during deposition of the Mount Laurel than during deposition of the Red Bank and Vincentown sands.

ENGLISHTOWN FORMATION AND MAGOTHY FORMATION

In the Coastal Plain of northern New Jersey, the Englishtown is a major aquifer. In outcrop, it consists of a gray, fine to medium quartz sand interstratified with dark-gray and brown, carbonaceous clayey silt. The sand is typically crossbedded, micaceous, lignitic, and slightly feldspathic. The formation thickness is consistently 90 to 100 ft in the study area. Nichols (1976) found that Englishtown lithologies in the shallow subsurface strongly resembled those of its outcrop belt. According to Zapecza (1989), the Englishtown thickens in southernmost Monmouth County (Point Pleasant) to 200 ft and can be recognized into Ocean County.

The Englishtown Sand exhibits depositional facies common to a major delta system, including delta-front and lower-delta-plain facies. These facies include nearshore sand deposits, such as barrier-shoreface and channel-mouth-bar facies, and marginal-marine deposits including distributary-channel and levee sediments.

The Magothy Formation is a subsurface unit in the study area; it is more than 200 ft thick in the USGS Freehold corehole. It typically contains intercalated dark, carbonaceous, clayey silts and light-colored sands; thick, laminated carbonaceous clays; and thick, crossbedded lignitic sands (Owens and Gohn, 1985). These facies are consistent with its interpretation as delta-front and prodelta deposits. The Magothy is the equivalent of the upper aquifer in the Potomac-Raritan-Magothy aquifer system (Zapecza, 1989).

RARITAN FORMATION

The Raritan Formation is more than 100 ft thick in the USGS Freehold corehole and unconformably underlies the Magothy. In Monmouth County, the Raritan contains the Farrington aquifer (Farrington Sand Member of the Raritan Formation) and is considered to be the equivalent of the middle aquifer of the Potomac-Raritan-Magothy aquifer system (Zapecza, 1989).

The Raritan is an excellent example of an unconformity-bounded sequence, which is nonmarine at its base but marine from the middle to the top (Sugarman and Owens, 1989). It represents a fining-upward transgressive sedimentary package reflecting a transition from upper-delta-plain to lower-delta-plain to delta-front and, finally, to shelf deposits.

POTOMAC FORMATION

Approximately 300 ft of sediments of the Potomac Formation underlie the Raritan and overlie the basement saprolite in the USGS Freehold corehole. Lithologies include fine to coarse sands, interbedded fine sands and clays with abundant lignite in many of the clays, and poorly sorted mixtures of gravel, sand, silt, and clay. The Potomac Formation in the study area is entirely continental and is characterized by upper-delta-plain and lower-delta-plain facies.

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10. Silcrete near Woodstown, New Jersey[1]

By John S. Wyckoff[2] *and* Wayne L. Newell[2]

INTRODUCTION

Silcrete boulders occur in a belt that extends northeast from Mannington to just north of Swedesboro, N.J. (fig. 10.1). They litter gentle slopes and flat surfaces near 40–50 ft in altitude. Summerfield (1983a) characterized silcrete as a highly siliceous, indurated sand formed within the shallow subsurface as a result of weathering of bedrock or surficial deposits and later cemented at low temperatures and pressures.

Quartzite boulders were previously noted in the New Jersey Coastal Plain by Cook (1855), Knapp and Salisbury (1917), Friedman (1954), Minard (1965), and J.P. Owens (U.S. Geological Survey, oral commun., 1988). Many previous workers did not distinguish among the various occurrences of quartzite having different origins, and previous studies reflect uncertainty concerning silcrete formation and characteristics. Often the silicrete boulders found near Woodstown have been dismissed as remnants of the "Fairton Quartzite" or as derivatives of Paleozoic quartzites from the Valley and Ridge Province. Knapp and Salisbury (1917) suggested that the boulders in the Woodstown area originated from a Miocene sandstone that once covered the Coastal Plain and was locally cemented by silica into a quartzite. Friedman (1954) described what he thought was a similar silica-cemented deposit cropping out along Cohansey Creek in Fairton, N.J., south of the boulder belt. Minard (1965) mapped the Woodstown quadrangle and noted the presence of quartzite boulders; however, he did not speculate on their origin (J.P. Owens, oral commun., 1988). The boulders that Minard described (plus others nearby) are the subject of this study.

We examined the macromorphological and micromorphological characteristics of the silcrete near Woodstown, and, in this paper, we suggest a mode of origin. The methods used in this study are similar to those used in other silcrete studies by Langford-Smith (1978), Wopfner (1978), and Summerfield (1982, 1983a, b, 1984).

[1]Prepared in cooperation with the New Jersey Geological Survey.
[2]U.S. Geological Survey, Reston, VA 22092.

Acknowledgments

John Evans and Elaine McGee, both of the U.S. Geological Survey (USGS), are acknowledged for their helpful and generous assistance with the scanning electron microscope and photographic work. We thank Robert Johnson (USGS), who performed the X-ray spectrometry and chemical analyses. We also thank and acknowledge Phil Schafer, Jim Owens, Lucy McCartan, and Milan Pavich, all of the USGS, for their excellent comments and reviews of this manuscript.

FIELD OCCURRENCE AND MACROMORPHOLOGICAL FEATURES

The silcrete boulders occur in a belt that extends northeast from Mannington to just north of Swedesboro

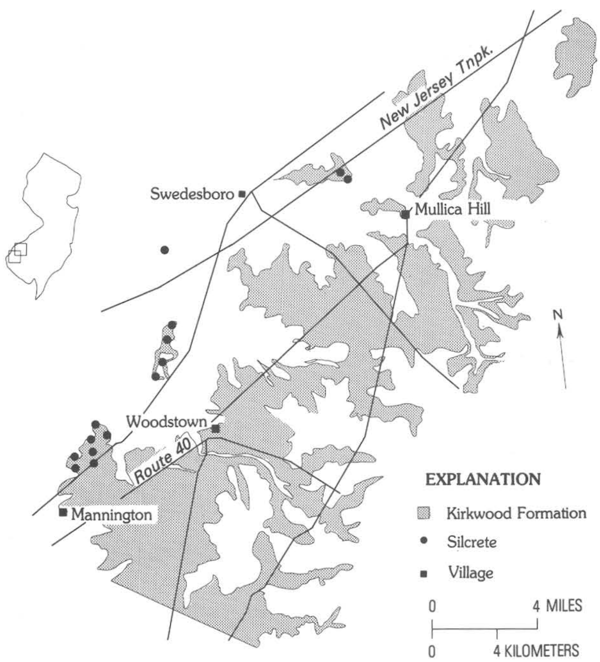

Figure 10.1. Distribution of the Kirkwood Formation and the associated silcrete boulders near Woodstown, in southwestern New Jersey.

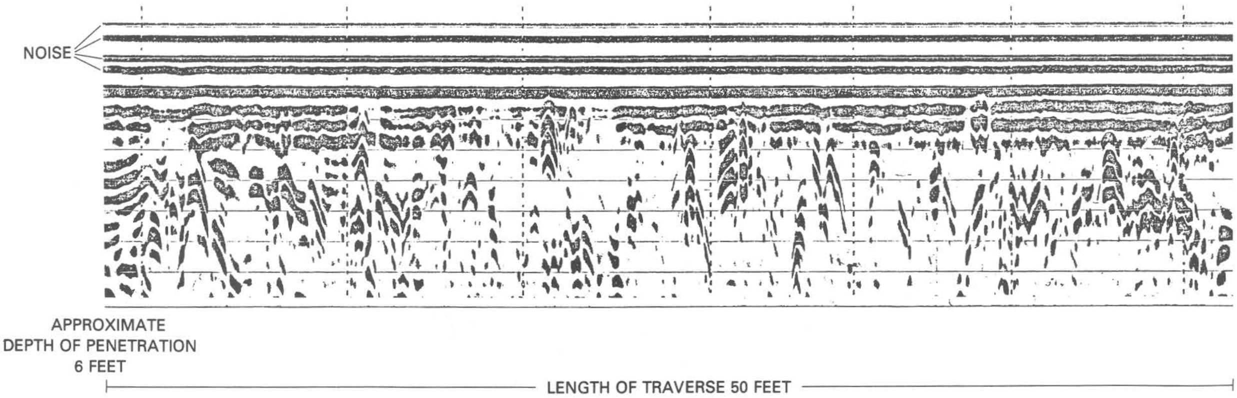

Figure 10.2. A ground-penetrating radar trace along a traverse just north of Mannington. Hyperbolic reflectors indicate point objects, which are silcrete boulders.

(fig. 10.1). The boulders (1.5–3 ft in diameter) litter the landscape and occur along scarps and terraces near 40–50 ft in altitude. Locally, boulders can be seen eroding out of the gently sloping scarps. They have been a perennial nuisance to farmers, who often place them in piles at the edges of their fields or along driveways. They were used in the oldest colonial house foundations and were also used as source material for projectile points by aborigines. Ground-penetrating radar surveys indicate that the boulders are closely packed and occur locally at depths of 40 in. or less (fig. 10.2).

Silcrete boulders are typically composed of well indurated quartz sand ranging from yellowish gray (5*Y* 7/2; color designations are from the Munsell Color Company, 1975) to grayish orange (10*YR* 7/4) with pinkish-gray (5*YR* 8/1) and light-gray (*N*7) mottles. Some silcrete boulders display soil-like structures and pedogenic fabrics (fig. 10.3). Such features include prisms, clay caps, and root casts. Morphologically, the boulders are quite variable, having both faceted and mammillary surface textures. The morphologic variability is the result of wind abrasion, dissolution, and reprecipitation. The variable nature of the boulders reflects their complex history of formation and subsequent destruction.

The present geographic distribution of the boulders coincides with the outcrop of the lower Miocene Kirkwood Formation. The morphologies and boulder distribution indicate that the silcrete formed in the shallow subsurface, above the Alloway Clay Member and within the Grenloch Sand Member of the Kirkwood Formation. Isphording and Lodding (1968) documented weathering within the Alloway Clay Member near Woodstown. They suggested that the kaolinization of the illite-rich Alloway Clay produced large kaolinite crystals. The alteration of illite to kaolinite liberated silica. The Alloway Clay may also be important hydrologically as an aquiclude creating a perched water table within the overlying Grenloch Sand. The Grenloch Sand was indurated by precipitation of the liberated silica. The detrital mineralogy of the silcrete matches the mineralogy of the Grenloch Sand; both consist of mature, multicyclic orthoquartzite.

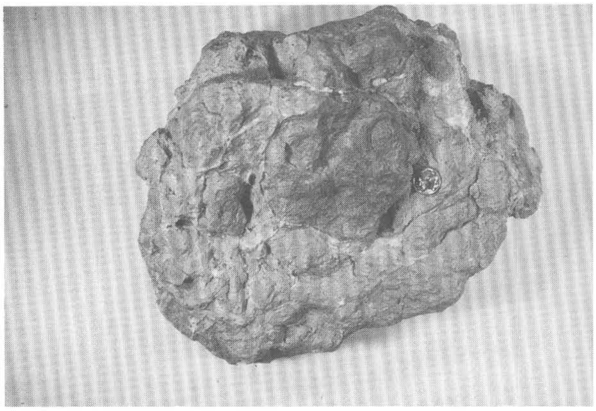

Figure 10.3. A silcrete boulder showing peds and clay skins. The globular feature in the center of the photograph is interpreted to be a ped. Surrounding the ped is a clay skin—a light-tan fine-grained matrix, which actually coats the exterior surfaces of the ped and the adjacent peds. Dime is for scale.

MICROMORPHOLOGICAL FEATURES

Petrographic and scanning electron microscope techniques reveal the complex texture and fabric of the silcrete. Petrographically, the silcrete ranges from a cement-supported to a grain-supported fabric. The framework quartz grains are very angular to subrounded and display embayed and corroded grain boundaries (fig. 10.4). Ilmenite is the second most abundant detrital mineral. Pyrite is only locally abundant (fig. 10.5) and is not volumetrically significant. However, the presence of pyrite within the

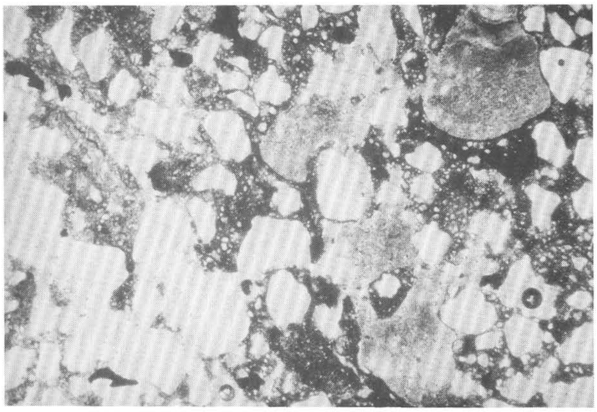

Figure 10.4. Photomicrograph of the silcrete in plane-polarized light. The white grains are quartz. Note the floating fabric and two generations of cement. The first phase is a darker cement, which initially coated most of the framework quartz grains and reduced primary porosity. The second phase, a lighter colored cement, filled the remaining voids. In the upper right corner of the photograph is a colloform feature. Field of photograph is 1.0 mm by 1.5 mm.

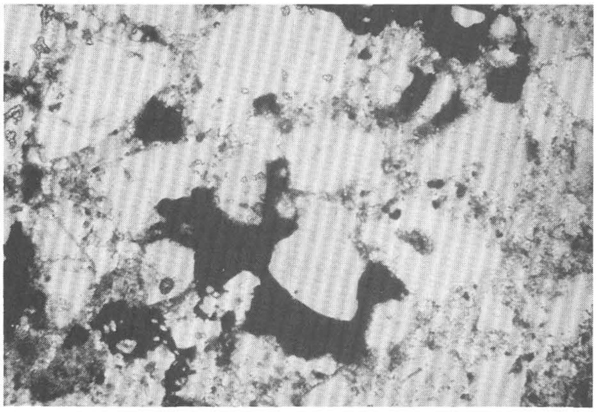

Figure 10.5. Photomicrograph of pyrite-rich zone within the silcrete in plane-polarized light. The quartz grain in the center of the photograph is not corroded where it is in contact with the pyrite (opaque material); however, the opposite boundary, where the quartz is in contact with the silica cement, shows some corrosion. Note the other corroded and noncorroded boundaries of the quartz grains. Field of view is 1.00 mm by 1.5 mm.

silcrete is unique, and no other silcrete studies have documented the association of pyrite and silcrete. We think that the pyrite may reflect the initial geological conditions prior to silcrete formation. Under reflected light and in thin section, authigenic pyrite can be seen rimming the ilmenite grains. In some samples, the pyrite coating surrounds adjacent quartz grains. Quartz grains surrounded by pyrite are not corroded, whereas quartz grains adjacent to silica cement are etched (fig. 10.5). The relation between the nature of the contact and the encompassing authigenic pyrite suggests a sequence of events in which the ilmenite weathered and produced iron oxides. The iron later reacted with bacteria-produced sulfides, under reducing conditions, to form pyrite. After formation of the pyrite, the framework quartz grains were corroded and had etched and embayed grain contacts. We suggest that, during the later dissolution events, much of the pyrite was also oxidized. In only a few instances, the pyrite was preserved as a result of rapid silicification, which enclosed the pyrite, protecting it from further destruction (fig. 10.6).

Thin sections of the silcrete display several generations of a variety of silica (quartz) cements. Microcrystalline quartz (crystals having an average length of 0.001–0.01 mm) cement is the most abundant, and macrocrystalline quartz (crystals that are discernible without a microscope) cement is the next most abundant. The least abundant cement consists of cryptocrystalline quartz (crystals too small to be observed with a microscope). The microcrystalline quartz cement occurs throughout, whereas macrocrystalline quartz is associated with root traces and voids. Cryptocrystalline quartz usually occurs in voids between previously cemented framework quartz grains. Cryptocrystalline quartz cement appears as argillans or colloform features. These structures are thought to be the result of periodic precipitation of quartz cement from a vertically fluctuating ground-water table (Frankel and Kent, 1938; Summerfield, 1983b). The reduction in primary porosity and permeability resulting from cementation would cause enrichment of interstitial water with silica, thus enhancing formation of colloform and argillan features (fig. 10.7).

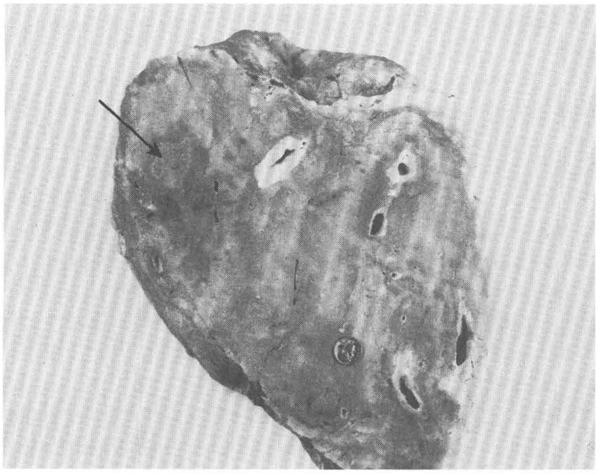

Figure 10.6. Slabbed silcrete boulder. The white zones are the result of what were once roots. The arrow indicates the mottling in the upper left part of the boulder, which is the light-gray (*N7*) zone where pyrite is present and unoxidized. Dime indicates scale.

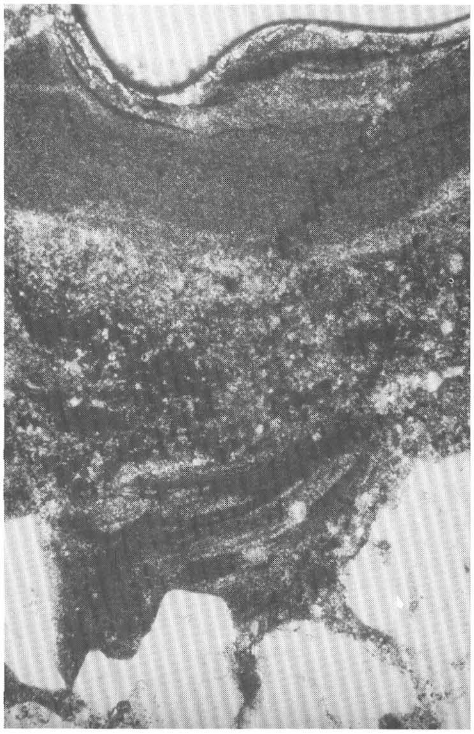

Figure 10.7. Photomicrograph of an argillan in plane-polarized light. Note the banding and the differences in the coarseness of the filling silica cement. Field of view is 0.20 mm by 0.31 mm.

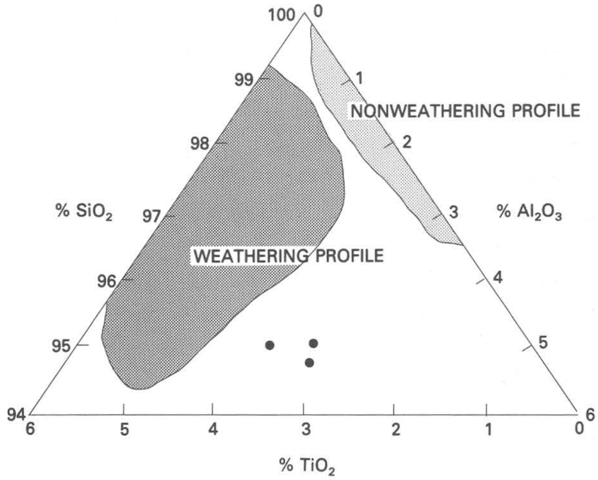

Figure 10.8. Ternary plot modified from Summerfield's (1983b) diagram of genetic compositional fields for silcrete from Africa. Results of three preliminary chemical analyses of the Woodstown, N.J., silcrete are plotted as black dots.

PRELIMINARY CHEMICAL ANALYSIS

Preliminary chemical analysis using X-ray fluorescence spectrometry described by Johnson and King (1987) shows that the silcrete contains about 97 percent silica, 2.5 percent titanium, and less than 0.5 percent iron, aluminum, and other elements. Previous studies of other silcretes have shown similar values for silica and titanium (Summerfield, 1983b). The absence of authigenic titanium-bearing minerals indicates that titanium was not mobilized and reprecipitated. The high titanium value reflects the abundance of detrital ilmenite. The abundance of ilmenite may have increased relative to that of other minerals as a result of selective weathering of more labile minerals lacking titanium.

Figure 10.8 shows the three analyses plotted on a ternary diagram modified from Summerfield (1983b). The analyses do not plot within either of Summerfield's two genetic compositional fields (weathering profile or nonweathering profile) for silcrete from Africa. Thus, the fields described by Summerfield (1983b) are inappropriate for silcrete in New Jersey. The parent material and (or) the processes producing the New Jersey silcrete must have been different from those producing the silcretes studied by Summerfield (1983b).

DISCUSSION AND SUMMARY

The formation of silcrete in New Jersey was controlled by the outcrop pattern of the members within the Kirkwood Formation and the paleogeomorphic setting, which influenced movement and chemical character of the ground water. Our model suggests that the silcrete formed within the shallow subsurface of a broad, vegetated valley bottom at the contact between the Grenloch Sand Member and the underlying Alloway Clay Member of the Kirkwood Formation (lower Miocene). The geomorphic setting was characterized by Owens and Minard (1979) as a northeast-trending valley with fluvial deposition during the late Miocene and early Pliocene. The silcrete probably formed during subtropical to warm-temperate climatic conditions characterized by ample precipitation and leaching. Wolfe and Poore (1982) reported that the Miocene was the warmest epoch in the Neogene whereas alternating warm and cool intervals prevailed during the Pliocene. Overall, Miocene and Pliocene temperatures were warmer than present temperatures in the northeastern U.S. Atlantic Coastal Plain (Wolfe and Poore, 1982). L.W. Ward (oral commun., 1988) uses benthic mollusks to infer subtropical to warm temperate, shallow marine conditions during the late Pliocene in Maryland and Virginia. Because the late Miocene and early Pliocene were warmer than any period in the Quaternary and because of the geologic and geomorphic settings that existed, we suggest that the silcrete was cemented during the Pliocene.

The formation of the silcrete was influenced by the difference in lithologies of the Grenloch Sand and the underlying Alloway Clay. Isphording (1970) noted extensive alteration of the Alloway Clay near Woodstown. Alteration resulted in the enrichment of kaolinite at the expense of illite and montmorillonite. Isphording and Lodding (1968) suggested that upward leaching by ground water caused the transformation of the various clay minerals to kaolinite and the concomitant release of silica. We, however, suggest that ground water moving both laterally and vertically altered the upper part of the Alloway Clay. Then, the silica-enriched water from the underlying weathered Alloway Clay precipitated and cemented the overlying Grenloch Sand. Another possible source of silica is the dissolution of the framework quartz grains prior to silicification. However, if this were a major factor, silcrete would probably be far more abundant in the siliciclastic sediments of the New Jersey Coastal Plain than it is.

The pyrite formed before dissolution and silicification. Reducing conditions prevailed during the formation of the pyrite. Subsequently, oxidizing conditions prevailed. During oxidizing conditions, etching and silicification occurred in an episodic manner. The silcrete formed in the shallow subsurface beneath a broad, level, vegetated valley bottom. Ground-water levels probably were at or close to the surface, as in a swamp. Then dissection of the stable surface produced a fluctuating ground-water table. Under oxidizing conditions and during continued movement of large quantities of water through the profile both laterally and vertically, the sediment was silicified. Rapid local silicification preserved pyrite from oxidation. Continued downcutting and dissection of the paleovalley bottom produced a scarp between the low terraces of the valley and the Coastal Plain uplands to the east. As the scarp backwasted, the scarp surface became armored with boulders, which slowed the rate of retreat and preserved the adjacent bottomland surface beneath which the silcrete is found. Today the silcrete does not exist as a continuous layer because the layer was subjected to a variety of physical and chemical processes, which have been active since the middle Pliocene. These include continued dissolution, freezing and thawing, and various slope processes. Freeze-thaw processes, which are known to have taken place during the Pleistocene, could have significantly contributed to the breakup of the layer and would explain its discontinuity.

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11. Hydrogeologic Units in the Coastal Plain of New Jersey and their Delineation by Borehole Geophysical Methods

By Otto S. Zapecza[1]

INTRODUCTION

An essential element of any ground-water investigation is the development of a working hydrogeologic framework. As part of the U.S. Geological Survey's Regional Aquifer-System Analysis (RASA) study of the northern Atlantic Coastal Plain, a regional hydrogeologic framework was developed for the Coastal Plain of New Jersey. The definition of hydrogeologic units (aquifers and confining units) within this framework is based on the hydraulic properties of the sediments and their correlation with established rock-stratigraphic units. This paper describes the hydrogeologic units in the Coastal Plain of New Jersey and how they can be delineated by using borehole geophysical techniques.

Within the Coastal Plain of New Jersey (fig. 11.1), most regional subsurface mapping is based on formal geologic (rock-stratigraphic) and chronologic (time-stratigraphic) units that are defined by lithologic and biostratigraphic correlations determined by analysis of borehole samples. An understanding and description of the regional ground-water flow system require a determination of the hydrogeologic framework. Many hydrogeologic-unit boundaries differ from boundaries of time- and rock-stratigraphic units. For example, a geologic formation may contain more than one aquifer, a formation may function as an aquifer in one area and as a confining unit in another, or an aquifer or confining unit may be composed of several geologic formations.

The characteristics of the ground-water system of the Coastal Plain are related directly to the geology. Water-bearing properties are a function of the lithology, thickness, and lateral extent of the various geologic formations. Therefore, knowledge of the depositional and postdepositional history of the geologic units is essential to the determination of hydrogeologic-unit boundaries and to an understanding of the ground-water flow system.

The New Jersey Coastal Plain is a seaward-dipping wedge of unconsolidated sediments of Cretaceous to Quaternary age. These sediments, for the most part, consist of clay, silt, sand, and gravel and are of continental, coastal, or marine origin. The Coastal Plain sediments thicken seaward from a featheredge at the Fall Line to greater than 6,500 ft at the southern tip of Cape May County (fig. 11.1).

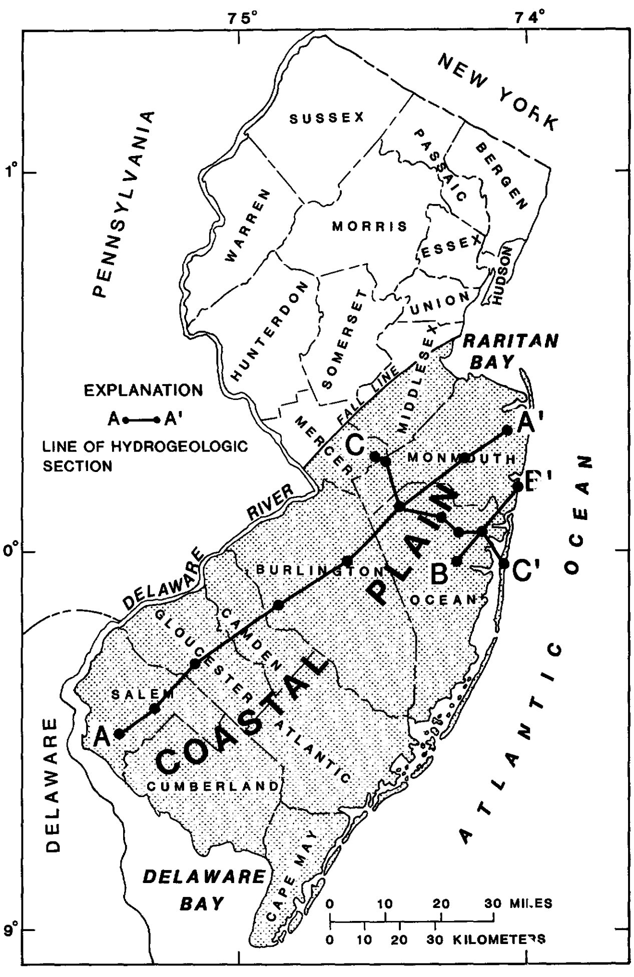

Figure 11.1. New Jersey Coastal Plain and lines of hydrogeologic sections shown in figures 11.2, 11.3, and 11.4.

[1]U.S. Geological Survey, West Trenton, NJ 08628.

This sedimentary wedge forms a complex ground-water system in which the sands and gravels function as aquifers and the silt and clay function as confining units.

DESCRIPTION OF GEOLOGIC AND HYDROGEOLOGIC UNITS

A stratigraphic column of geologic units and associated hydrogeologic units in the Coastal Plain of New Jersey is shown in table 11.1. The lowermost deposits of the Potomac Group and a large part of the Raritan Formation are fluvial or fluviodeltaic in origin and contain discontinuous lenses of gravel, sand, silt, and clay.

Cretaceous and most Tertiary sediments overlying the Raritan Formation were deposited in various shelf and beach environments during cyclic marine transgressions and regressions. Glauconite, common in this part of the geologic section, is indicative of middle to outer shelf deposition (Owens and Sohl, 1969, p. 259). Large concentrations of glauconite in association with fine-grained sediments are indicative of transgressive deposits, which formed during major incursions of the sea. Such units include the Merchantville, Marshalltown, and Navesink Formations, the Hornerstown Sand, and the Manasquan Formation. Sedimentary sequences that coarsen upward and overlie the major glauconitic units are termed regressive deposits. These deposits accumulated in inner shelf, nearshore, and beach areas during the slow retreat of the sea. Such units include the Englishtown Formation, Wenonah Formation, Mount Laurel Sand, Red Bank Sand, Vincentown Formation, Kirkwood Formation, and Cohansey Sand. Generally, transgressive deposits form confining units within the Coastal Plain and regressive deposits form aquifers.

Continental deposition returned to the Coastal Plain during late Tertiary and Quaternary times. The Beacon Hill Gravel and Bridgeton, Pensauken, and Cape May Formations are composed primarily of fluvial sand and gravel (Owens and Minard, 1979, p. D1).

The principal aquifers of the New Jersey Coastal Plain are the Kirkwood-Cohansey aquifer system, the Atlantic City 800-foot sand of the Kirkwood Formation, the Wenonah-Mount Laurel aquifer, the Englishtown aquifer system, and the Potomac-Raritan-Magothy aquifer system. Minor aquifers include the Rio Grande water-bearing zone of the Kirkwood Formation, the Piney Point aquifer, the Vincentown aquifer, and the Red Bank Sand.

The Kirkwood-Cohansey aquifer system is the major unconfined aquifer. It consists of hydraulically connected sediments of the Kirkwood Formation, Cohansey Sand, and overlying surficial deposits. The Cohansey Sand is confined by clays within the Cape May Formation in the peninsular part of Cape May County. The Kirkwood Formation contains thick, silty clay units and two confined aquifers along the New Jersey Coast. The upper confined aquifer—the Rio Grande water-bearing zone—is used mainly in Cape May County. The principal aquifer—the Atlantic City 800-foot sand—is the lower aquifer.

The predominantly fine grained, low-permeability geologic units between the aquifers of the Kirkwood Formation and the Wenonah-Mount Laurel aquifer are grouped together as one composite confining unit. However, parts of the Red Bank Sand, Vincentown Formation, Shark River Formation, and Piney Point Formation are used locally as a source of water.

The Cretaceous sediments of New Jersey contain a major aquifer and two major aquifer systems. The Wenonah-Mount Laurel aquifer consists of the coarse-grained upper part of the Wenonah Formation and the Mount Laurel Sand. The fine-grained lower part of the Wenonah Formation and the Marshalltown Formation form a relatively thin, leaky, confining unit between the Wenonah-Mount Laurel aquifer and the Englishtown aquifer system. The Englishtown aquifer system consists of upper and lower sand facies in southeastern Monmouth County and northeastern Ocean County. Silt and clay of the Merchantville Formation and Woodbury Clay form the most extensive confining unit in the New Jersey Coastal Plain. This confining unit separates the Englishtown aquifer system from the Potomac-Raritan-Magothy aquifer system. The Potomac-Raritan-Magothy aquifer system contains five mappable hydrogeologic units of variable extent. The five units include three aquifers, designated the upper, middle, and lower aquifers on the basis of stratigraphic position within the system, and two confining units that lie between the aquifers. A more detailed description of the individual hydrogeologic units within the New Jersey Coastal Plain, including structure-contour and thickness maps, is presented in a report by Zapecza (1989).

DELINEATION OF HYDROGEOLOGIC UNITS

The primary goal in developing a hydrogeologic framework is to define and delineate areas of permeable and relatively impermeable rock. This goal is achieved by mapping on the basis of lithology. In the Coastal Plain of New Jersey, the cyclic deposition of fine-grained sediments (clay and silt) that alternate with coarser grained sediments (sand and gravel) makes the use of borehole geophysical logs ideal for lithologic correlation. Geophysical logs are made by lowering sensors in boreholes or wells and recording the corresponding measurements at the surface. Borehole geophysical logs—primarily natural gamma-ray and electric logs—provide continuous objective records of the lithologies penetrated by the borehole or well. Correlation of geophysical logs with detailed geologic and paleontologic data from numerous coreholes has shown that many geologic formations or parts of formations have the same or similar geophysical characteristics throughout their areal

Table 11.1. Geologic and hydrogeologic units in the Coastal Plain of New Jersey
[Modified from Zapecza, 1989, table 2]

SYSTEM	SERIES	GEOLOGIC UNIT	LITHOLOGY	HYDROGEOLOGIC UNIT	HYDROLOGIC CHARACTERISTICS
Quaternary	Holocene	Alluvial deposits	Sand, silt, and black mud.	Undifferentiated	Surficial material, commonly hydraulically connected to underlying aquifers. Locally some units may act as confining units. Thicker sands are capable of yielding large quantities of water.
		Beach sand and gravel	Sand, quartz, light-colored, medium- to coarse-grained, pebbly.		
	Pleistocene	Cape May Formation	Sand, quartz, light-colored, heterogeneous, clayey, pebbly.		
Tertiary	Miocene	Pensauken Formation			
		Bridgeton Formation		Kirkwood-Cohansey aquifer system	A major aquifer system. Ground water occurs generally under water-table conditions. In Cape May County, the Cohansey Sand is under artesian conditions.
		Beacon Hill Gravel	Gravel, quartz, light-colored, sandy.		
		Cohansey Sand	Sand, quartz, light-colored, medium- to coarse-grained, pebbly; local clay beds.		
		Kirkwood Formation	Sand, quartz, gray and tan, very fine to medium-grained, micaceous, and dark-colored diatomaceous clay.		
				Confining unit Rio Grande water-bearing zone Confining unit	Thick diatomaceous clay bed occurs along coast and for a short distance inland. A thin water-bearing sand is present in the middle of this unit.
				Atlantic City 800-foot sand	A major aquifer along the coast.
				Composite confining unit	Poorly permeable sediments.
	Oligocene	Piney Point Formation[1]	Sand, quartz and glauconite, fine- to coarse-grained.	Piney Point aquifer	Yields moderate quantities of water.
	Eocene	Shark River Formation			
		Manasquan Formation	Clay, silty and sandy, glauconitic, green, gray, and brown; contains fine-grained quartz sand.		Poorly permeable sediments.
	Paleocene	Vincentown Formation	Sand, quartz, gray and green, fine- to coarse-grained, glauconitic, and brown clayey, very fossiliferous, glauconite and quartz calcarenite.	Vincentown aquifer	Yields small to moderate quantities of water in and near its outcrop area.
		Hornerstown Sand	Sand, clayey, glauconitic, dark-green, fine- to coarse-grained.		Poorly permeable sediments.
Cretaceous	Upper Cretaceous	Tinton Sand	Sand, quartz and glauconite, brown and gray, fine- to coarse-grained, clayey, micaceous.		
		Red Bank Sand		Red Bank Sand	Yields small quantities of water in and near its outcrop area.
		Navesink Formation	Sand, clayey, silty, glauconitic, green and black, medium- to coarse-grained.		Poorly permeable sediments.
		Mount Laurel Sand	Sand, quartz, brown and gray, fine- to coarse-grained, slightly glauconitic.	Wenonah-Mount Laurel aquifer	A major aquifer.
		Wenonah Formation	Sand, very fine- to fine-grained, gray and brown, silty, slightly glauconitic.	Marshalltown-Wenonah confining unit	A leaky confining unit.
		Marshalltown Formation	Clay, silty, dark-greenish-gray; contains glauconitic quartz sand.		
		Englishtown Formation	Sand, quartz, tan and gray, fine- to medium-grained; local clay beds.	Englishtown aquifer system	A major aquifer. Two sand units in Monmouth and Ocean Counties.
		Woodbury Clay	Clay, gray and black, and micaceous silt.	Merchantville-Woodbury confining unit	A major confining unit. Locally the Merchantville Formation may contain a thin water-bearing sand.
		Merchantville Formation	Clay, glauconitic, micaceous, gray and black; locally very fine grained quartz and glauconitic sand are present.		
		Magothy Formation	Sand, quartz, light-gray, fine- to coarse-grained. Local beds of drak gray lignitic clay. Includes Old Bridge Sand Member.	Potomac-Raritan-Magothy aquifer system: Upper aquifer	A major aquifer system. In the northern Coastal Plain, the upper aquifer is equivalent to the Old Bridge aquifer and the middle aquifer is equivalent to the Farrington aquifer. In the Delaware River Valley, three aquifers are recognized. In the deeper subsurface, units below the upper aquifer are undifferentiated.
		Raritan Formation	Sand, quartz, light-gray, fine- to coarse-grained, pebbly, arkosic; contains red, white, and variegated clay. Includes Farrington Sand Member.	Confining unit Middle aquifer	
	Lower Cretaceous	Potomac Group	Alternating clay, silt, sand, and gravel.	Confining unit Lower aquifer	
Pre-Cretaceous		Bedrock	Precambrian and lower Paleozoic crystalline rocks, metamorphic schist, and gneiss; locally Triassic sandstone and shale and Jurassic diabase are present.	Bedrock confining unit	No wells obtain water from these consolidated rocks, except along Fall Line.

[1] Of Olsson and others, 1980.

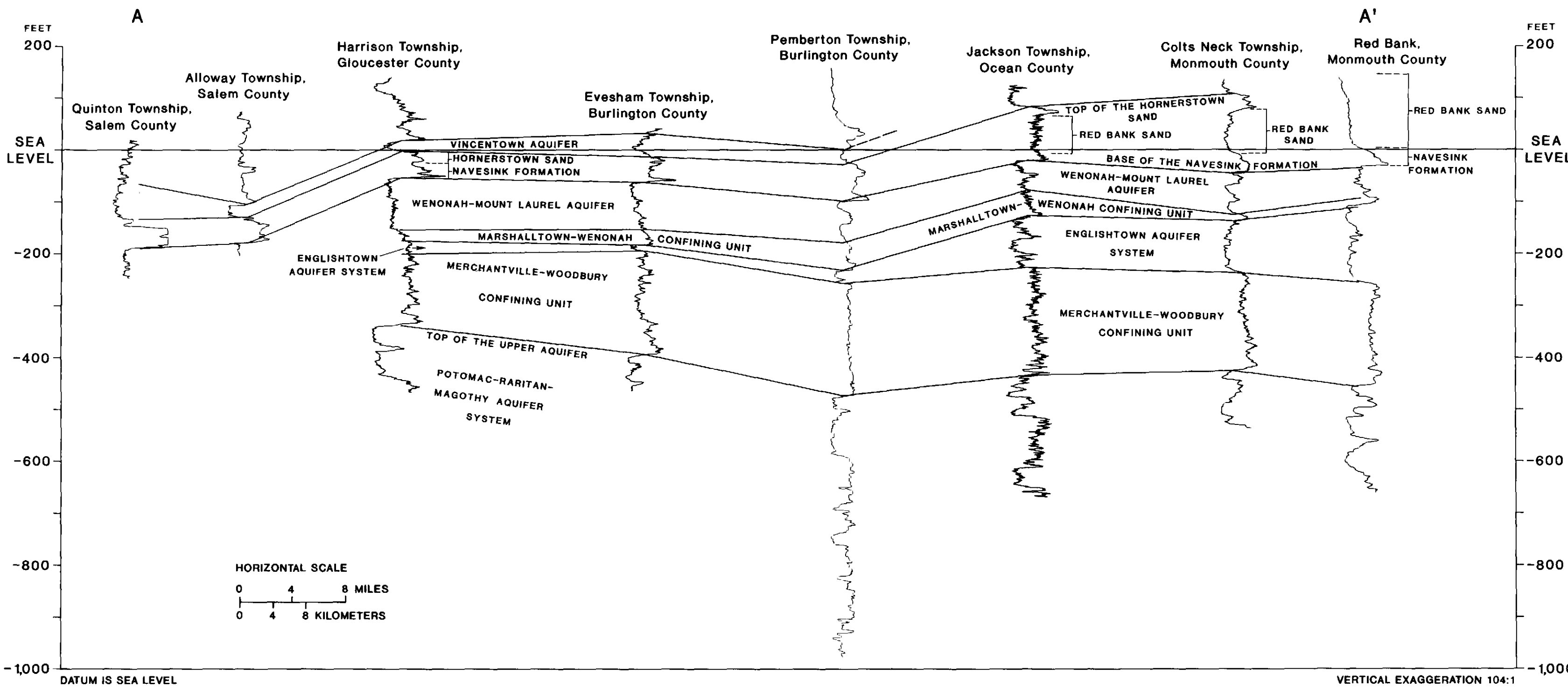

Figure 11.2. Hydrogeologic section *A–A'* from Quinton Township, Salem County, N.J., to Red Bank, Monmouth County, based on gamma-ray logs. The line of section is shown in figure 11.1. From Zapecza (1989, pl. 4).

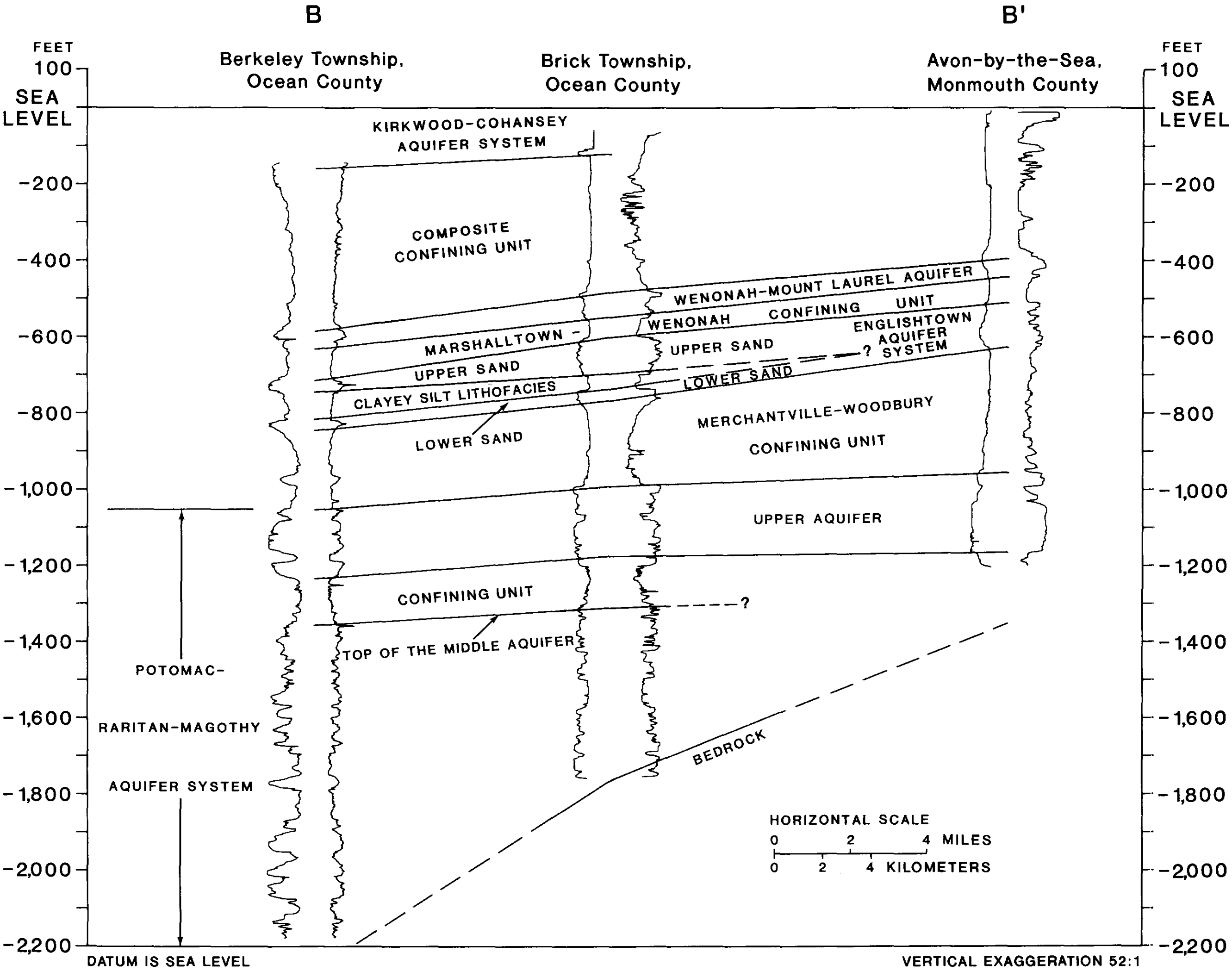

Figure 11.3. Hydrogeologic section *B–B'* from Berkeley Township, Ocean County, N.J., to Avon-by-the-Sea, Monmouth County, based on electric logs. The line of section is shown in figure 11.1. From Zapecza (1989, pl. 5).

extent. Distinctive "signatures" and characteristic patterns on natural gamma-ray logs and electric logs can mark contacts between aquifers and confining units more reliably than can drillers' logs or geological descriptions of drill cuttings.

Natural gamma-ray logs are graphical plots of the rate of emission of gamma rays from the formations penetrated by the borehole. In general, silt and clay have much higher natural gamma activity than do clean quartz sands and carbonates because clays concentrate radioactive elements through ion exchange and adsorption. Feldspars and micas, which decompose readily into clay, also contain small amounts of the gamma-emitting radioisotope potassium-40 (Keys and MacCary, 1971, p. 65). Gamma radiation increases to the right on the gamma-ray log. Therefore, permeable sediments such as sand and gravel, which generally have low radioactivity, are indicated by log deflections toward the left, whereas silt and clay, which generally contain radioisotopes, are indicated by log deflections toward the right.

Figures 11.2, 11.3, and 11.4 are hydrogeologic sections showing correlations made on the basis of borehole geophysical logs. Section *A–A'* (fig. 11.2) is a series of natural gamma-ray logs of hydrogeologic units that extend approximately 90 mi along strike from Salem County to Monmouth County.

Correlation of natural gamma-ray logs in section *A–A'* (fig. 11.2) shows the lateral continuity of some major hydrogeologic units. Of note is the uniformity of characteristic gamma-log signatures for the Merchantville-Woodbury confining unit and the Navesink Formation and Hornerstown Sand. The Merchantville-Woodbury confining unit has a consistent thickness of about 200 ft along the section. In contrast, the Englishtown aquifer system thins markedly to the southwest. From Burlington County north, the typical signature of the Navesink Formation and Hornerstown Sand

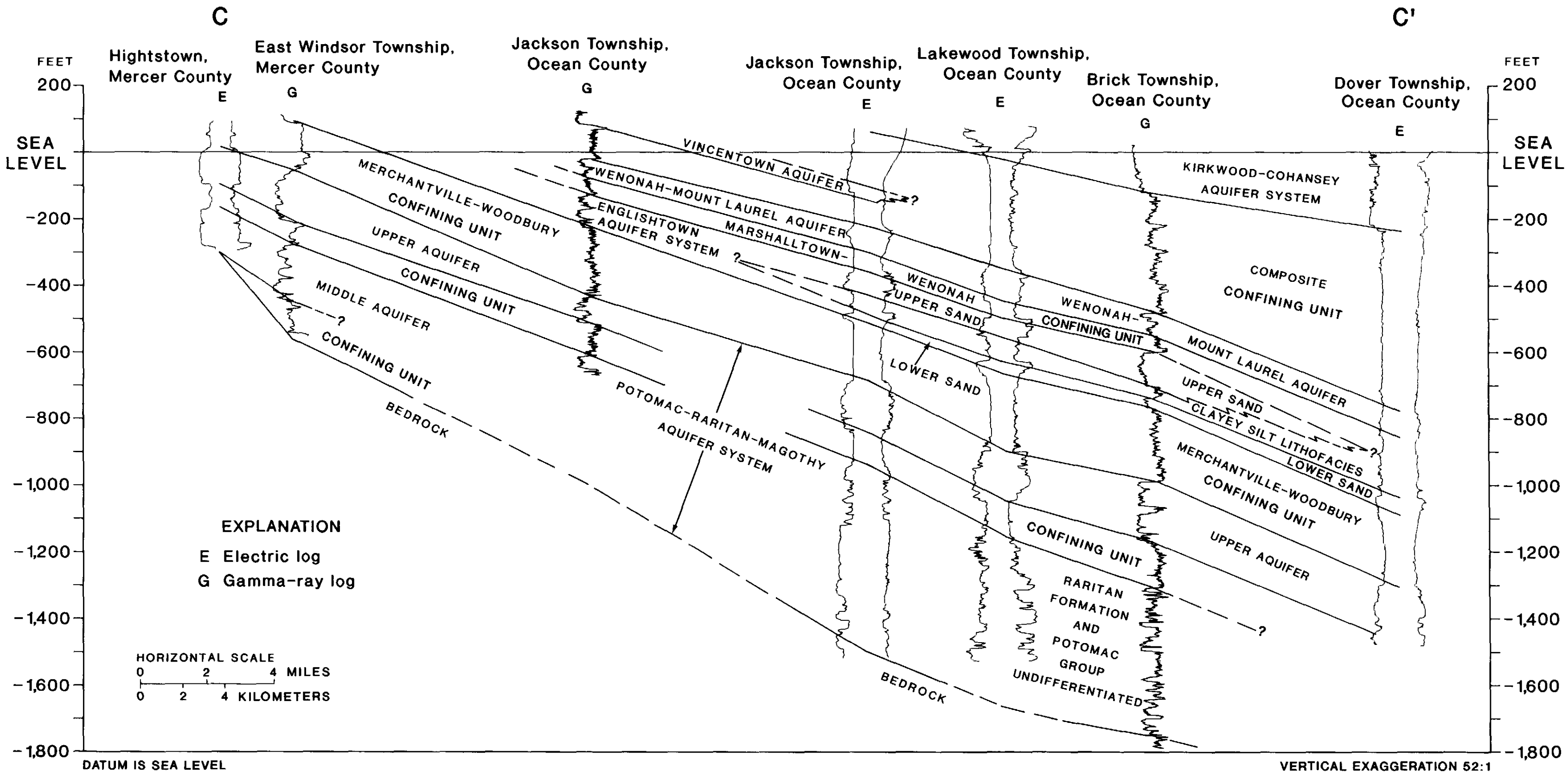

Figure 11.4. Hydrogeologic section C–C′ from Hightstown, Mercer County, N.J., to Dover Township, Ocean County, based on electric and gamma-ray logs. The line of section is shown in figure 11.1. From Zapecza (1989, pl. 3).

shows a progressively increasing separation between high-radiation deflections. This is caused by the northeastward thickening wedge of the Upper Cretaceous Red Bank and Tinton Sands.

Section *B–B′* (fig. 11.3) shows the correlation of hydrogeologic units by use of three electric logs. The electric logs in figure 11.3 are dual-track logs that consist of the spontaneous-potential curve on the left-hand track and a conventional single-point resistance curve on the right-hand track.

The spontaneous-potential curve is a record of small changes in voltage caused by electrochemical reactions between the borehole fluid and the surrounding formation materials. In general, sands cause deflections to the left and clays cause deflections to the right.

The single-point resistance curve is a record of electrical resistance of formation materials penetrated by the borehole. Most sands and gravels are more resistant to the flow of electric current than are silts and clays. Sands and gravels are indicated by sharp deflections to the right on the electric log; silts and clays are indicated by deflections to the left.

Section *B–B′* (fig. 11.3), from northeastern Ocean County to southeastern Monmouth County, shows the characteristic electric-log patterns for several major aquifers and confining units. Simply, the aquifers are identified by the divergence of the two curves, whereas the confining units are identified by the convergence of the curves.

Section *C–C′* (fig. 11.4), from eastern Mercer County to northeastern Ocean County, shows the downdip correlation of hydrogeologic units made by using electric and natural gamma-ray logs. Similarities in geophysical characteristics of the units, as shown on the logs, permit simple and accurate correlation of the principal aquifers and confining units.

In addition to correlation of geophysical logs with geologic and paleontologic data from cores, the interpretation of hydrogeologic units discussed herein is substantiated by (1) long-term water-level measurements of the principal aquifers and their response to ground-water withdrawals (Walker, 1983; Eckel and Walker, 1986; Zapecza and others, 1987) and (2) the vertical location of screened intervals in major supply wells.

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Central Atlantic Coastal Plain

12. A Summary of the Geological Evolution of Chesapeake Bay, Eastern United States[1]

By Steven M. Colman,[2] Jeffrey P. Halka,[3] *and* C.H. Hobbs III[4]

INTRODUCTION

The seaward margin of the U.S. Atlantic Coastal Plain has fluctuated through time, from near the Fall Line to near the edge of the present Outer Continental Shelf, owing to changes in relative sea level. The strata that underlie the Coastal Plain were deposited in environments that ranged from fully terrestrial to fully marine. Estuarine environments are critical components of the Coastal Plain; they represent the interface, otherwise known as the shoreline, between the marine and terrestrial depositional systems. The Quaternary evolution of estuaries has important implications for both documenting the history of sea-level changes and interpreting ancient coastal-plain strata.

In this paper, we briefly summarize the Quaternary history of the Chesapeake Bay, the largest of the many Coastal Plain estuaries on the Atlantic coast. This summary is based on recent syntheses of a wide variety of data (Colman and others, 1988, 1990; Colman and Mixon, 1988) on the history and evolution of the bay.

DATA AND METHODS

The Quaternary stratigraphic record in the Chesapeake Bay and the Delmarva Peninsula area is interpreted primarily from three basic types of data: (1) almost 1,600 mi of shallow-penetration, high-resolution, seismic-reflection profiles collected in the main part of the Chesapeake Bay (fig. 12.1); (2) onshore geologic mapping; and (3) boreholes drilled both onshore and in the bay for engineering work, water wells, and stratigraphic studies.

The seismic-reflection data were collected by using both boomer-type systems and 3.5- to 5-kHz systems (Colman and Hobbs, 1987, 1988; Colman and Halka, 1989a, b). The seismic signals were filtered between 300 Hz and 5 kHz and were recorded at a 0.25-s sweep rate. Loran-C was used for navigation during the seismic-reflection surveys.

Results of recent detailed surficial geologic mapping and descriptions of boreholes for the southern Delmarva Peninsula have been published by Mixon (1985). Additional unpublished core data were used to refine ideas about the locations and depths of the ancient channels of the Susquehanna River beneath the Delmarva Peninsula (Colman and Mixon, 1988). Boreholes in the bay itself are concentrated along the bridge and tunnel crossings (Ryan, 1953; Hack, 1957; Harrison and others, 1965) and near the bay mouth (Meisburger, 1972; Colman and Hobbs, 1987).

PALEOCHANNELS OF THE SUSQUEHANNA RIVER

The Quaternary stratigraphy beneath the Chesapeake Bay is dominated by paleochannels cut into the underlying Tertiary marine deposits by the Susquehanna River and its tributaries and by the sediments that fill those channels. We have identified three distinct generations of these paleochannel systems, which we informally call the Cape Charles, the Eastville, and the Exmore paleochannels, in order of increasing age (fig. 12.1). All three channels cross beneath the southern Delmarva Peninsula, and each is named for a geographic feature on the peninsula.

Seismic-reflection and borehole stratigraphic data clearly show that the three paleochannel systems are of different ages and that the sediments that fill them are separated by significant unconformities. The courses of the paleochannels are rarely coincident, although they commonly intersect. Their relative ages can be determined by map patterns and by crosscutting relationships seen on seismic-reflection profiles. The three paleochannel systems have been mapped throughout the bay; their courses projected from the seismic-reflection data in the bay coincide exactly with their known positions onshore (fig. 12.1).

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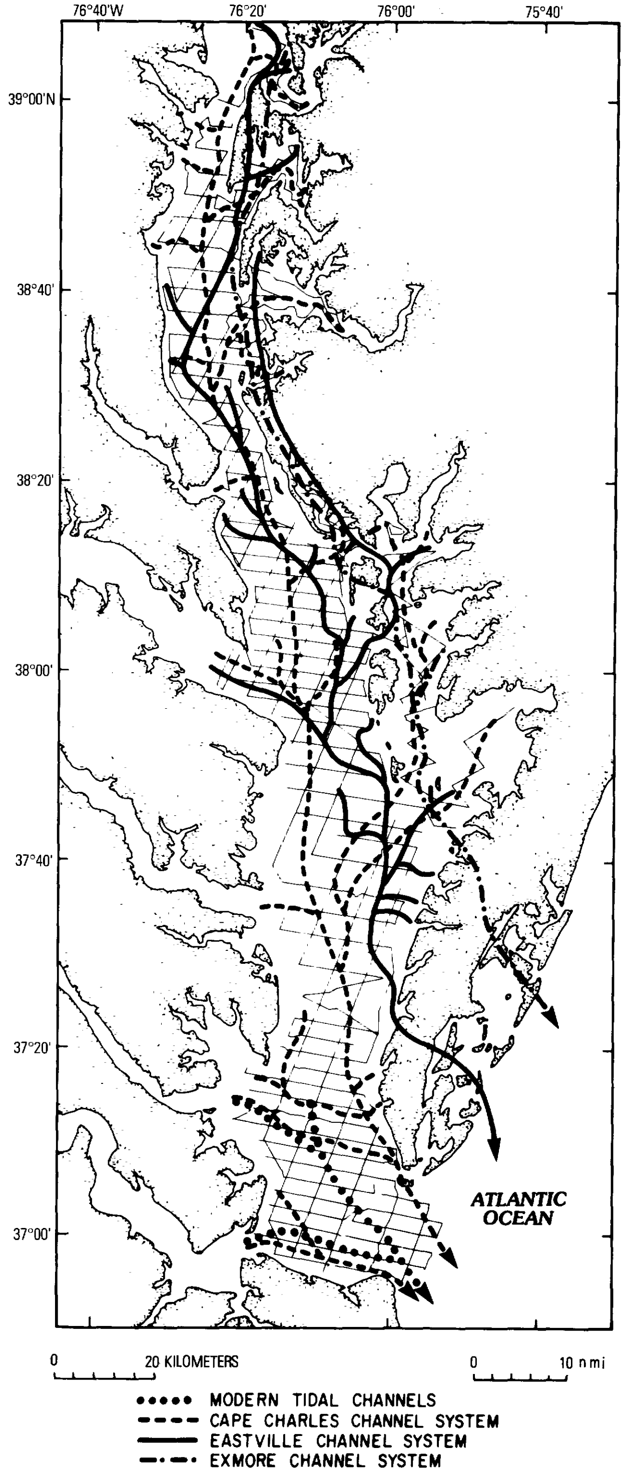

Figure 12.1. Modern tidal channels in the Chesapeake Bay and the three ancient channel systems of the Susquehanna River. The channel systems are listed in order of increasing age; see text. The Cape Charles paleochannel crosses beneath Fishermans Island at the southern tip of the Delmarva Peninsula. The light grid shows tracklines along which seismic-reflection profiles were collected. n mi, nautical miles.

The paleochannel-fill sequences have been divided into two units, whose seismic-reflection attributes are distinctly different. The lower unit of each fill is characterized by relatively strong, irregular, discontinuous reflections, whereas the upper unit of each fill is characterized by relatively weak, long, smooth, continuous, gently dipping reflections. These seismic characteristics, together with lithologic and paleontologic data from relatively deep boreholes (Harrison and others, 1965; Mixon, 1985), indicate that the lower channel-fill unit of each paleochannel is a fluvial deposit, typically consisting of coarse sand and fine gravel. The upper unit of each paleochannel fill, in contrast, was deposited either in restricted river-estuary to open-bay environments or in nearshore-marine environments at the bay mouth. The lithologies of the upper units of the paleochannels are commonly complex, consisting of interbedded muddy sand, silt, and peat. The upper, estuarine, units are finer grained than the lower, fluvial, units, and the estuarine units become finer grained both upsection and upbay.

Where the paleochannels underlie present land areas, their internal structure and fill lithology are known from well logs and stratigraphic boreholes. The Eastville paleochannel is especially well documented where it crosses beneath the Delmarva Peninsula. Mixon (1985) divided the channel fill into several units and showed that, on the Delmarva Peninsula, the paleochannel is overlain by a barrier-spit complex. Both the channel geometry and the fill stratigraphy derived from the borehole data are remarkably similar to those derived from the seismic-reflection profiles of channels beneath the bay.

The geometry and stratigraphy of the paleochannel systems indicate that the channels were formed during periods of low sea level, when the mouth of the Susquehanna River was far out on the present continental shelf. The geometries of the paleochannel systems are similar. The main trunk channel of each system is about 1 to 2.5 mi wide and about 100 to 160 ft deep. Longitudinal profiles of the paleochannels are irregular, and the overall gradients of the trunk channels within the bay are unexpectedly low. During the last major low sea-level stand, about 18,000 yr ago, sea level was perhaps 280 ft below present sea level on the mid-Atlantic Continental Shelf (Dillon and Oldale, 1978). Near the eastern margin of the bay, the bases of the channels are about 200±20 ft below present sea level, and they presumably grade to the deeper lowstand shorelines on the Outer Continental Shelf.

The channel systems show progressively less relation to the present configuration of the Chesapeake Bay with increasing age (fig. 12.1). The channels are relatively close together in the northern part of the Chesapeake Bay, but they diverge significantly toward the southeast; all three cross beneath the present Delmarva Peninsula. Where they cross the peninsula, the major paleochannels are progressively younger toward the south.

The primary reason for this systematic divergence and southward age progression is the southward progradation of the Delmarva Peninsula during major interglacial high sea-level stands (Colman and Mixon, 1988). The latest episode of this progradation process is evident in the late Holocene history of the bay mouth, where the axial channel of the bay has been displaced as much as 7.5 mi in the last few thousand years (Colman and others, 1988). This progradation of the peninsula and the southward migration of the mouth of the bay were episodic, occurring only during the highest of interglacial high sea-level stands (Colman and Mixon, 1988). As sea level fell following a major interglaciation, the displaced estuarine channel became the new fluvial channel, the previous generation of the fluvial channel and its fill were preserved, and the course of the Susquehanna River was altered.

AGES OF THE PALEOCHANNELS AND CHANNEL-FILL SEQUENCES

Evidence for the ages of the paleochannels comes from a variety of chronometric and stratigraphic data. The ages of the Cape Charles paleochannel and its fill are relatively well known because they represent the most recent sea-level cycle and because radiocarbon ages are available for the channel fill. The paleochannel is correlated with marine oxygen-isotope stage 2 (Colman and Mixon, 1988), the peak of which occurred about 18,000 yr ago (Imbrie and others, 1984). The Cape Charles paleochannel has been only partly filled by the Holocene transgression, and the channel itself is clearly related to the low sea-level stand associated with the last major glaciation, the late Wisconsinan. Radiocarbon ages from the channel fill range from about 8,000 to 15,000 yr before present (Harrison and others, 1965; Meisburger, 1972).

Each of the older paleochannels is assumed to correlate with an interval of low sea level of about the same magnitude as that of the late Wisconsinan glaciation and oxygen-isotope stage 2. Each of the older paleochannels is filled with estuarine sediments and overlain by barrier-spit deposits on the Delmarva Peninsula, and no major unconformities exist within these sequences (Colman and Mixon, 1988). Therefore, each of the paleochannels is inferred to correlate with a major glaciation immediately followed by a major interglaciation. These major glacial-interglacial transitions have been called terminations (Broecker and van Donk, 1970); the Cape Charles paleochannel and its Holocene fill represent termination I (Colman and Mixon, 1988). The barrier-spit deposits that overlie the paleochannels on the Delmarva Peninsula represent the last events of previous terminations and thus constrain the ages of the paleochannels.

Uranium-series, uranium-trend, and amino acid age estimates exist for the two ancient barrier systems on the Delmarva Peninsula; the ages of these and nearby deposits have been the subject of considerable discussion and argument, which have been reviewed in relation to the history of the bay by Colman and Mixon (1988). Uranium-series and amino acid age estimates are incompatible for some deposits; ages estimated by both methods conflict with stratigraphic interpretations of other deposits; and some of the uranium-series age estimates do not closely correspond to known times of high sea level. Nevertheless, the barrier spit that overlies the Eastville paleochannel appears to correlate with the last major (Sangamon) interglaciation and with oxygen-isotope stage 5. Accordingly, the Eastville paleochannel presumably dates from oxygen-isotope stage 6, about 150,000 yr ago (Colman and Mixon, 1988). The age of the barrier spit that overlies the Exmore paleochannel is more problematic, but Colman and Mixon (1988) have suggested that these deposits may correlate with oxygen-isotope stage 7 (about 200,000 yr ago) or with stage 11 (about 400,000 yr ago). The Exmore paleochannel likely correlates with the next older stage (stage 8, about 270,000 yr ago, or stage 12, about 430,000 yr ago).

EVOLUTION OF THE CHESAPEAKE BAY

The Quaternary evolution of Chesapeake Bay is intimately related to major eustatic sea-level changes. The record of these changes in the Chesapeake Bay area consists of three generations of paleochannels of the Susquehanna River system, representing low sea levels, and three generations of barrier-spit and channel-fill deposits, representing high sea levels. This unusual record contains features related to both maximum and minimum sea levels, along with nearly complete sedimentary records of three major transgressions. The record has a climax aspect, preserving mainly evidence of the highest and lowest sea levels. On the basis of available age information, the sea-level changes recorded in the Chesapeake Bay area appear to correlate well with the marine oxygen-isotope record and to represent the last few hundred thousand years of sea-level history.

The present Chesapeake Bay is only the latest in a series of at least three generations of the bay, each configured differently. During each period of the bay's existence, progradation of the Delmarva Peninsula caused southward migration of the bay mouth. As sea level fell following each estuarine episode, the bay drained and the fluvial channel of the Susquehanna River incised in a new location, south and west of its former position. As a result, the stratigraphy and morphology of the bay generally became younger toward the south and west. The evolution of Chesapeake Bay shows that coastal-plain estuarine environments can be geologically dynamic, on both short and long time scales.

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13. Dynamics of Upper Paleocene Deposits in the Salisbury Embayment, Virginia and Maryland

By Lucy E. Edwards[1]

The Paleogene sedimentary record in the Salisbury embayment reflects the combined influences of depositional, erosional, tectonic, eustatic, and paleogeographic factors. Eight cores from the Coastal Plain of Virginia and Maryland provide insight into the history of the embayment.

Graphic correlation provides a means for separating various components of the depositional history of an area. The technique, introduced by Shaw (1964), uses two-axis graphs as a means of expressing time-equivalency between stratigraphic sections. The aspect of the technique to be emphasized here is that the line of correlation (LOC) between any section and the composite standard for the area can be used to compare relative rates of sediment accumulation, relative differences in times of onset of deposition, and relative differences in the extent of nondeposition and erosion. Relative age can be expressed in terms of composite standard unit values. Relative rates can be expressed in terms of the slopes of the individual LOC's. Details and applications of the method were provided by Shaw (1964), Miller (1977), Sweet (1979), Edwards (1984, 1989), and Dowsett (1988).

The lower part of the Paleogene section in the Salisbury embayment consists of the lower Paleocene Brightseat Formation (not present everywhere), the upper Paleocene Aquia Formation, the upper Paleocene or lower Eocene Marlboro Clay (not present everywhere), and the lower Eocene Nanjemoy Formation. The first and last appearance datums of dinocyst taxa in eight coreholes were used in the graphic correlation procedure. The dinocyst data are unpublished data of the author, some of which were summarized in Edwards and others (1984), and data from Witmer (1987). This report presents only the results from the Aquia Formation and Marlboro Clay.

Figure 13.1*A* shows a simplified thickness map for the Aquia Formation and Marlboro Clay in the study area. The thinnest accumulation (at the Carters Corner corehole) is assigned a value of one; contour lines represent multipliers of this value. These sediments are thinnest in a narrow north-northeast-trending zone in the western part of the study area (at the Carters Corner and Lake Jefferson coreholes) and thickest in the eastern part (particularly in the northeastern part where these units are 2.5 times as thick at the Solomons corehole as they are at the Carters Corner corehole). Steep gradients occur near the area of thinning. The graphic correlation technique is used to explore the components of this pattern.

For the eight sections studied, the Office Hall section was chosen as the reference section because it was well sampled and centrally located. The Lake Jefferson section was correlated to it to produce an initial composite section; then the Carters Corner section was correlated to this composite to produce a revised composite section. The other cores were added in the same fashion in the order: Ashton, Oak Grove, Fort McLean, Haynesville, and Solomons. As might be expected for coastal-plain deposits, bends occurred in several of the LOC's (bends indicate changes in relative sediment accumulation rates).

Because of the rather consistent locations of bends in the LOC's, the upper Paleocene sediments are discussed in terms of two units, here labeled UP–I and UP–II. Although UP–I is roughly (but not exactly) equivalent to the Piscataway Member of the Aquia Formation and UP–II is roughly (but not exactly) equivalent to the Paspotansa Member of the Aquia Formation and the Marlboro Clay, it is important to note that these are not lithostratigraphic units, but rather, depositional packages recognized by the graphic correlation procedure.

One variable that contributes to the pattern of thicknesses is the difference in the timing of the onset of deposition. The relative position of the base of the Aquia Formation in each core can be noted with respect to the oldest composite unit value found in any core. One can then compute the amount of time (in composite units) for which deposits are missing at the base of UP–I in each core. The percentage of UP–I time for which deposits are missing from each core is contoured in figure 13.1*B* (0 percent missing from the Oak Grove core). This figure shows a preexisting structural low centered around Oak Grove,

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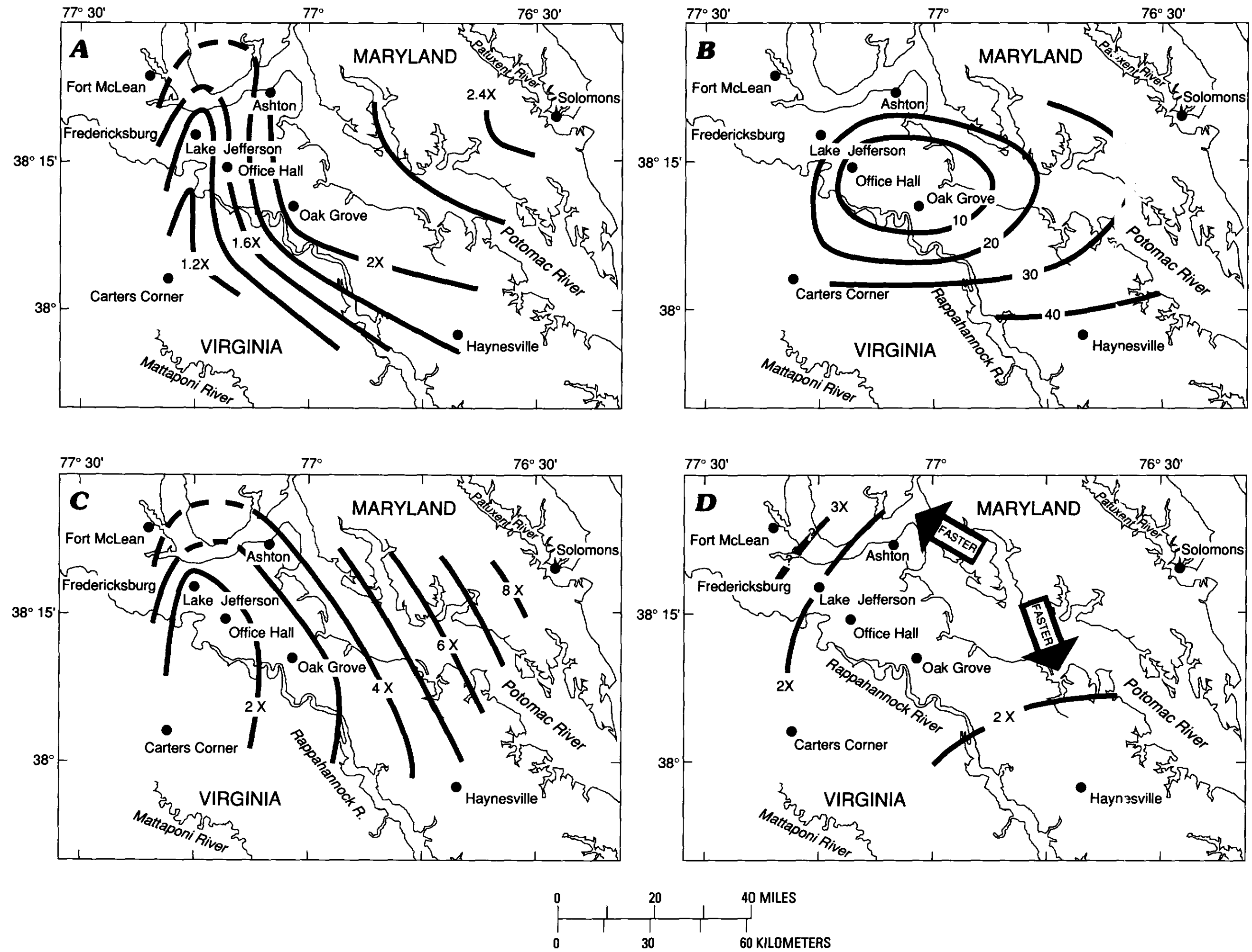

Figure 13.1. Data on the Aquia Formation and Marlboro Clay from eight coreholes in the Salisbury embayment, Maryland and Virginia. *A*, Simplified thickness for the Aquia Formation and Marlboro Clay in the study area. The thinnest value (at the Carters Corner corehole) is 1; contours represent multipliers of this thickness. *B*, The timing of the onset of deposition of UP–I. Values plotted are the percentage of UP–I time for which deposits are missing because of nondeposition in each core. The Oak Grove section has 0 percent nondeposition. *C*, Relative rates of deposition during UP–I time. Lowest observed value (at the Carters Corner corehole) is 1; contours represent multipliers of this rate. *D*, Relative rates of deposition during UP–II time. Lowest observed value (at the Solomons corehole) is 1; contours represent multipliers of this rate.

which was gradually filled during UP–I time. The exact alignment of this low is poorly defined.

Another interesting facet of the basin history is the relative rate of deposition during the time deposition occurred and for which sediment remains. This rate is found from the slope of the individual LOC's. Periods of time without deposition and periods of time represented by sediments that have since been eroded are not incorporated into this rate. Figure 13.1*C* shows contours of this rate for UP–I time. This figure has been scaled to give the lowest observed rate (at the Carters Corner section) a value of 1, and contours represent multipliers of this value. The rate at the Solomons section is greater than nine times the rate at the Carters Corner section. The pattern appears to represent mainly differences in sediment supply; there is little influence of the preexisting structure shown in figure 13.1*B*. The relatively low rate in the vicinity of Carters Corner, Office Hall, and Lake Jefferson may represent differential subsidence or slow deposition over an emerging structural high.

Figure 13.1*D* shows the changing picture of relative accumulation during UP–II time. Here also, the lowest rate (at the Solomons section) is given a value of 1, and contours represent multipliers of this rate. The range of variation is less than that during UP–I time; the highest rate in UP–II time is less than four times the lowest rate. Because there are relatively few dinocyst datums late in UP–II time, calculated rates may not accurately account for periods of erosion, but the general picture is probably good. Again, the pattern reflects differences in subsidence or sediment supply. During UP–II time, the section at Haynesville received a lot of sediment while the section at Solomons received relatively little.

Figures 13.1*B–D* show some of the factors that combined to produce the record shown in figure 13.1*A*. Thus, we see relatively thin Aquia Formation and Marlboro Clay deposits at Lake Jefferson and Carters Corner, slightly thicker deposits at Office Hall, and still thicker deposits at Ashton, Fort McLean, Oak Grove, and Haynesville. The thickest deposits are at Solomons. The thinnest deposits result from a fairly late onset of deposition coupled with relatively low rates of deposition during the remainder of Paleocene time. The somewhat thicker Office Hall and Oak Grove sections result from an early onset of deposition combined with relatively low to moderate depositional rates. The deposits at Ashton, Fort McLean, and Haynesville have thicknesses similar to those at Oak Grove but are the response to different factors: late onset of deposition but somewhat higher depositional rates. The thickest deposits result from a very high rate during the latter part of UP–I time. This rate overwhelmed the effects of late onset of deposition and a slowing rate during UP–II time.

Thus, the Aquia Formation and Marlboro Clay in the Salisbury embayment reflect a complex interplay of many factors. Preexisting and emerging structures combined with major changes in sediment supply to produce varying thicknesses of deposits and varying depositional rates.

Acknowledgments

I thank R.B. Mixon and T.G. Gibson for drilling and for collaborative research on the cores.

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14. Late Pleistocene and Holocene Development of Delaware Bay

By Harley J. Knebel[1]

INTRODUCTION

The geologic evolution of Delaware Bay since late Pleistocene time has been outlined in two recent papers by Knebel and Circé (1988) and Knebel and others (1988). In these reports, the primary data are medium-penetration seismic-reflection profiles collected along 282 nmi of tracklines in the lower Delaware Bay (fig. 14.1*A*). Interpretations of these subbottom profiles used previously published core and drill-hole data, as well as information on modern sedimentary environments within and around the study area.

In the following discussion, I summarize the major aspects of the development of Delaware Bay from the two papers cited above. The reader should consult the full text of these papers for appropriate supporting references and for a more definitive treatment of the data.

LATE PLEISTOCENE AND HOLOCENE GEOLOGIC HISTORY

Two ancestral drainage systems of the Delaware River traversed the lower Delaware Bay during late Pleistocene time (fig. 14.1*B*). One system can be traced southeastward beneath the northern half of the survey area to where it intersects the shoreline about 3 nmi north of the tip of Cape May. The trunk valley of this system has relief of 59 to 105 ft and axial depths of 125 to 154 ft below present sea level, which deepen eastward down the bay. The width of this valley ranges from 2 to 4 nmi; the average width exceeds 3 nmi.

The trunk valley of the northern drainage system connects with a large filled paleovalley that runs beneath Cape May (fig. 14.1*B*) and extends onto the modern continental shelf. Drill-hole data from Cape May indicate that the bay-peninsula valley system was eroded into the Cohansey Sand (Miocene age) when sea level fell to at least −210 ft, probably during the late Illinoian glacial maximum. Subsequently, the valley was filled with transgressive strata as sea level rose during Sangamonian time. The transgressive sequence reached more than 98 ft in thickness beneath the lower bay area, and it consisted of basal fluvial sands and gravels overlain by heterogeneous estuarine fine sands and muds.

At the height of the Sangamonian sea-level transgression, littoral and nearshore processes built the Cape May peninsula southward over the buried northern drainage system. Coastal deposition at this time also created a contiguous southwest-trending submarine ridge that extended part way across the present bay-mouth area and formed the nucleus of a modern bay-mouth bathymetric sill. The growth of the peninsula and the formation of the submarine ridge shifted the entrance to the bay considerably southward of its early Sangamonian position.

A second ancestral drainage system of the Delaware River was eroded beneath the southern half of Delaware Bay when sea level fell (to about −279 ft) during the late Wisconsinan glacial maximum (fig. 14.1*B*). As sea level was lowered, the Cape May peninsula and adjoining submarine ridge became a highland, which forced the ancestral Delaware River to follow a course around the southern end of the ridge and to flow through the present bay-mouth area. The drainage system included (1) a trunk valley of the Delaware River beneath the central part of the bay; (2) a tributary valley formed by the convergence of secondary rivers along the Delaware coast; and (3) a channel of the Maurice River that followed along the western flank of the Cape May peninsular ridge (fig. 14.1*B*). Like its northern (late Illinoian) counterpart, the trunk valley of this system has a southeast trend, relief of 59 to 105 ft, widths of 2 to 4 nmi (>3 nmi average), and a thalweg that deepens from −128 to −154 ft down the bay. The system was cut into a variety of coastal-plain deposits of Miocene and younger age.

Delaware Bay developed from the southern drainage system during the Holocene transgression. The paleogeography of the bay at different positions of lowered sea level (figs. 14.1*C*,*D*) was controlled primarily by the topography of the late Wisconsinan erosional surface. This surface was not appreciably altered during the transgression, either by

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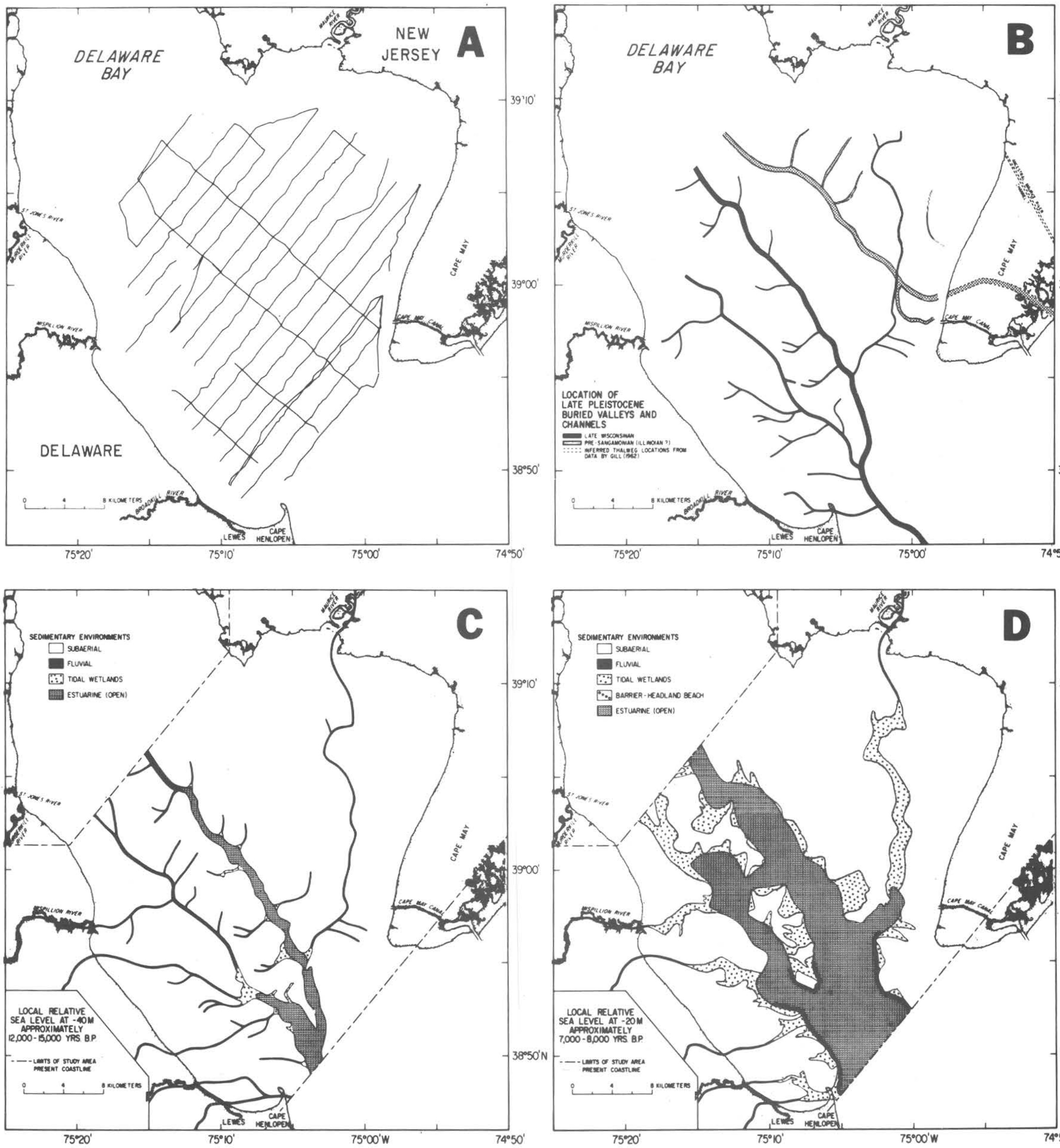

Figure 14.1. Maps of Delaware Bay showing (*A*) tracklines along which seismic-reflection profiles were collected; (*B*) thalweg locations of ancestral drainage systems of the Delaware River; (*C*) paleogeography and sedimentary environments of the ancestral estuary when local relative sea level was at −131 ft (−40 m) approximately 12,000–15,000 yr ago; and (*D*) paleogeography and sedimentary environments of the ancestral estuary when local relative sea level was at −66 ft (−20 m) approximately 7,000–8,000 yr ago. *A* and *B* modified from Knebel and Circé (1988). *C* and *D* modified from Knebel and others (1988).

differential crustal movement or by erosion due to waves and tidal currents.

During the Holocene transgression, the configuration of the ancestral Delaware estuary largely determined the areal distribution of subaerial, fluvial, tidal-wetland, barrier-beach, headland-beach, and open-estuarine sedimentary deposits. Subaerially deposited (pretransgressive) sediments now beneath the bay consist primarily of gravelly, fine to coarse sands or mixtures of sands and muds that range in color from orange to blue gray to white. Fluvial sediments are light-colored, gravelly, coarse sands that commonly are 10 ft thick. Tidal-wetland deposits typically are gray-brown muds containing plant fragments and peat layers, although some deposits have variable amounts of sand, gravel, and shells. Barrier-headland beach sediments are dominantly fine to coarse sands with gravel and shell fragments. Open-estuarine deposits range from clayey silts containing thin beds of fine sand (turbidity-maximum deposits) to interlaminated muds and sands (reworked bay deposits) to clean, coarse sands (flood-tidal deposits).

Fluvial conditions existed within the bay from the time of the late Wisconsinan lowstand (about 16,000–18,000 yr ago) until the shoreline reached the present bay-mouth area. During this interval, coarse fluvial sediments accumulated along the axes of the main channels of the drainage system in response to the elevation of base level. Tidal activity did not affect sedimentation within the bay until relative sea level had reached about −164 ft.

When relative sea level was at −131 ft (−40 m), about 12,000–15,000 yr ago, a small estuary existed within the bay area (fig. 14.1*C*). The estuary included a narrow tidal river (less than 1 nmi wide), which extended more than 16 nmi up the bay along the former trunk valley of the Delaware River, as well as a small contiguous embayment at the bay mouth. Primarily coarse-grained fluvial and tidal-river sediments accumulated within the bay area at this time.

When relative sea level was at −98 ft (−30 m), about 10,000–11,000 yr ago, the developing estuary encompassed two passages (0.5–2 nmi wide), which were formed by the drowned major valleys of the ancestral drainage system. Near the present bay mouth, these passages merged into a relatively wide inlet channel (2–3 nmi wide). At this time, fine-grained sediments accumulated rapidly in the lower bay area, where the turbidity-maximum depocenter was located, and organic muds were deposited in tidal wetlands that occupied the mouths of most tributaries. Fluvial sediments, meanwhile, accumulated within those stretches of the tributaries that were not affected by the tides, and short barrier and headland beaches were present along the margins of the inlet channel.

When local relative sea level was at −66 ft (−20 m), about 7,000–8,000 yr ago, the two major passages of the estuary were joined, except for a series of small intervening islands, and the estuary comprised about 30 percent of the present lower bay area (fig. 14.1*D*). By this time, the turbidity-maximum depocenter had moved landward of the bay area, with the result that fine-grained sedimentation decreased in the open estuary. In addition, the fetch and tidal prism of the open estuary now caused large waves and strong tidal currents to (1) scour and rework previously deposited transgressive sediments, (2) transport sediments into the lower bay area from the continental shelf, and (3) form nearly continuous barrier and headland beaches along the lower bay shoreline (fig. 14.1*D*).

When relative sea level was at −33 ft (−10 m), about 5,000–6,000 yr ago, the estuary was nearly continuous and covered about 60 percent of the present bay area. This increase in continuity and size of the open estuary further enhanced the processes of sediment reworking and transport. Moreover, the wave regime continued to increase the extent of barrier and headland beaches, especially along the southern margin of the bay, and it probably was responsible for the formation of shallow subtidal flats (less than 13 ft deep) near the shore. Tidal wetlands, meanwhile, covered relatively broad expanses along the lower reaches of the tributary channels.

Modern Delaware Bay is characterized by two open-estuarine sedimentary environments. One environment is distinguished by tidal-current transport of the surficial sands and by erosion of previously deposited estuarine strata; these conditions are prevalent under the relatively deep central part of the bay. The second environment is characterized by either sediment reworking or accumulation, processes which dominate across the shallow bay margins.

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15. Nature and Timing of Deformation of Upper Mesozoic and Cenozoic Deposits in the Inner Atlantic Coastal Plain, Virginia and Maryland

By R.B. Mixon,[1] D.S. Powars,[1] *and* D.L. Daniels[1]

INTRODUCTION

U.S. Geological Survey (USGS) studies of Coastal Plain tectonism in Virginia and Maryland began in the middle 1970's. The two objectives were to recognize and map geologic structures in upper Mesozoic and Cenozoic deposits and to determine the nature and timing of deformation in Inner Coastal Plain areas. The ultimate goal was to help assess earthquake risk with regard to the safe siting and construction of nuclear generating stations. As the program wound down in the early 1980's, the emphasis shifted to integrated studies of the structural and stratigraphic framework of both the Inner and Outer Coastal Plain.

PREVIOUS INVESTIGATIONS

The initial field investigations were of the upper Mesozoic and Cenozoic deposits of the Inner Coastal Plain of Virginia (table 15.1) between Washington, D.C., and Richmond, Va. This area was selected for study because previous work (Mixon and others, 1972) had indicated faulting of Coastal Plain beds and because the intense dissection of the thin Coastal Plain formations near the Fall Line provides good outcrops and an abundance of structural control. Our early studies (Mixon and Newell, 1976, 1977, 1978, 1982; Newell and others, 1978) recognized and described the Stafford fault system (fig. 15.1), a series of en echelon, northeast-striking, northwest-dipping, high-angle reverse faults that displace both the Paleozoic crystalline rocks of the Piedmont and the overlying Coastal Plain formations. The Stafford system is known to extend for more than 33 mi along the Virginia Fall Line and the northeast-trending reach of the Potomac River; its position supports the hypothesis that the Fall Line and major river deflections along it have been tectonically influenced (McGee, 1888; Mixon and Newell, 1977).

The history of fault movement along the Stafford system, as determined from outcrop and borehole data and trench studies (Mixon and Newell, 1977, 1978, 1982), indicates that compressional deformation began at least as early as the Early Cretaceous. Of special significance is a detailed trench study of the Dumfries fault zone west of Stafford, Va., where Ordovician slate, phyllite, and schist are thrust at a high angle over the Lower Cretaceous Potomac Formation (Newell and others, 1978). Crosscutting relations of subsidiary faults exposed in the trench clearly show reverse faulting confined to Potomac Formation beds of Early Cretacous age. Major deformation also took place in the middle(?) Tertiary (post-Nanjemoy and pre-Calvert time) and in the late Tertiary (post-Calvert and pre-Choptank time (Mixon and Powars, unpub. data)). At least some Pliocene deformation along the Stafford system is indicated by as much as 1.5 ft of reverse-fault offset of the base of the upland gravel in an artificial cut near the U.S. 1 highway bypass in south Fredericksburg, Va. (Potomac Electric Power Company, 1973; Mixon and Newell, 1982). About 1.0 ft of fault offset of the base of high-level Rappahannock River terrace deposits by the Fall Hill fault, a major strand of the Stafford system, indicates middle to late Pliocene or, possibly, younger faulting (Mixon and Newell, 1978, 1982).

During studies of the Stafford fault system, Mixon and Newell (1977) noted the striking alignment of zones of compressional faulting in the Virginia and Maryland Coastal Plain with linear zones of early Mesozoic extensional faulting in the Piedmont and in the crystalline basement beneath the Coastal Plain (fig. 15.1). The Coastal Plain compressional fault zones and the early Mesozoic extensional faults are aligned, in turn, with Paleozoic thrusts and shear zones in the underlying basement rock. For example, in Virginia, the Stafford fault system is aligned with the border faults of the early Mesozoic Farmville basin trend and with the Spotsylvania magnetic lineament (Neuschel, 1970; Pavlides, 1980), a Piedmont feature representing a broad zone of en echelon faults or shear zones in the crystalline basement. The Lakeside mylonite (Farrar, 1984), which is adjacent to and roughly aligned with the Farmville basin (fig. 15.1), appears to represent a major shear zone in the Piedmont rocks along

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Table 15.1. Age, thickness, and generalized lithology of rock units in the Inner Coastal Plain of Virginia
[Color terms from Goddard and others (1948)]

Rock unit	Age	Thickness (ft)	Lithology
Terrace deposits	Pleistocene and Pliocene	0–50	Sand, gravel, clay-silt; gray, yellowish gray, and brown.
Upland gravel	Pliocene	0–50	Sand, gravel, silt; gray, yellowish orange, and reddish brown.
Yorktown Formation	Pliocene	0–90	Fine to coarse sand, gravelly in part; gray, yellow; contains *Ophiomorpha*.
Eastover Formation	Late Miocene	0–40	Fine shelly sand and silt; dark gray to bluish gray.
Choptank and Calvert Formations.	Early and middle Miocene	0–135	Fine sand, clay-silt; olive gray, diatomaceous.
Nanjemoy Formation:			
Woodstock Member	Early Eocene	0–50	Shelly, glauconitic quartz sand; muddy, micaceous, greenish gray.
Potapaco Member	Early Eocene	50–90	Shelly, glauconitic quartz sand; muddy, micaceous, greenish gray.
Marlboro Clay	Paleocene	0–30	Kaolinitic clay-silt, red to gray.
Aquia Formation	Paleocene	0–125	Shelly, glauconitic quartz sand; light to dark olive gray.
Brightseat Formation	Paleocene	0–15	Fine quartzose sand and silt; micaceous, olive gray to olive black.
Potomac Formation	Early and Late Cretaceous	0–1,400	Feldspathic sand, gravel, and illite-montmorillonite-kaolinite clay.

the Stafford–Spotsylvania–Farmville trend. Similarly, the Brandywine fault system in southern Maryland (Glaser, 1971; Jacobeen, 1972; McCartan, 1989) includes northeast-trending Cretaceous and Tertiary reverse faults and structural highs that closely parallel the northwestern edge of an early Mesozoic rift basin present beneath the Coastal Plain cover. The Brandywine fault system and the underlying rift basin are aligned with the Richmond and Taylorsville early Mesozoic basins in Virginia. Deep test holes, water wells, and an aligned, east-dipping gravity gradient between the Brandywine and Richmond areas suggest continuity of faulting, rift-basin deposits, and basement terranes in Virginia and Maryland according to Mixon and Newell (1977), Wentworth and Mergner-Keefer (1983), and Mixon and Powars (1984). These workers concluded that Cretaceous-Cenozoic faulting in the Virginia and Maryland Coastal Plain followed preexisting zones of weakness in the crust marked by early Mesozoic extensional faults and pre-Mesozoic compressional shear zones. The style of faulting in this area is similar to that of the southeastern Atlantic Coastal Plain (Prowell and O'Connor, 1978; Prowell, 1988).

TAYLORSVILLE BASIN-SKINKERS NECK ANTICLINE STRUCTURAL TREND

Our more recent investigations (Mixon and Powars, 1984; Mixon and others, 1987; Powars, 1987) are concentrated east of Fredericksburg, Va., in King George and Caroline Counties, about 10–20 mi east and downdip of the Stafford fault system (figs. 15.1, 15.2). This area is of particular interest tectonically because it straddles the trace of the early Mesozoic Taylorsville basin border faults as projected from the Virginia Piedmont northeastward beneath a thin to thick cover of Coastal Plain deposits. Additionally, the Hylas cataclastic rocks, which mark a major zone of late Paleozoic ductile shearing and thrusting that borders the west sides of the Richmond and Taylorsville basins in the Piedmont (Bobyarchick and Glover, 1979), project northeastward along the buried part of the Taylorsville basin in the study area. Our mapping in the Coastal Plain along the trend of these old structural discontinuities shows that the regional eastward dip of the overlying Cretaceous and Tertiary strata is interrupted at Skinkers Neck on the Rappahannock River by a long, linear, northeast-trending anticlinal fold (fig. 15.2). The fold repeats part of the upper Paleocene–lower Eocene stratigraphic section, including the uppermost Aquia Formation and the lower Nanjemoy Formation (table 15.1). The dip reversal and repetition of section result in a wider than expected Paleogene outcrop belt in northeastern Virginia (Milici and others, 1963; Mixon and others, 1989).

The principal fold structure, herein named the Skinkers Neck anticline,[2] extends from the edge of the Coastal Plain near Bowling Green, Va., north-northeastward for more than 25 mi to the vicinity of Fairview Beach on the Potomac River (fig. 15.2). Our structure-contour and isopach maps for Cretaceous and Tertiary mapping horizons define a 3- to 5-mi-wide, low-amplitude, anticlinal fold that

[2]Mixon and Powars (1984) used the term "Skinkers Neck structure" in a more restricted sense for the narrow zone of faulting and folding along the northwest flank of the anticlinal structure.

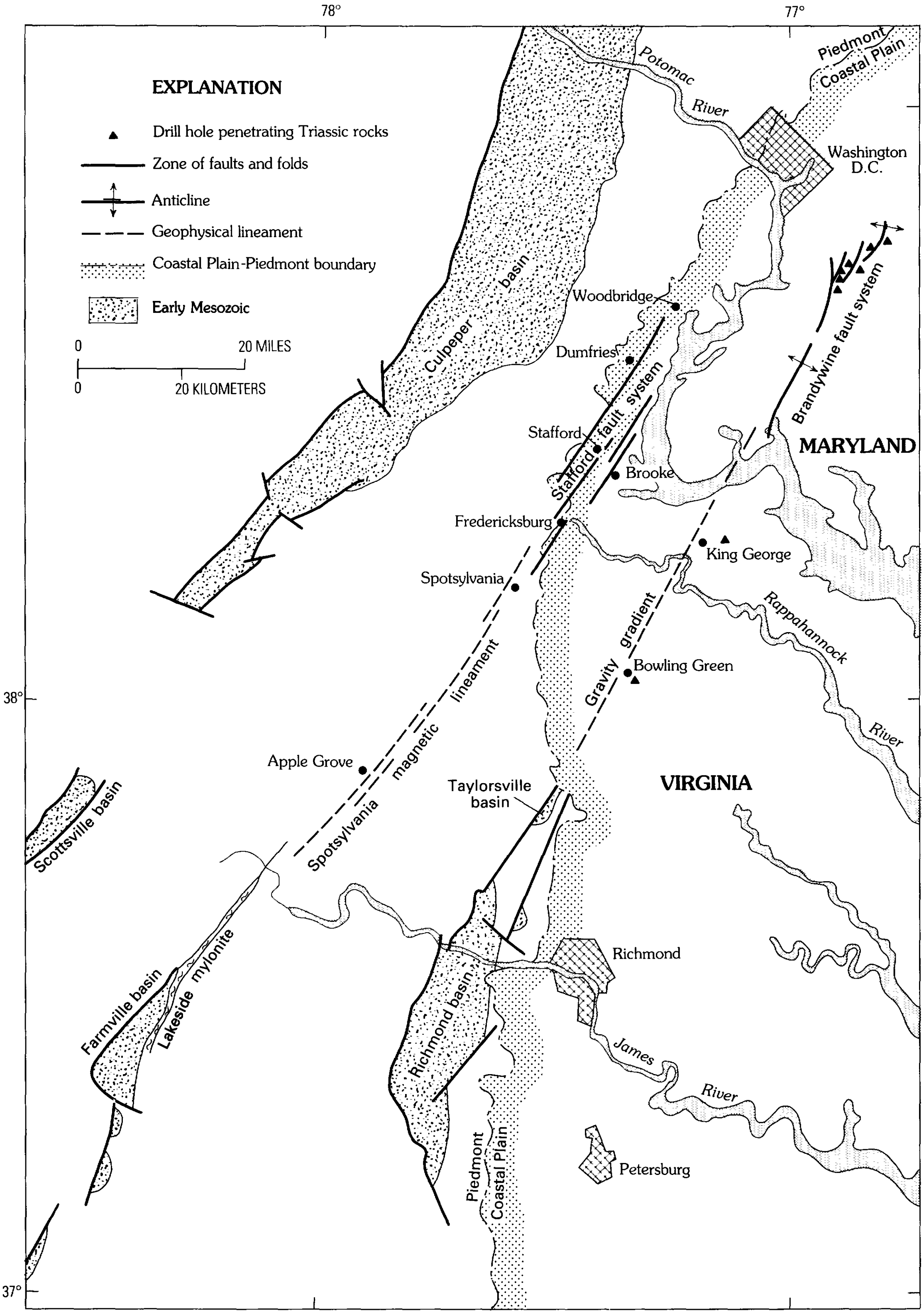

Figure 15.1. Alignment of the Stafford and Brandywine fault systems, early Mesozoic basins, and geophysical lineaments. Modified from Mixon and Newell (1982).

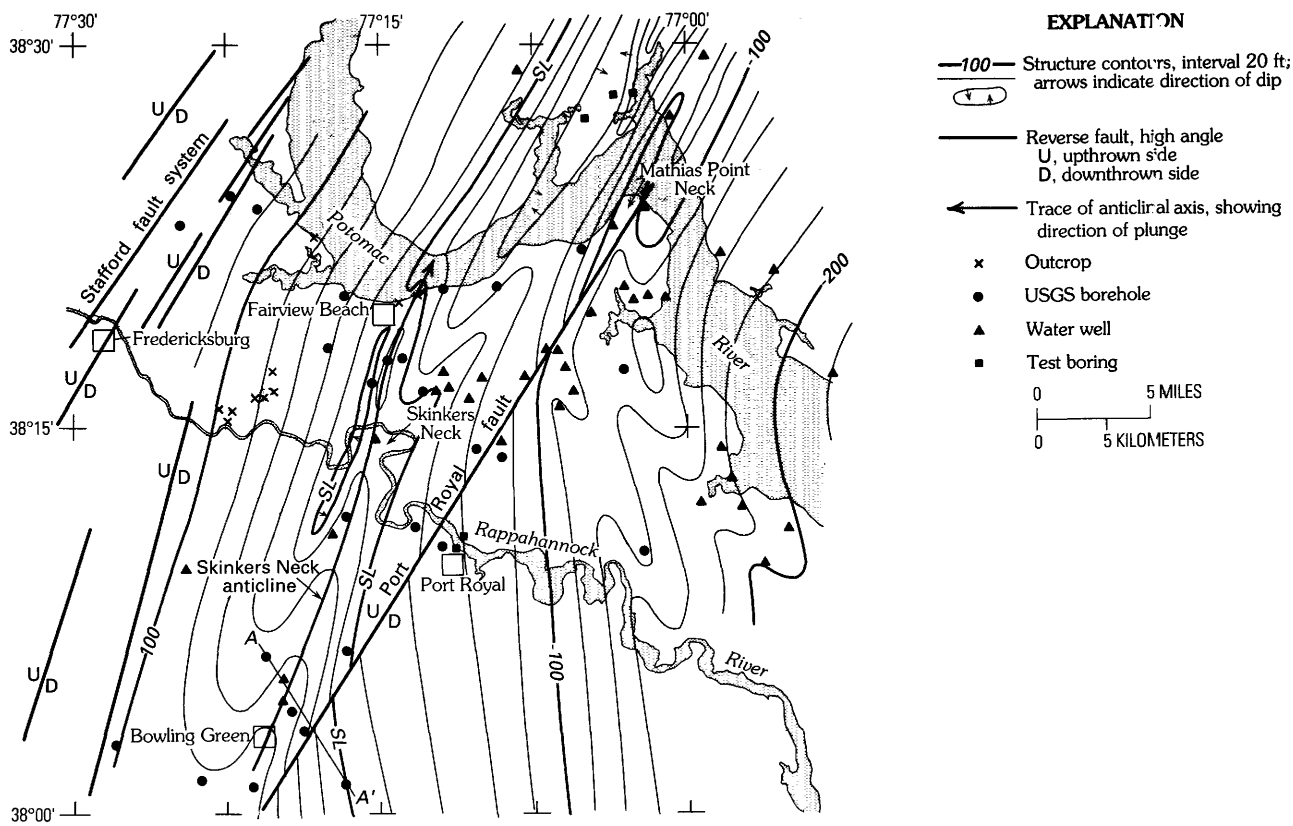

Figure 15.2. Structure contours on the top of the Paleocene beds in the area of the Skinkers Neck anticline and Port Royal fault, northeastern Virginia and southwestern Maryland Coastal Plain. Cross section *A–A'* is shown in figure 15.5. SL, sea level.

is steeper to the northwest (figs. 15.2, 15.3). The axis of the gently northeast-plunging fold crosses the Rappahannock River at Skinkers Neck, a land area bounded on three sides by a large, northward-projecting meander bend of the river (see USGS topographic map of the Rappahannock Academy 7.5-minute quadrangle). The pronounced northward deflection of the Rappahannock River at this locality is thought to reflect the impingement of the river against the uplifted, relatively resistant, clayey and silty Miocene and Eocene strata of the northwest flank of the Skinkers Neck anticline. A detailed Bouguer gravity survey (fig. 15.4; see also regional gravity map of Daniels and Leo, 1985) helps to locate more precisely the western edge of the buried Taylorsville basin in this area. The basin edge and the axis of the Skinkers Neck anticline are approximately coincident.

A northwest-southeast cross section (fig. 15.5) near Bowling Green, Va., provides some of the more detailed data presently available for interpreting the nature and timing of post-early Mesozoic deformation along the Skinkers Neck anticline. Control for the cross section consists of continuous cores and geophysical logs from four USGS coreholes and two water wells. All holes penetrated the Tertiary section and bottomed out in the upper part of the Lower and Upper(?) Cretaceous Potomac Formation.

In the cross section, the northwest side of the structural high is shown as a southeast-dipping, high-angle reverse fault. Interpretation as a reverse fault, rather than a flexure, is based on (1) the abrupt, up-to-the-southeast displacement of the Paleogene beds as seen in outcrop and in our coreholes along the Rappahannock River (Mixon and others, 1989, sheet 2), (2) seismic-reflection lines in southern Maryland (Jacobeen, 1972; Potomac Electric Power Company, 1973) showing northeast-trending reverse faults involving the crystalline basement rock and overlying Coastal Plain strata, and (3) analogy with the nearby Stafford fault system, which clearly shows reverse-fault offset of Cretaceous and Paleogene formations (Mixon and Newell, 1982; Mixon and others, 1989). Across the fault, vertical separation of the unconformity between the Potomac Formation and the directly overlying upper Paleocene Aquia Formation (fig. 15.5) is about 50 ft, whereas vertical separation of the basal beds of the Nanjemoy Formation (lower Eocene) is only about half as much. These relations and the absence of the Marlboro Clay (fig. 15.5) on the upthrown side of the fault indicate an episode of faulting, uplift, and erosion in post-Aquia and pre-Nanjemoy time. Post-Nanjemoy faulting and uplift are also indicated, but the amount of deformation is partly obscured by extensive erosion of the Nanjemoy prior to deposition of the lower

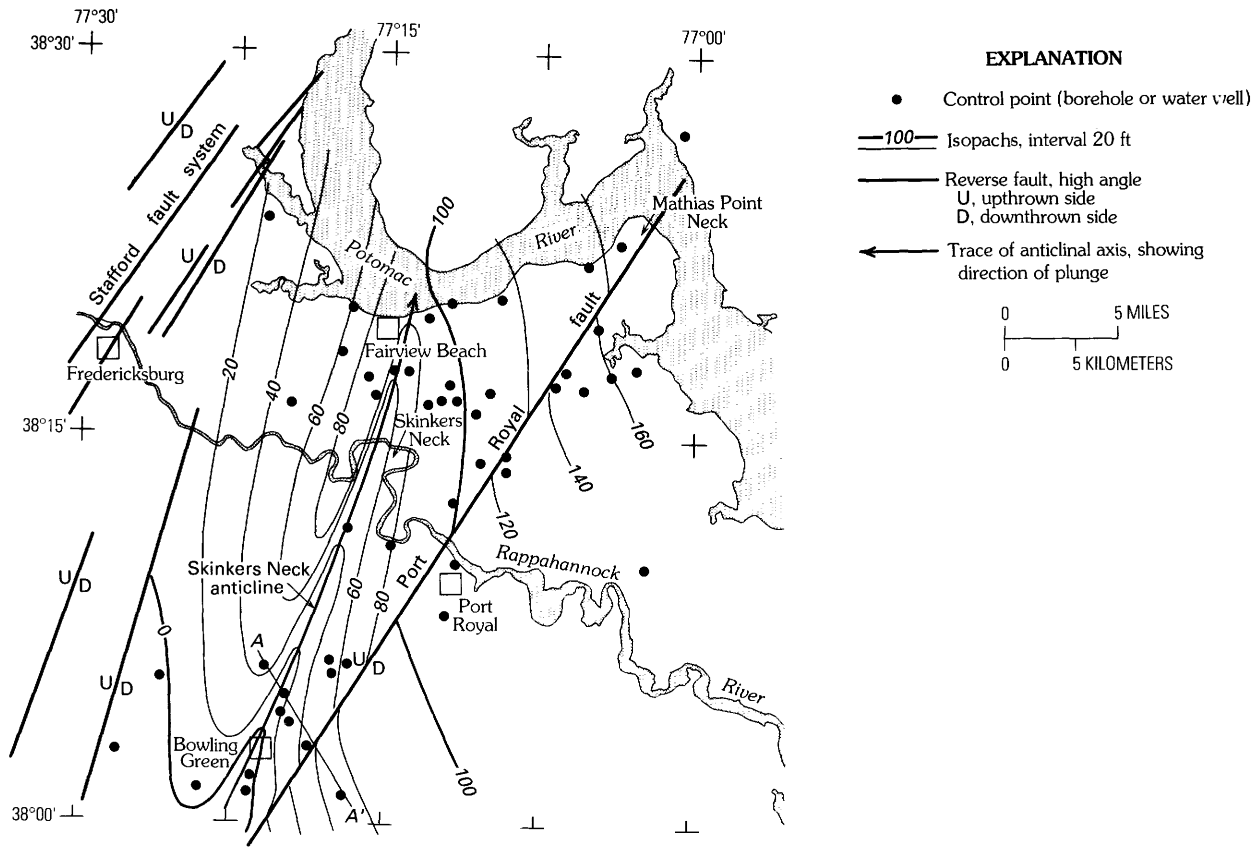

Figure 15.3. Isopachs of lower Eocene beds (Nanjemoy Formation) in the area of the Skinkers Neck anticline and Port Royal fault, northeastern Virginia. Cross section *A–A′* is shown in figure 15.5. Modified from Mixon and Powars (1984).

and middle Miocene Calvert Formation. Slight uplift and possible minor fault offset (5–10 ft) of the Calvert and Choptank Formations along the anticline are inferred on the basis of dip reversals (west dip) mapped in these units in King George County, Va. (Powars, 1987). The available drill data and surface mapping are not sufficient to clearly show the presence or absence of small-scale faulting and folding that may affect both the upper Miocene Eastover Formation and the Pliocene Yorktown Formation.

As shown in the cross section, the Nanjemoy Formation thins markedly from southeast to northwest over the crest of the Skinkers Neck anticline. Our mapping of individual members of the Nanjemoy along the line of section indicates that the thinning occurs mainly by truncation of upper and middle Nanjemoy beds by the pre-Calvert erosion surface (Mixon, Powars, and L.E. Edwards, unpub. data). The abruptly thicker Nanjemoy section in the Lonesome Gulch borehole (fig. 15.5) is the result of preservation of younger beds in the relatively downfaulted block west of the Skinkers Neck anticline. The uppermost Aquia Formation and the Marlboro Clay are also truncated across the structure. However, some of the variation in thickness of the Aquia in this area occurs in the lower part of the unit; thicker deposits fill paleotopographic lows, and deposits are thinner over paleotopographic highs. Marked thinning also occurs west of Bowling Green near the margin of the depositional basin where successively younger Aquia beds lap across the Cretaceous strata onto the crystalline basement rock (Mixon, unpub. data). The thinning of the Nanjemoy and Aquia caused by truncation of the upper parts of the units—rather than by depositional thinning of the units as a whole—suggests episodic uplift and erosion of the anticline.

PORT ROYAL FAULT

A few miles east and downdip of the axis of the Skinkers Neck anticline, the structure-contour and isopach maps of the Paleocene and Eocene beds (figs. 15.2, 15.3) delineate the Port Royal fault, a zone of low-displacement, northwest-dipping, high-angle reverse faulting and monoclinal folding that strikes southwestward from Mathias Point Neck on the Potomac River to Port Royal and central Caroline County, Va. The sense of displacement is down to the coast (down to the southeast). In King George County, where the structure is well developed, displacement of the top of the Cretaceous beds is about 50 ft and is thought to result mainly from faulting. Displacement of the top of the Paleocene section (Aquia Formation and Marlboro

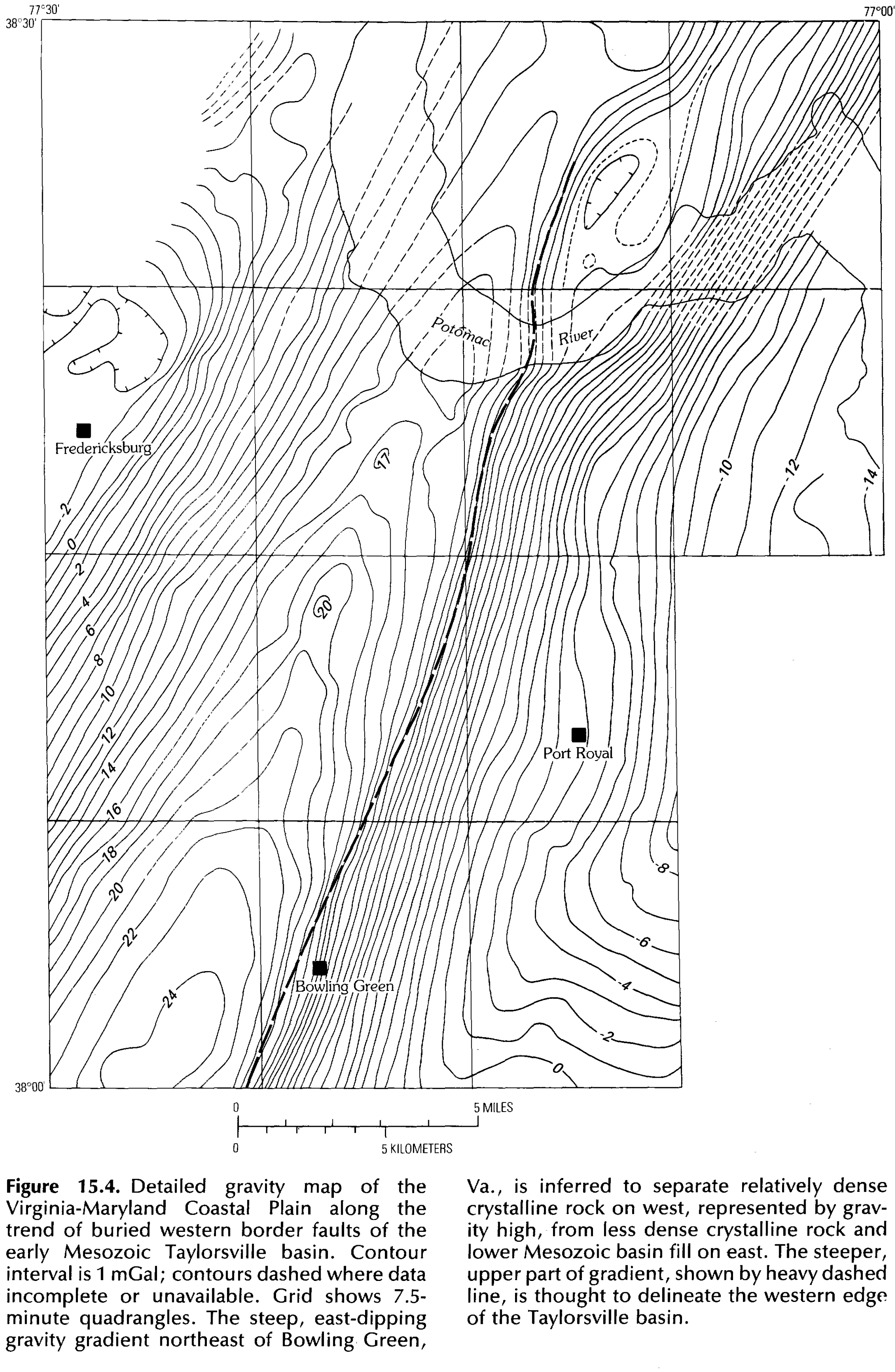

Figure 15.4. Detailed gravity map of the Virginia-Maryland Coastal Plain along the trend of buried western border faults of the early Mesozoic Taylorsville basin. Contour interval is 1 mGal; contours dashed where data incomplete or unavailable. Grid shows 7.5-minute quadrangles. The steep, east-dipping gravity gradient northeast of Bowling Green, Va., is inferred to separate relatively dense crystalline rock on west, represented by gravity high, from less dense crystalline rock and lower Mesozoic basin fill on east. The steeper, upper part of gradient, shown by heavy dashed line, is thought to delineate the western edge of the Taylorsville basin.

Clay) appears to be about 30 ft. The overlying Eocene and Miocene section is also deformed, but much of the deformation may have occurred by folding rather than faulting. Displacement decreases southwestward along the Port Royal fault and, where the structure projects into the Bowling Green cross section (fig. 15.5), appears to be

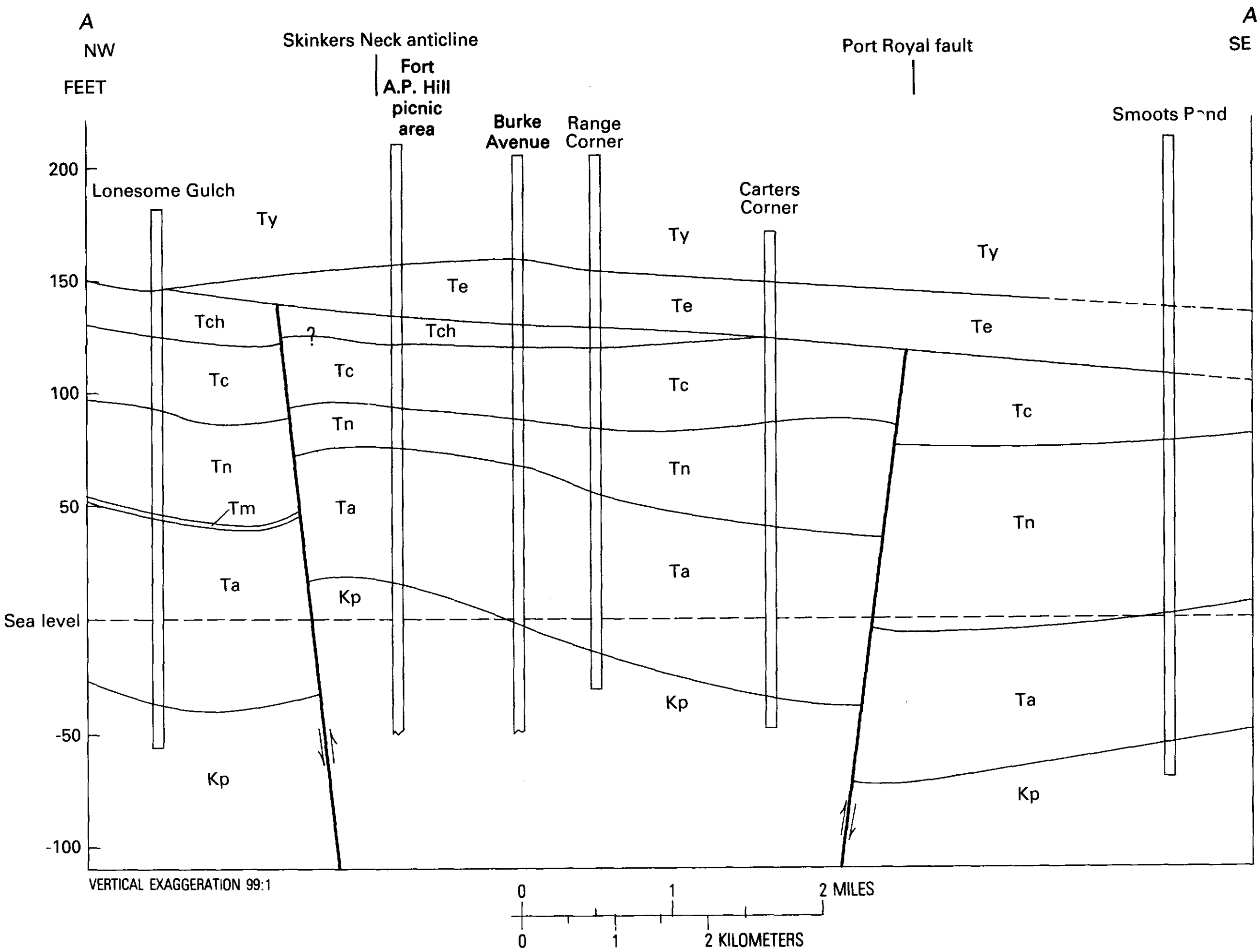

Figure 15.5. Geologic cross section of the Skinkers Neck anticline and Port Royal fault, Bowling Green area, northeastern Virginia Coastal Plain. Location of section is shown in figures 15.2 and 15.3. Ty, Yorktown Formation; Te, Eastover Formation; Tch, Choptank Formaticn; Tc, Calvert Formation; Tn, Nanjemoy Formation; Tm, Marlboro Clay; Ta, Aquia Formation; Kp, Potomac Formation.

expressed as gentle folding and very low displacement faulting. The marked thickening of the Nanjemoy Formation between the Carters Corner and Smoots Pond coreholes (fig. 15.5) is accentuated by preservation of beds on the relatively downwarped side of the structure. The absence of the Choptank Formation at the Carters Corner core site and southeastward along the line of section is a function of slight relative uplift of beds on the northwest side of the Port Royal fault and the angle of truncation of the unconformably overlying Eastover Formation. The Choptank Formation thickens to the northwest and is thickest in the structural low just west of the Skinkers Neck anticline.

SUMMARY AND CONCLUSIONS

1. En echelon, high-angle reverse faults of moderate displacement and low-amplitude folds are the two main types of structures identified in upper Mesozoic and Cenozoic deposits of the Virginia and Maryland Coastal Plain. The dominantly reverse nature of the faul'ing, the northeast trend of major structures, and the history of fault movement suggest that much of the Atlantic Coastal Plain has undergone northwest-southeast compression for at least 110 m.y. (since the Early Cretaceous).
2. Two extensive systems of late Mesozoic and Cenozoic faults have been mapped in the Coastal Plain of Virginia and Maryland. The Stafford fault system, paralleling the Fall Line in northeastern Virginia, is defined by northwest-dipping reverse faults that thrust Piedmont crystalline rocks southeastward at high angles over Coastal Plain strata. The mirror image Skinkers Neck-Brandywine zone, parallel to and about 10–15 mi east of the Stafford fault system, includes southeast-dipping reverse faults and asymmetric anticlinal folds that are steeper to the northwest.
3. Maximum reverse-fault offsets of the unconformity between the crystalline basement rock and the directly

overlying Coastal Plain strata (generally the Lower Cretaceous Potomac Formation) range from 100 to 250 ft. The amount of displacement of rock units by major reverse faults decreases upward through the Coastal Plain section, indicating recurrent movement through time. For example, offset of the base of the Paleocene Aquia Formation by any one of the Stafford faults is commonly about half as much as offset of the base of the directly underlying Lower Cretaceous Potomac Formation.

4. Analysis of outcrops and trench exposures of the Stafford faults and of our cross sections of the Skinker's Neck anticline near Bowling Green (fig. 15.5) and King George, Va. (Mixon and Powars, unpub. data), indicate episodes of faulting, folding, uplift, and erosion in the Early Cretaceous, in latest Paleocene or earliest Eocene time, in the late early(?) Eocene (post-Nanjemoy and pre-Calvert), and in the middle Miocene (post-Calvert and pre-Choptank). Some gentle arching and (or) minor faulting of the upper middle Miocene Choptank Formation, the Pliocene "upland gravel," and high-level terrace deposits of latest Pliocene age have also occurred (Mixon and Newell, 1978, 1982).
5. The aligned Stafford fault system, Spotsylvania magnetic lineament, Lakeside mylonite zone, and border faults of the early Mesozoic Farmville basin trend (fig. 15.1) are believed to mark a long, narrow zone of ductile and brittle deformation that extends from the outer Piedmont northeastward into the Inner Coastal Plain of Virginia. In northeastern Virginia, relative down-to-the-southeast displacement of the crystalline basement rocks and the Coastal Plain deposits along the Stafford fault system controls the position of the Fall Line and the northeast-trending reach of the Potomac River estuary. We believe that the zone of deformation and associated downwarping of the inner edge of the Coastal Plain continues northeastward through Maryland and northern Delaware along the margin of the Chesapeake-Delaware embayment.
6. Several workers have discussed the close association of the early Mesozoic border faults of the Richmond and Taylorsville basins with the late Paleozoic Hylas mylonite zone and have cited these structures as an example of preexisting fault control for Mesozoic basins (Bobyarchick and Glover, 1979; Swanson, 1986). Our mapping in the Coastal Plain along the strike of the Hylas zone and adjacent rift basins has identified Cretaceous and Tertiary structures, including the Skinkers Neck anticline and related faults, that are superposed on buried border faults of the Taylorsville basin, present in the subsurface in this area. These relationships indicate continued release of compressional stress along the Hylas zone and (or) reactivation and reversal of movement on the extensional faults of the early Mesozoic basins. The deformational history of the Coastal Plain strata in the Skinkers Neck area indicates the continuation of a compressional stress field through much of late Mesozoic and Cenozoic time. The Port Royal fault and fold, a few miles east and downdip from the Skinkers Neck anticline, probably also represent reactivation of early Mesozoic faults present beneath the Coastal Plain cover.

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16. National Water-Quality Assessment Program Activities on the Delmarva Peninsula in Parts of Delaware, Maryland, and Virginia

By Patrick J. Phillips,[1] Robert J. Shedlock,[1] *and* Pixie A. Hamilton[2]

INTRODUCTION

The Delmarva Peninsula is one of three ground-water project areas in the U.S. Geological Survey's pilot National Water-Quality Assessment (NAWQA) Program. The pilot program will test and refine concepts and approaches for conducting a full-scale national water-quality assessment (Hirsch and others, 1988). At present (1988), the NAWQA Program consists of seven pilot projects: three that focus on ground water and four that focus on surface water. The goals of a full-scale NAWQA Program would be to provide a consistent national description of current water-quality conditions, to define long-term trends in water quality, and to relate observed water-quality conditions and trends to natural and human factors. This report describes the main objectives and approaches of the pilot ground-water NAWQA project in the Delmarva Peninsula. Because the range of hydrogeologic and geomorphic settings on the Delmarva Peninsula is similar to that in other areas along the U.S. Atlantic Coastal Plain, results from this project are expected to have broad interest and applicability.

The Delmarva Peninsula lies east of Chesapeake Bay and west of the Atlantic Ocean and Delaware Bay; it includes most of Delaware and parts of Maryland and Virginia (fig. 16.1). The total land area of the peninsula is about 6,050 mi^2. Agriculture is the largest single land use in the study area, and it accounted for about 48 percent of the total land area in 1972–76 (U.S. Geological Survey, 1979a, b; 1981a, b). The remaining land was woodland (31 percent), wetland (13 percent), urban land (7 percent), and barren land (1 percent). The study area is in the Coastal Plain physiographic province and is relatively flat; most of the study area is less than 100 ft above sea level. The peninsula contains a central upland bordered by broad lowlands. These lowlands are fringed by tidal wetlands near the coastlines. Part of the central upland consists of broad flats separated by deeply incised streams, whereas another part is poorly drained and hummocky.

The aquifer system consists of a sequence of nine confined sand aquifers that dip to the east or southeast (Cushing and others, 1973). This sequence is overlain by a regionally extensive surficial aquifer that is mostly unconfined. The surficial aquifer is the main source of recharge to the underlying confined aquifers, which, with one exception, are truncated by the surficial aquifer.

The three NAWQA ground-water pilot projects have five major activities: (1) analysis of available information; (2) low-density sampling throughout the study areas for a wide range of constituents (referred to as a regional survey); (3) targeted sampling at a higher density in several smaller areas for selected groups of constituents; (4) supplemental geohydrologic and geochemical measurements; and (5) establishment of a network of wells for long-term sampling.

ANALYSIS OF AVAILABLE INFORMATION

Existing data on ground-water quality and ancillary information have been analyzed to provide an initial assessment of regional ground-water quality and to test methods for investigating water quality in relatively large regions, such as the Delmarva Peninsula. Hamilton and others (1989) described the sources, availability, accessibility, and geographic distribution of existing water-quality data for the study area and provided a preliminary analysis of these data.

REGIONAL SURVEY

One of the objectives of the regional survey is to describe ground-water quality, both areally and with depth, for the principal aquifers in the study area. A further objective is to provide a basis for improving the understanding of water-quality variations and problems in the study area, especially with regard to differences in land use, soils, and hydrogeologic setting. The samples collected for the regional survey will be analyzed for all major cations and anions, numerous trace metals and other trace elements,

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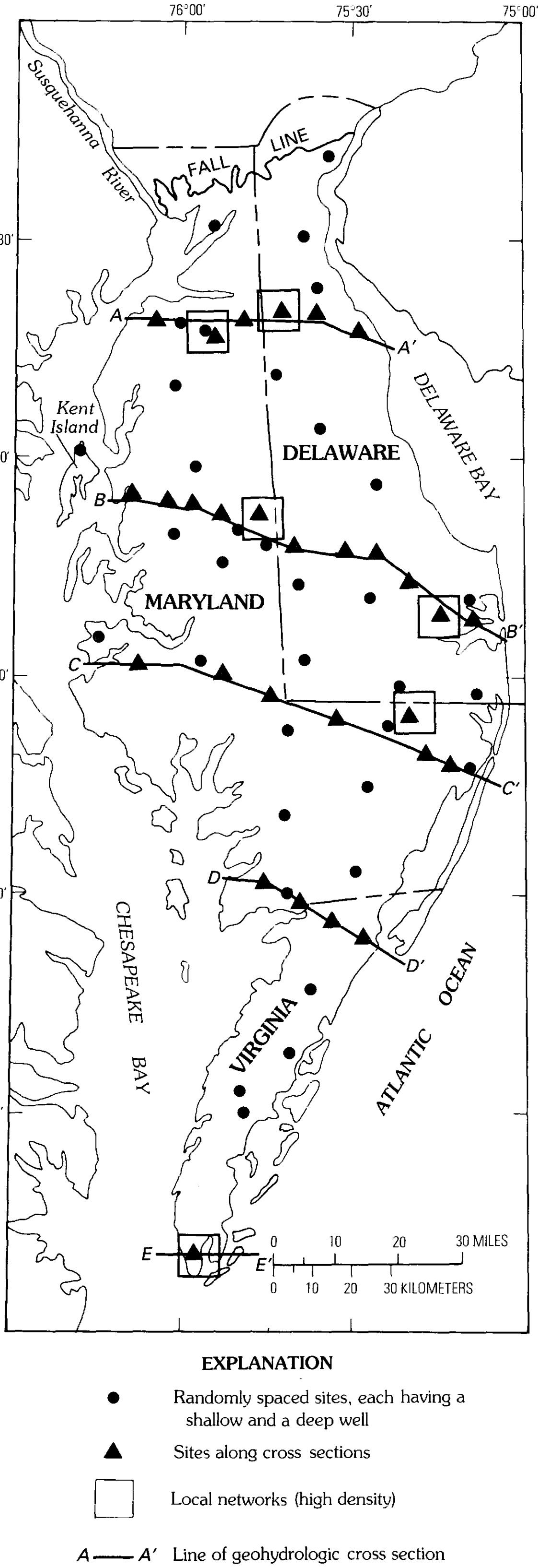

Figure 16.1. Network of wells in the surficial aquifer planned for the Delmarva pilot project.

radionuclides, and organic constituents. The organic constituents include dissolved organic carbon, a suite of volatile organic compounds, and selected pesticides. Sampling sites will be distributed throughout each aquifer and will be unbiased toward known or suspected problem areas. The well network (fig. 16.1) will consist of three sets of wells: (1) a spatially stratified network of wells in the surficial aquifer; (2) a set of wells in the surficial aquifer along east-west cross sections; and (3) a set of wells in each of the nine confined aquifers (not shown in fig. 16.1).

The spatially stratified network of the surficial aquifer is intended to provide a representation of water quality in the Delmarva Peninsula at the broadest scale. Thirty-five sites were selected randomly throughout the peninsula; each site has a shallow and a deep well. Wells in the second set in the regional survey are along several lines of section that trend roughly east-west and extend from Chesapeake Bay to Delaware Bay or to the Atlantic Ocean (fig. 16.1). These cross sections will provide approximately 30 additional sampling sites in the surficial aquifer in areas representative of different hydrogeologic and geomorphic settings. These cross sections were chosen primarily to depict changes in the thickness and hydrogeologic characteristics of the surficial aquifer system and to show regional profiles of the water table. The third set of wells in the regional survey network is in the confined aquifer system. Three wells will be chosen for each aquifer; one well will be in the subcrop zone, and two other wells will be located progressively farther downdip and, presumably, farther downgradient.

The results of the regional survey are expected to allow broad comparisons between deep and shallow ground water in the surficial aquifer and between waters in the confined aquifers and those in the surficial aquifer. The total number of wells to be sampled during the regional survey will range from 120 to 150. Supplemental sampling also may be done to collect samples from various land-use areas, hydrogeologic and geomorphic settings, and classes of wells that may have been underrepresented in the regional survey.

TARGETED SAMPLING

The regional survey wells will be supplemented by wells in small watersheds in a variety of hydrogeologic and geomorphic settings (fig. 16.1). This set of wells, known as the high-density network, consists of wells placed along transects in watersheds of a few square miles or less to investigate water-quality patterns along local ground-water flow paths. Samples collected from this network will be analyzed for major ions, nutrients, selected pesticides, and selected radionuclides. Because the watersheds for these transects are located along the major east-west cross sections, they should provide a link between local and regional patterns of water quality.

Results of water-quality sampling from the high-density network will be used for several purposes. First, they will provide a measure of variations in water quality in more detail than is possible with the regional survey network. Second, these results will be used to relate the distribution of water-quality constituents to ground-water-flow patterns and to local variations in topography, soils, and land use.

Finally, water-quality data collected at wells in the high-density network and at the regional-survey sites in agricultural areas will be used to study the distribution of agricultural chemicals in ground water in various hydrogeologic settings. Information on current and past crop-rotation practices and on use of specific pesticides will help to determine which pesticides' distribution will be analyzed. Temporal variations in the concentrations of agricultural chemicals in ground water will be studied by sampling selected wells before, during, and after the growing season for 1 year or more.

GEOHYDROLOGIC AND GEOCHEMICAL MEASUREMENTS

Supplemental geohydrologic and geochemical measurements will be made on samples from both the regional and local networks to study the relations among patterns of ground-water flow and water quality. These measurements will be made to study recharge-discharge relations in the shallow ground-water system, to determine differences in environmental isotopes in different parts of the aquifer system, and to investigate mineral-water interactions that may affect water quality. These efforts will provide physical and chemical data on human and natural effects on ground water.

Recharge-discharge relations in the shallow ground-water system will be studied by observing temporal variations in water-table profiles in the high-density network and by using these profiles as the basis for cross-sectional simulations of the shallow ground-water system. The simulations will be used to investigate the horizontal and vertical dimensions of shallow ground-water-flow systems. The investigations of recharge-discharge relations are expected to help in understanding the relations between water quality and physical features of the land in more detail than can be obtained by use of traditional maps of surficial geology, soils, and land use.

Environmental isotopes have been used in the last 30 years to study recharge-discharge relations in aquifer systems, particularly in shallow flow systems. (See Back and Freeze, 1983, p. 294–297, for a short review of the subject.) Samples will be collected from wells in the subcrop areas of the confined aquifers and from selected wells in the high-density network. These samples will be analyzed for tritium and the stable isotopes of oxygen and hydrogen. Results from this effort will complement simulations of vertical flow patterns in the shallow ground-water system. The results will be used primarily to determine the vertical penetration of local flow patterns so that the potential for anthropogenic substances to migrate into deep ground-water supplies can be assessed.

Chemical analyses of ground-water samples from both the regional survey and targeted sampling efforts will be used to identify mineral-water interactions that influence the chemical composition of ground water. In addition, petrographic observations and chemical analyses of soils and aquifer material will help to explain geochemical processes that may influence the transport and fate of trace elements and manmade organic constituents.

ESTABLISHMENT OF A LONG-TERM SAMPLING NETWORK

A long-term sampling network will be established to evaluate temporal trends in ground-water quality in the study area. Collection of water-quality data from these wells will continue after the completion of the Delmarva pilot project. This network will include wells representing waters of different ages (where "age" indicates the time since water entered the ground-water-flow system), hydrogeologic and geomorphic settings, and land uses. The first step in determining the long-term sampling network will involve assigning the wells sampled during the study to groups based on the estimated age of the water in the well. These ages will be determined from studies of recharge-discharge relations, environmental isotopes, and hydrochemical water type. Although sites will be chosen to reflect a range of ground-water ages, most of the wells in the long-term sampling network will tap aquifers containing relatively young waters.

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17. Campanian to Quaternary Depositional Sequences in the Baltimore Canyon Trough and Their Relations to Deposits underlying the Middle U.S. Atlantic Coastal Plain

By C. Wylie Poag[1]

INTRODUCTION

The Baltimore Canyon trough is a large depositional basin seaward of the Middle Atlantic Coastal Plain in Virginia, Maryland, Delaware, and New Jersey (fig. 17.1). Several compilations of geologic, paleontologic, and seismostratigraphic data from the trough have been published in the last few years (various papers in Poag, 1985b; Poag and others, 1987; Ross and Haman, 1987; Sheridan and Grow, 1988).

For this region, 55 boreholes and multichannel seismic-reflection profiles collected along 10,000 km of tracklines constitute the principal foundation for the stratigraphic and depositional framework (Poag, 1985a, 1987a; Poag and Valentine, 1988). Poag (1987a) illustrated and described this framework along a chronostratigraphic section that stretches 700 km from the Virginia Coastal Plain to the lower part of the Atlantic Continental Rise in the Hatteras basin (Deep Sea Drilling Project Site 603). The chief feature of this section is a succession of widespread, unconformity-bounded (allostratigraphic) depositional units (Poag, 1897b).

The stratigraphic position of each unit conforms remarkably well with the supercycles of the sequence-stratigraphy model (Vail and others, 1977; Haq and others, 1987; Greenlee and others, 1988). The bounding unconformities can be recognized on the basis of biostratigraphic and lithostratigraphic discontinuities in outcrops and boreholes (Ward, 1984; Olsson and others, 1987; Olsson and Wise, 1987; Poag and Low, 1987; Poag, 1989; Miller and others, 1990) and (or) by angular relations between reflections on seismic profiles (Vail and others, 1977; Poag and Schlee, 1984; Poag and Mountain, 1987). Stratigraphic successions documented at the offshore boreholes can be extrapolated for hundreds of kilometers along the extensive network of interconnecting seismic profiles (Schlee, 1981; Poag and Mountain, 1987; Poag and Sevon, 1989). Recent studies of planktonic microfossils in Coastal Plain deposits (Olsson and others, 1987; Poore and Bybell, 1988; Poag, 1989; Poag, unpub. data) allow detailed correlation between units under the Coastal Plain and offshore allostratigraphic units.

Using this stratigraphic and depositional framework, I have mapped the structural configuration, thickness, and distribution patterns of each allostratigraphic unit (Poag, 1987a; Poag and Mountain, 1987; Poag and Sevon, 1989). The isopach maps are particularly useful for locating depocenters, tracing the principal routes of sediment dispersal, and determining the dominant provenance of terrigenous components in each allostratigraphic unit. The following discussion is based on a limited selection of isopach maps and seismostratigraphic sections for a few of these allostratigraphic units.

CAMPANIAN ALLOSTRATIGRAPHIC UNIT

I will set the stage for a discussion of the Cenozoic depositional units by first describing the salient features of the Upper Cretaceous deposits. Seaward of what now are the Middle Atlantic States of Virginia, Maryland, Delaware, and New Jersey, the Upper Cretaceous (Campanian) margin featured a broad, gently sloping continental shelf; it was about 200 km wide and was covered by a blanket of terrigenous to marine and hemipelagic deposits less than 100 m thick (Poag, 1987a; Poag and Sevon, 1989). In contrast, seaward of what now is Long Island, the Campanian shelf was the site of several deltaic prisms thicker than 200 m. The greatest sediment thickness by far, however, accumulated on the continental rise as submarine fans >1,000 m thick near the base of the Long Island platform (fig. 17.2). The pattern of isopachs is evidence that the Campanian sediments came mainly from the north via the ancient Connecticut River and an undetermined river (or rivers) in eastern Massachusetts (which drained the New England Highlands; Poag and Sevon, 1989). Similar relations characterized Maastrichtian deposition.

During the Paleocene and early Eocene, however, the dominance of northern sediment sources diminished, as the central Appalachian Highlands became a major source

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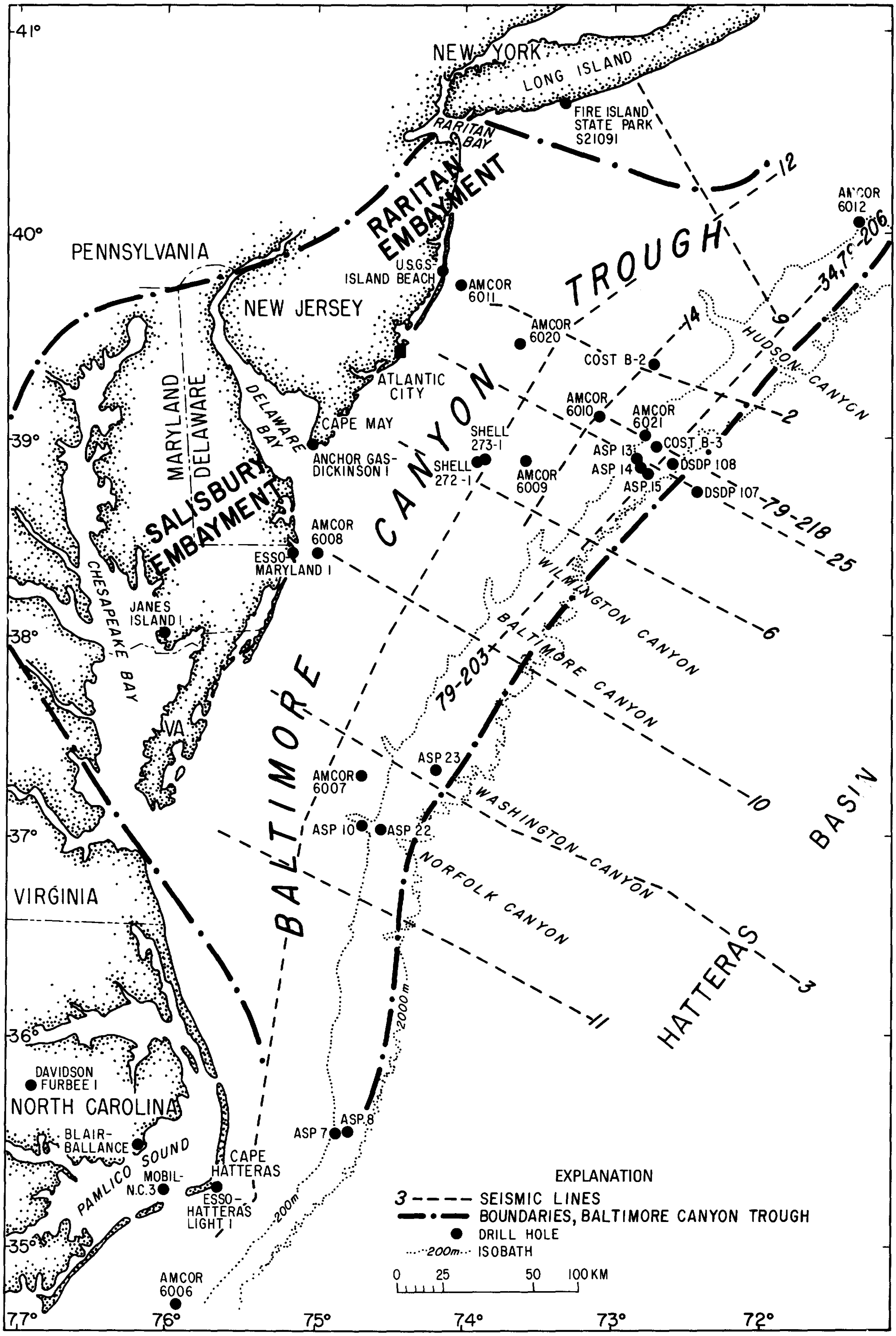

Figure 17.1. Locations of principal physiographic and geologic features, selected boreholes, and tracklines along which multichannel seismic-reflection profiles were collected in the Baltimore Canyon trough. AMCOR, Atlantic Margin Coring Project; ASP, Atlantic Slope Project; COST, Continental Offshore Stratigraphic Test; DSDP, Deep Sea Drilling Project.

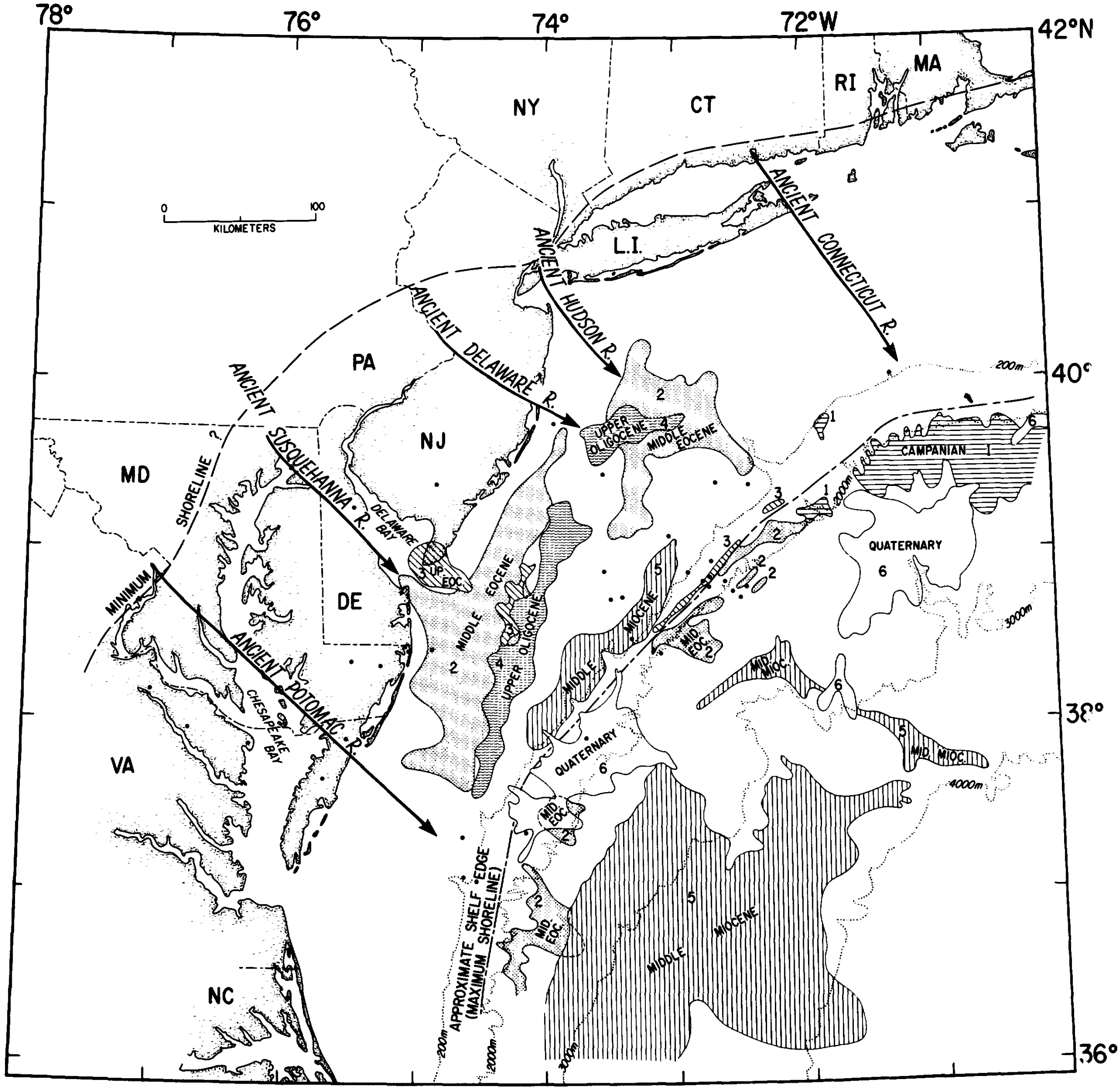

Figure 17.2. Selected principal depocenters of the Baltimore Canyon trough and northern Hatteras basin during the Campanian to Quaternary. Depocenters are numbered (1–6) in chronological order from oldest (Campanian) to youngest (Quaternary). Principal drainage systems are indicated by heavy arrows; solid circles mark key boreholes; fine dotted lines are bathymetric contours indicating meters below present sea level. Note temporal and spatial shifts in sediment source between western (central Appalachian Highlands) and northern (Adirondack and New England Highlands) terrains. Depocenters for other Cenozoic sedimentary sequences are not illustrated in order to avoid cluttering the figure.

terrain. In addition, carbonate deposits became more abundant in shelf and deeper lithotopes (Poag, 1985a, 1987a).

MIDDLE EOCENE ALLOSTRATIGRAPHIC UNIT

By the end of the middle Eocene, a significant rise in relative sea level had trapped most continent-derived sediments on the continental shelf; this system was very different from the earlier Cenozoic and Late Cretaceous depositional systems (Poag and Sevon, 1989). In the study area, the most striking sedimentary feature of the middle Eocene was an elongate (~200-km-long), current-dominated delta (fig. 17.2) reminiscent of the modern Amazon Delta (by analogy, it was probably a submarine delta). Its >400-m-thick depocenter is directly offshore from what now is Delaware Bay. Thus, by middle Eocene time, the dominant provenance of terrigenous detritus had shifted to the west (to the central Appalachian Highlands), and sediment moved seaward via the ancient Susquehanna and Delaware Rivers, which took more direct routes eastward to the Atlantic than they do today (Poag and Sevon, 1989).

The Adirondack Highlands appear to have been an important, though secondary, sediment source, as evidenced by a large bird-foot delta (>200 m thick), which built out seaward of the ancient Hudson River mouth (fig. 17.2). Scattered, thin wedges (~200 m thick) of chiefly carbonate sediments formed slope aprons and submarine fans on the upper continental rise (Poag, 1987a).

UPPER EOCENE ALLOSTRATIGRAPHIC UNIT

The present distribution of upper Eocene sediments is quite limited compared to that of most other Cenozoic units (fig. 17.2). A small, relatively thin depocenter (~100 m thick) is centered around what now is Cape May, N.J.; it was penetrated by the Anchor Gas-Dickinson No. 1 well (Brown and others, 1972; Poag, 1985a). A second small depocenter appears to have overlapped part of the seaward slope of the underlying middle Eocene delta, and an elongate, thin wedge of upper Eocene sediment is present at the shelf edge off New Jersey. This patchy distribution is partly a product of widespread erosion that took place during the early and late Oligocene.

UPPER OLIGOCENE ALLOSTRATIGRAPHIC UNIT

Lower Oligocene strata in the study area are too poorly known to map, but upper Oligocene strata have been identified in several outcrops (Ward, 1984; Poag and Ward, 1987) and boreholes (Poag, 1985a, 1989; Poore and Bybell, 1988) on the Coastal Plain and offshore. The principal depocenter (fig. 17.2) is marked by an elongate, narrow prism (~200 km long, >300 m thick), which accumulated along the seaward slope of the underlying middle Eocene delta. The chief source of sediment remained the central Appalachian Highlands (Poag and Sevon, 1989), as the ancient Susquehanna drainage system was dominant. Also, as in the middle Eocene, a secondary depocenter developed off the mouths of the ancient Hudson and Delaware Rivers (fig. 17.2). Upper Oligocene depocenters are not yet known from the Hatteras basin, but thin, scattered patches of upper Oligocene strata are present on the outer shelf and slope, where they consist of chiefly pelagic carbonate deposits (Poag, 1985a; Poag and others, 1987). The patchy distribution of the upper Oligocene allostratigraphic unit appears to be, like that of the upper Eocene unit, the result of erosion.

MIDDLE MIOCENE ALLOSTRATIGRAPHIC UNIT

Progradation of middle Cenozoic depositional units toward the shelf edge continued through the early and middle Miocene. By the end of the middle Miocene, a shelf-slope profile similar to the modern one had been constructed (Poag, 1987a). The middle Miocene rate of sediment accumulation was an order of magnitude greater than that during any preceding interval of the Cenozoic, and this accelerated deposition formed a huge shelf-edge depocenter (>1,300 m thick) due east of the mouth of the ancient Susquehanna River (fig. 17.2; Poag and Sevon, 1989). Thick (>1,400 m) middle Miocene deposits also accumulated on the continental rise in large submarine fans and contourite drifts, which contain predominantly terrigenous constituents (Mountain and Tucholke, 1985; Tucholke and Mountain, 1986). The ancient Delaware and Potomac river systems were secondary sediment-dispersal routes at this time, but northern source terrains (Adirondack and New England Highlands) appear to have been less important.

UPPER MIOCENE TO QUATERNARY ALLOSTRATIGRAPHIC UNITS

Continued seaward progradation during the late Miocene, Pliocene, and Quaternary built relatively thick (300–400 m), elongate depocenters that perched along the shelf edge (Poag, 1987a; Poag and Sevon, 1989), but the bulk of deposition (>800 m) during this time interval took place on the continental rise. Two broad depocenters (~100 km wide) on the continental rise contain >600 m of upper Miocene to Quaternary sediments (fig. 17.2). These sediments appear to have been channeled mainly through submarine canyons that incised the shelf edge. Large submarine fans and contourite drifts continued to characterize much of the deep-water deposition in the Hatteras basin during this time (Tucholke and Mountain, 1986). Both western (central Appalachian Highlands) and northern (Adirondack and New England Highlands) source terrains contributed large quantities of sediment during the Quaternary.

SUMMARY AND CONCLUSIONS

In summarizing, let me re-emphasize that the Upper Cretaceous and Cenozoic deposits of the Middle Atlantic Coastal Plain accumulated along the western near-shore margin of a basin whose principal depocenters (for almost every allostratigraphic unit) are well offshore from the present Coastal Plain. The positions of these depocenters appear to be related directly to (1) the relative position of sea level, (2) the rate of sediment accumulation, and (3) the provenance of the dominant sedimentary constituents for each allostratigraphic unit. These three properties have been subject to large-scale shifts through the last 80 m.y. Upper Cretaceous (Campanian-Maastrichtian) deposits are thickest on the northern part of the continental rise, whereas the continental shelf off the Middle Atlantic Coastal Plain received only a relatively thin blanket of Upper Cretaceous sediment.

By middle Eocene time, however, a major rise in relative sea level trapped most terrigenous detritus on the continental shelf and allowed central Appalachian drainage systems to build a large submarine delta off Delaware Bay. A subsequent major pulse of terrigenous deposition brought central Appalachian detritus to unusually thick middle Miocene depocenters at the shelf edge and on the continental rise. During the rest of the Cenozoic, the continental rise hosted the thickest depocenters, and sediment was supplied in large volumes by northern sources, as well.

During the 80 m.y. of Campanian to Quaternary time, the shelf edge was near its present location (fig. 17.2); its position was controlled by a buried Upper Jurassic reef escarpment whose original relief was 1.5–2.0 km (Poag, in press). In contrast, the shoreline (fig. 17.2) shifted laterally 200 km or more during this time interval, reaching nearly to (or beyond?) the shelf edge during the lowest stands of sea level (for example, during Quaternary glaciations). During lowstands and the early stages of subsequent marine transgressions, erosion was widespread and often severe, modifying significantly the original distribution patterns and thicknesses of the allostratigraphic units (especially where the deposits were already thin) and creating their unconformable boundaries.

It is essential that the regional perspective provided by the depositional and erosional history of the Atlantic offshore basins be incorporated into any interpretation of the various geologic and paleontologic aspects of the Atlantic Coastal Plain.

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18. Uppermost Mesozoic and Cenozoic Geologic Cross Section, Outer Coastal Plain of Virginia[1]

By D.S. Powars,[2] R.B. Mixon,[2] *and* Scott Bruce[3]

INTRODUCTION

Since 1985, a cooperative research effort has existed between the U.S. Geological Survey (USGS) and the Virginia State Water Control Board (VA–SWCB) to drill and study five coreholes in the Outer Coastal Plain of Virginia. The locations of the five coreholes—Dismal Swamp, Fentress, Kiptopeke, Exmore, and Jenkins Bridge—and of boreholes to basement from previous investigations are shown in figure 18.1. All these locations except Exmore are VA–SWCB research stations where individual wells monitor each aquifer penetrated.

The Dismal Swamp and Fentress coreholes were continuously cored from the surface downward into Lower Cretaceous strata. The Exmore corehole was continuously cored from the surface downward into middle Eocene strata dominated in the lowermost 50 ft by reworked Lower Cretaceous boulders. The Jenkins Bridge corehole was continuously cored from the surface down to 490 ft and from 1,200 to 1,333 ft; it bottomed in Lower Cretaceous strata. The Kiptopeke corehole was continously cored from the surface down to 1,330 ft, where it just reached middle Eocene strata similar to those penetrated in the Exmore corehole, and from 1,733 to 1,753 ft; it bottomed in Lower Cretaceous strata.

This report is a preliminary summary of ongoing lithostratigraphic investigations and includes biostratigraphic data from USGS colleagues on dinoflagellates (Lucy E. Edwards), Foraminifera (C. Wylie Poag and Thomas G. Gibson), Tertiary mollusks (L.W. Ward), Cretaceous mollusks (Norman F. Sohl), diatoms (George W. Andrews), pollen (Ronald J. Litwin), and ostracodes (Gregory S. Gohn). This report also includes lithologic (D. Powars, site log) and geophysical data from corehole MW4–1 in the City of Chesapeake (fig. 18.1). That corehole penetrated Cretaceous to Pleistocene strata and was continuously cored from 350 to 996 ft depth. The cored interval penetrated Lower Cretaceous to Miocene strata and is part of an aquifer storage and recovery project for the City of Chesapeake.

Before these holes were drilled, knowledge of the subsurface of the Outer Virginia Coastal Plain was derived primarily from studies of water-well cuttings and geophysical logs (Cederstrom, 1945a, b; Hansen, 1967; Sinnott and Tibbitts, 1968; Brown and others, 1972; Cushing and others, 1973; Meng and Harsh, 1988). The only available cores were from selected intervals in the Taylor #1 oil test hole and the Moores Bridge well (labeled NOR–T–12 in fig. 18.1) and from selected intervals at a few wells drilled by the VA–SWCB.

The new core data enable us for the first time to establish well-defined standard sections for Virginia's Outer Coastal Plain, which is divided by the mouth of the Chesapeake Bay into two areas: the Eastern Shore and southeastern Virginia. Lithologic, paleontologic, and geophysical data from these coreholes are being correlated with data from other wells in order to define the stratigraphic and structural configuration of this area.

Acknowledgments

Special thanks are given to the drillers who made this study possible, including the USGS team of Dennis W. Duty and Donald G. Queen and the VA–SWCB team of John P. Greason and Jay A. Owens. We thank Sherman Stairs, manager of the Eastern Shore of Virginia National Wildlife Refuge; Jim Owens, manager of the Dismal Swamp National Wildlife Refuge; military personnel at the Naval Auxiliary Landing Field Fentress; Dr. Robert E. Baldwin, chief scientist of the Eastern Shore Agriculture Experiment Station; and M. Carter Davis, Jr., landowner at the Jenkins Bridge site, for their cooperation and access to their properties. Finally, Steve Schindler and Greg Gohn's assistance in the scientific review was very helpful and is greatly appreciated.

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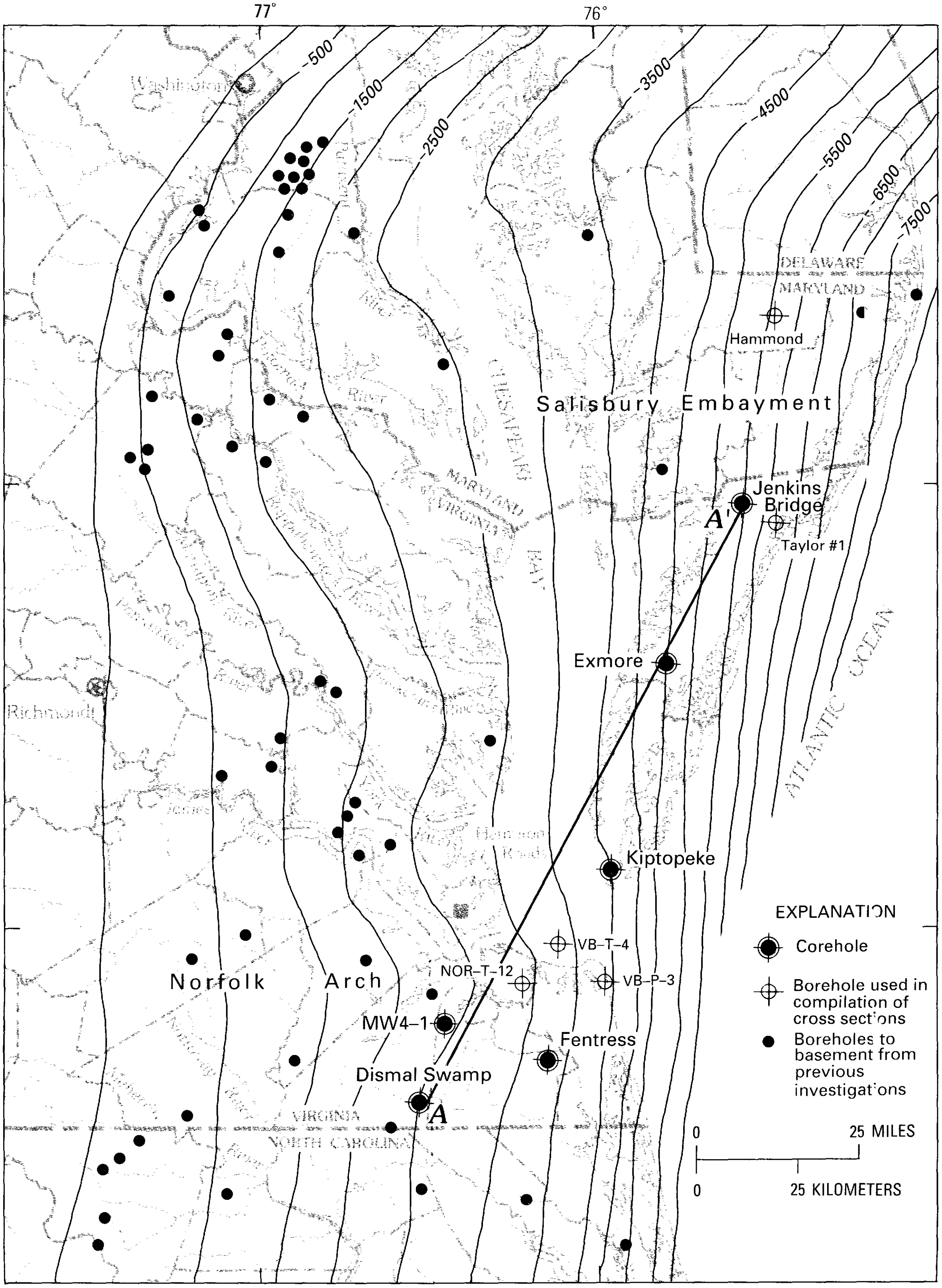


Figure 18.1. Structure contours on the top of pre-Cretaceous basement in the Virginia Coastal Plain. Contour interval equals 500 ft. Compiled from Brown and others (1972), Hansen and Edwards (1986), Bayer and Milici (1987), Wilkes and others (1989), and unpublished data of the Virginia State Water Control Board and the U.S. Geological Survey.

GEOLOGIC SETTING

As shown in figures 18.1 and 18.2, the Outer Coastal Plain of Virginia includes parts of two major structural features: the tectonic downwarp known as the Salisbury embayment on the north and the Norfolk arch on the south. The arch and basin configuration is typical for the Atlantic Coastal Plain and is generally characterized by stratigraphic thinning or by truncation of Cretaceous and Tertiary formations across the arches.

In cross section *A–A′* (fig. 18.3), the northeast flank of the Norfolk arch is defined by the northeast-sloping basement surface and the southward stratigraphic thinning of Cretaceous and Tertiary strata across the arch. The southern end of the cross section is farther west than the northern end and, therefore, is updip structurally. Also, the NOR–T–12 well, the VB–T–4 test boring and the Kiptopeke corehole (see figs. 18.1 and 18.3) are several miles east of the cross section line and, therefore, are downdip structurally.

Cederstrom (1945b) postulated a fault zone that generally trends west from Hampton Roads along the James River to the Fall Line because of his interpretation that Eocene beds (including what we now subdivide into Paleocene and Eocene strata) abruptly thicken north of the James River. Our data suggest that the water-covered area at the mouths of Chesapeake Bay and the James River is the most likely place for an eastward extension of Cederstrom's James River-Hampton Roads fault zone; such an extension would help explain the wide stratigraphic variations that exist between the Eastern Shore of Virginia and southeastern Virginia.

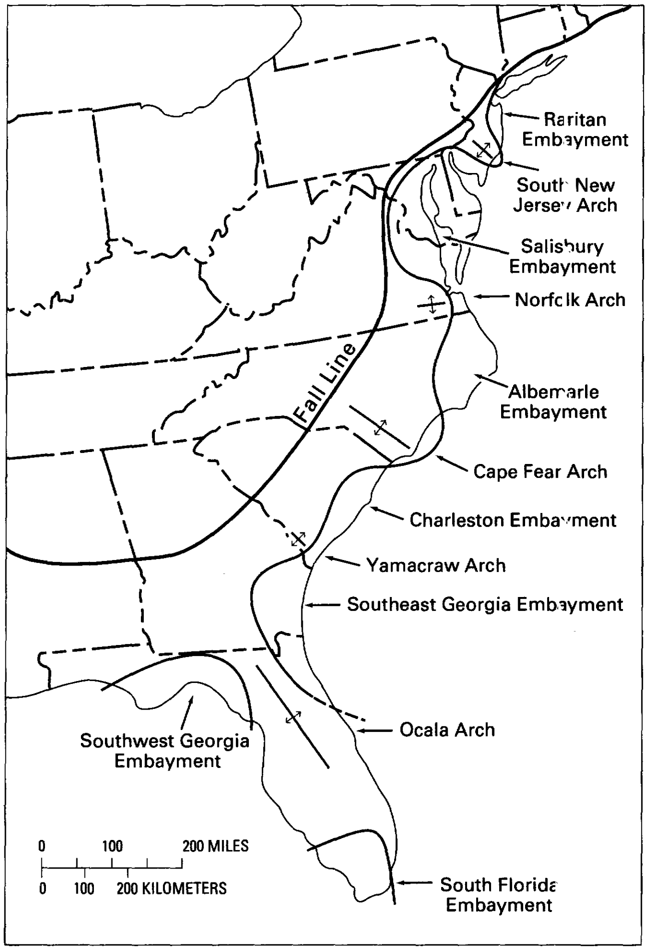


Figure 18.2. Principal basins and arches of the U.S. Atlantic Coastal Plain. Modified from Ward and Powars (1989, fig. 1).

BASEMENT ROCKS

The only test holes to reach basement along the Outer Coastal Plain of Virginia are the Dismal Swamp (−1,850 ft), the NOR–T–12 (−2,576 ft), and the Taylor #1 (−6,028 ft). Data from these holes and from seismic, gravity, and magnetic surveys indicate that the basement consists of Precambrian, Paleozoic, and lower Mesozoic (Triassic) rocks.

LOWER AND LOWER UPPER CRETACEOUS DEPOSITS

The Lower Cretaceous and lower Upper Cretaceous beds (Potomac Formation) consist of gray to greenish-gray, in part mottled red, brown, and yellow, poorly sorted, fine to coarse, quartzose and feldspathic sand, gravel, silt, and clay. Clays are primarily illite-smectite, illite, montmorillonite, and kaolinite and are of fluvial-deltaic origin; locally, they intertongue with glauconitic sands of marine origin. In the Taylor #1 test hole near the northern end of our cross section, the Potomac Formation is 4,690 ft thick and includes pollen zones K1, I, II, and III (Robbins and others, 1975). In contrast, the Lower Cretaceous and lower Upper Cretaceous deposits to the southeast in the Dismal Swamp corehole are only 1,290 ft thick. In the Dismal Swamp hole, only pollen zone IIB and a barren interval, which may represent zone III, were recognized in cored intervals (R.J. Litwin, written commun., 1988). The presence of some or all of the older pollen zones (K1, I, and II) in the uncored interval in the lowermost 767 ft is likely. As shown in cross section *A–A′* (fig. 18.3), these older pollen zones are nearly 3,180 ft of strata in the Taylor #1 test hole and, therefore, represent the largest lateral variation in stratal thickness among the studied lithic units.

UPPER CRETACEOUS DEPOSITS YOUNGER THAN POLLEN ZONE III

Upper Cretaceous deposits younger than pollen zone III were penetrated in the subsurface of Virginia's Eastern Shore and in southeastern Virginia, where they range in thickness from 121 ft in the Jenkins Bridge corehole (fig.

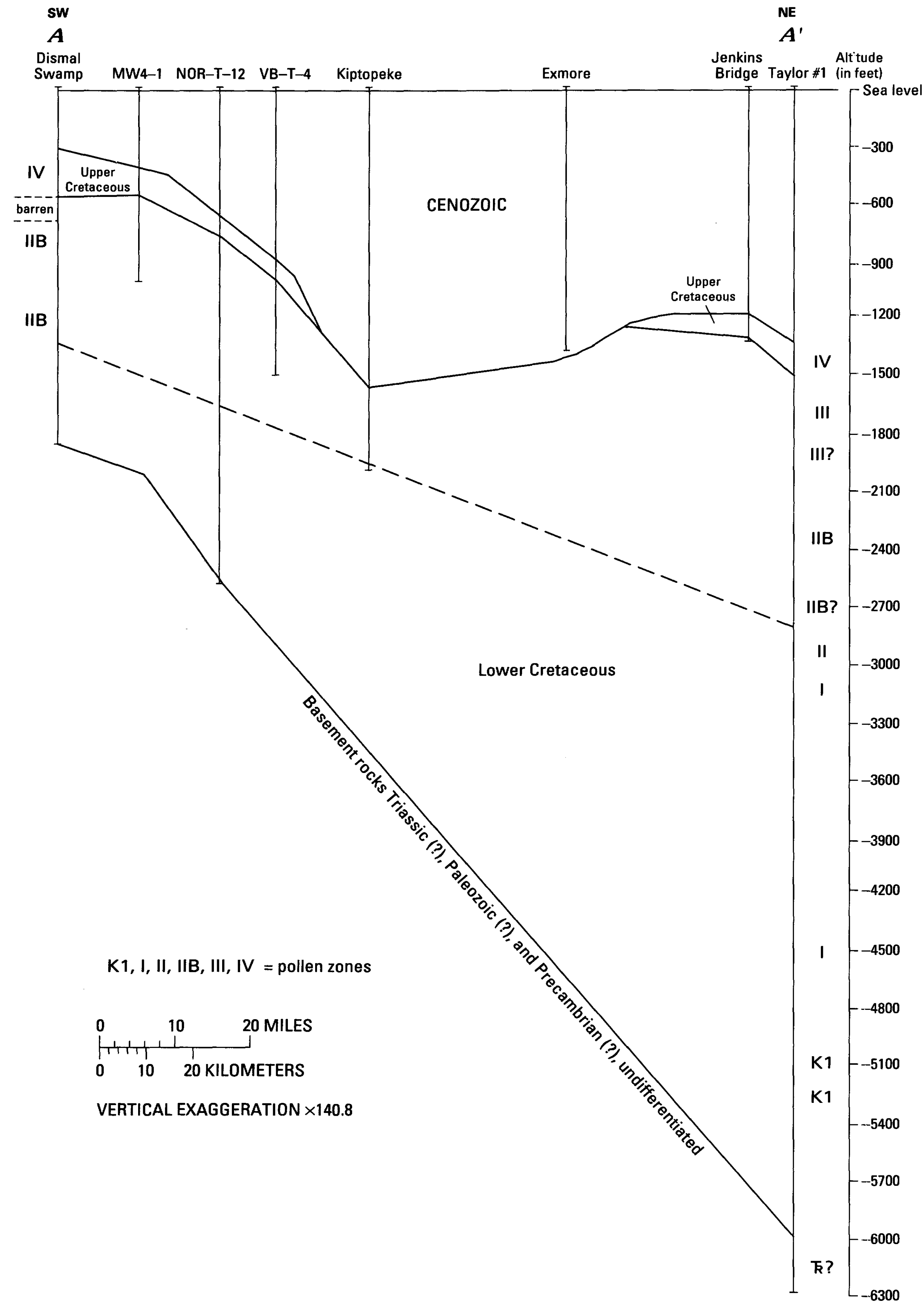


Figure 18.3. Diagrammatic geologic cross section to basement, Outer Coastal Plain of Virginia. Pollen data for Taylor #1 from Robbins and others (1975). Line of section is shown in figure 18.1.

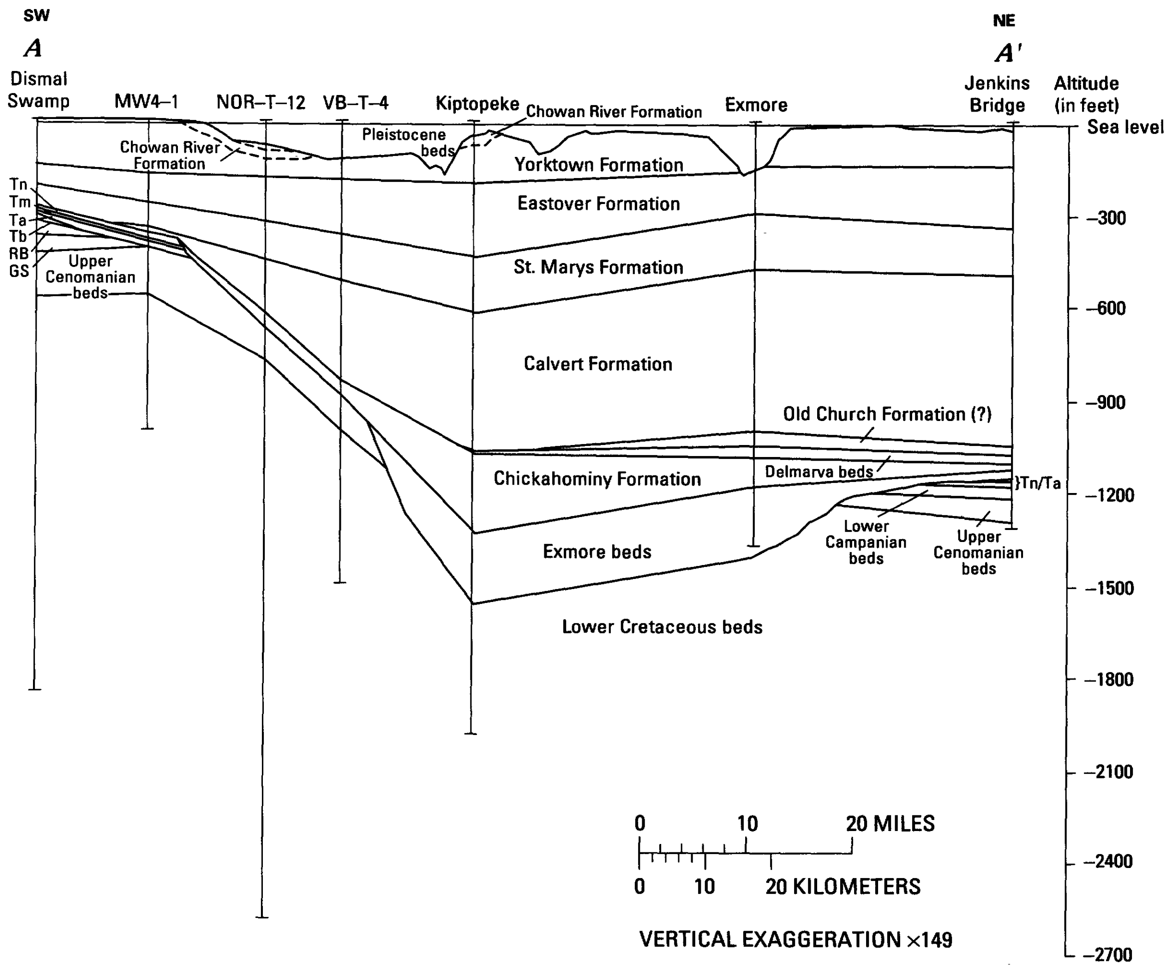


Figure 18.4. Uppermost Mesozoic and Cenozoic geologic cross section, Outer Coastal Plain of Virginia. GS, glauconitic sand unit; RB, red beds; Tb, Brightseat Formation; Ta, Aquia Formation; Tm, Marlboro Clay; Tn, Nanjemoy Formation. Line of section is shown in figure 18.1. This section shows more detail for the upper part of the section shown in figure 18.3.

18.3) to 348 ft in the Fentress corehole. As can be seen in cross section *A–A'* (fig. 18.4), the Upper Cretaceous deposits in southeastern Virginia (upper Cenomanian beds, glauconitic sand unit, and red beds) are much thicker than those in the Eastern Shore area (upper Cenomanian and lower Campanian beds).

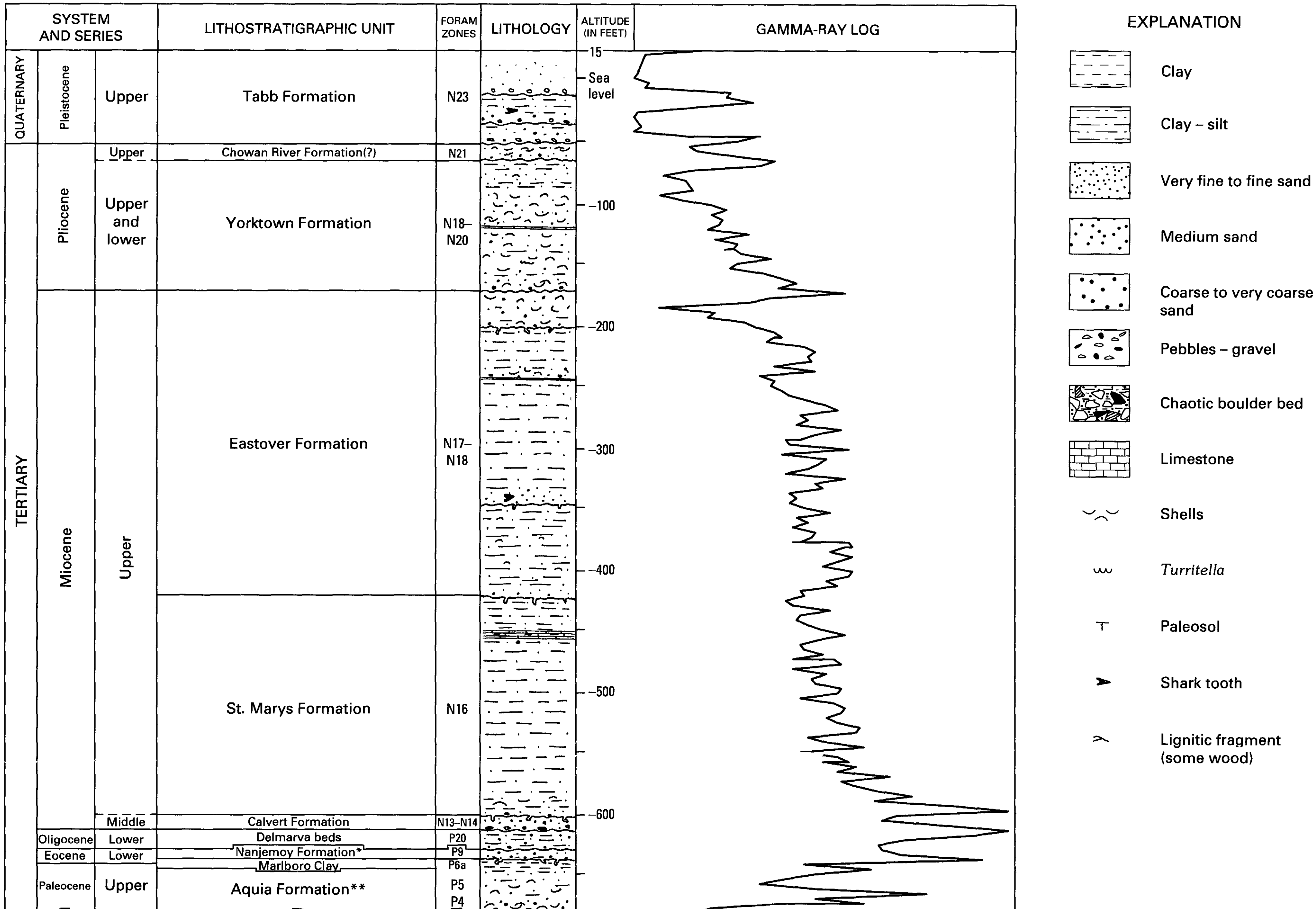
Southeastern Virginia

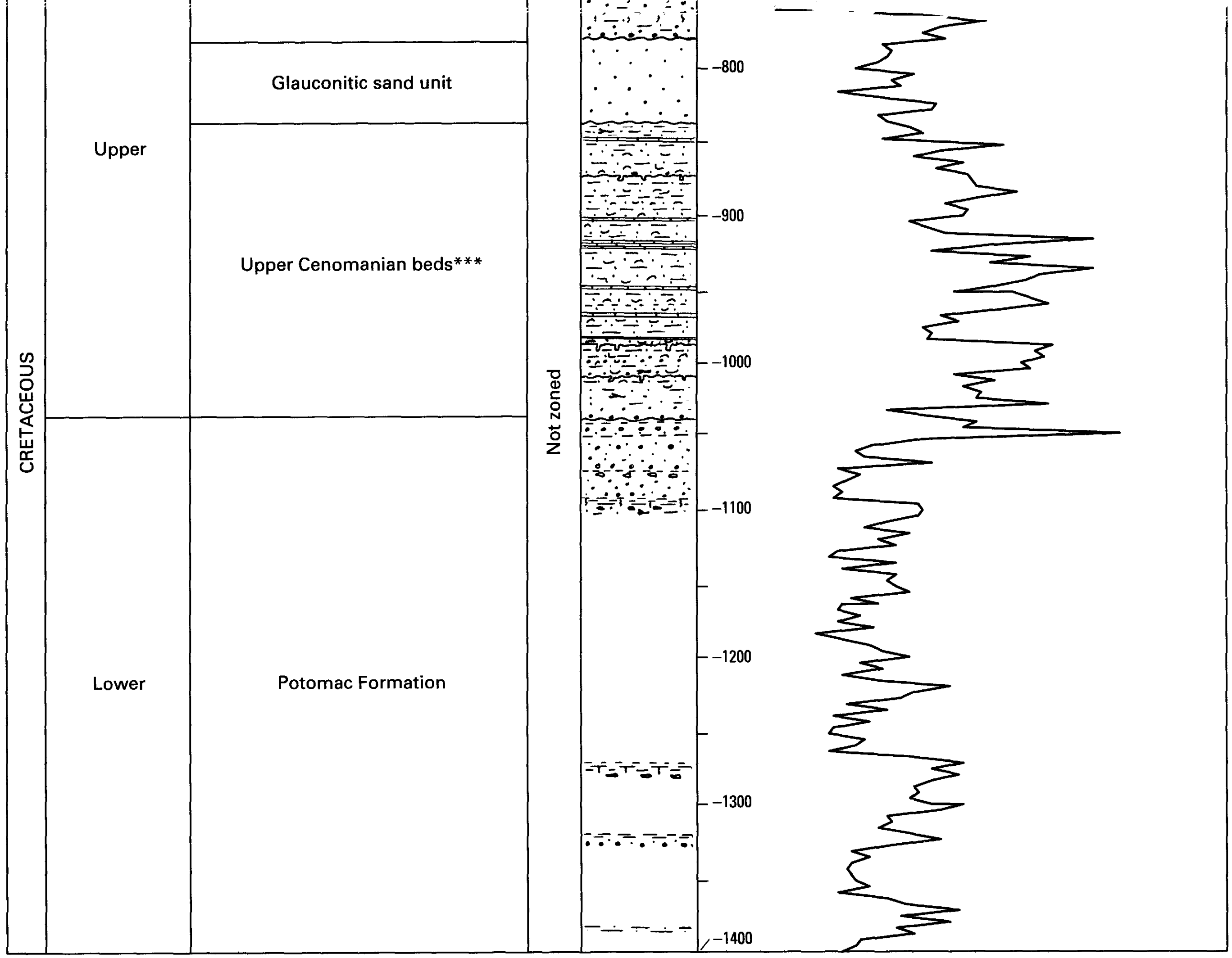
In southeastern Virginia, Upper Cretaceous deposits can be divided into three lithic units; in ascending order, they are upper Cenomanian beds consisting of marine and deltaic deposits, a glauconitic sand unit, and red beds. These units are described below.

Upper Cenomanian beds.—The upper Cenomanian beds (equivalent to the Raritan Formation of New Jersey) range from 150 to 200 ft in thickness and were found throughout the study area as the lowest lithic unit of the Upper Cretaceous deposits younger than pollen zone III. In the Fentress corehole (fig. 18.5), the bottom 50 ft of this lithic unit consists of two fining-upward sequences. Abundant wood fragments suggest a nearshore-shelf to deltaic depositional environment. At Fentress, the base of the upper Cenomanian deposits is marked by a sharp lithic contact at −1,035 ft elevation, where an olive-gray to light-gray, micaceous, lignitic, loose, medium to very coarse glauconitic quartz sand sharply overlies the Potomac Formation, which there consists of greenish-gray, silty and sandy clay, which grades downward into medium to coarse, crossbedded sand.

In the Dismal Swamp (fig. 18.6) corehole, the lowest upper Cenomanian deposits consist of glauconitic sands, which grade upward into shelly, glauconitic silt, clay, and sand of marine origin. The base of the upper Cenomanian is an erosional unconformity; dark-gray to black, clayey, glauconitic, phosphatic, pyritic, carbonaceous, pebbly quartz sand overlies and is burrowed down into the Potomac Formation, consisting of white to light-gray clay showing some yellow mottling. Within a couple of feet, these Lower Cretaceous beds grade downward into multicolored sands

FENTRESS COREHOLE

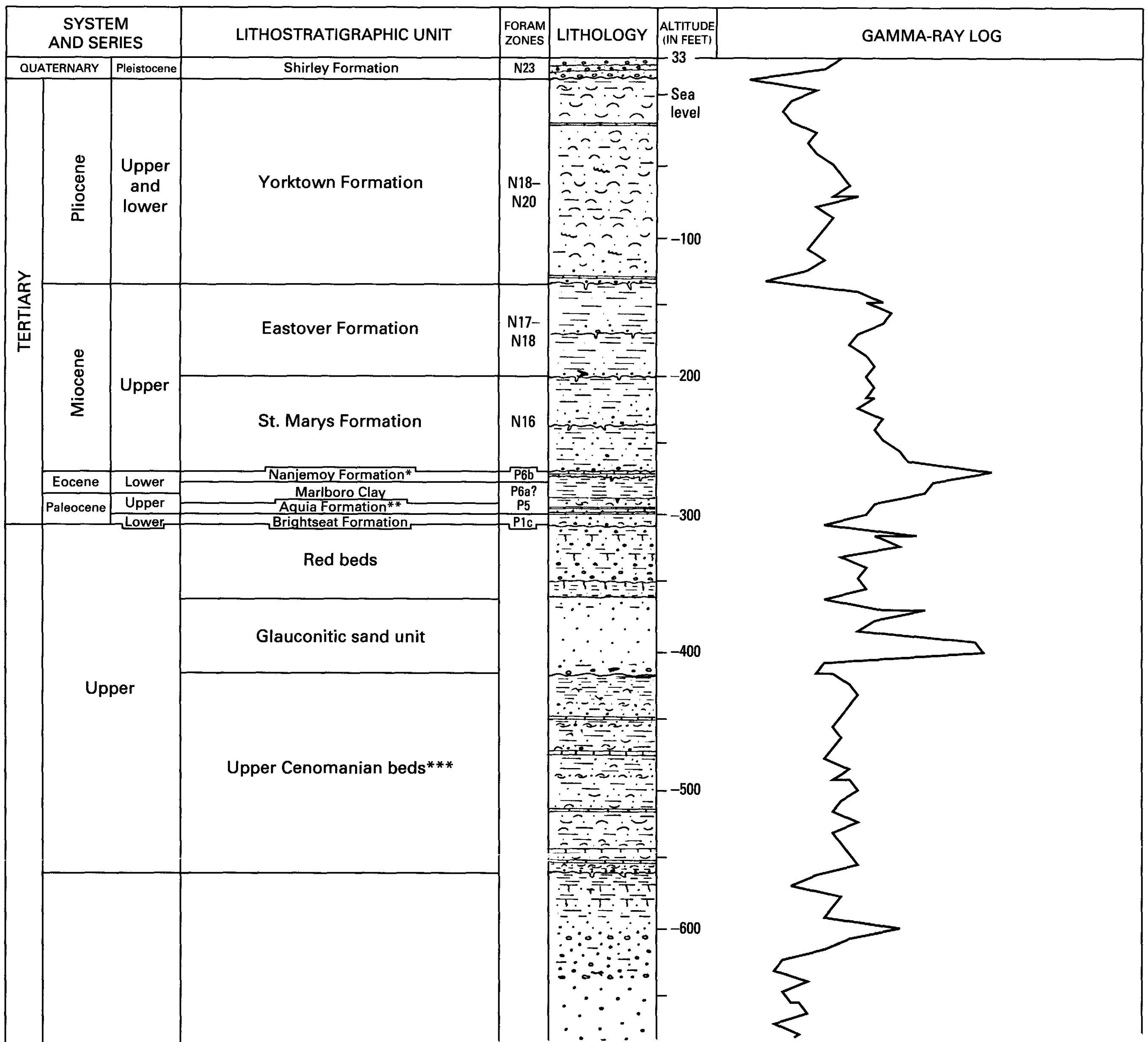


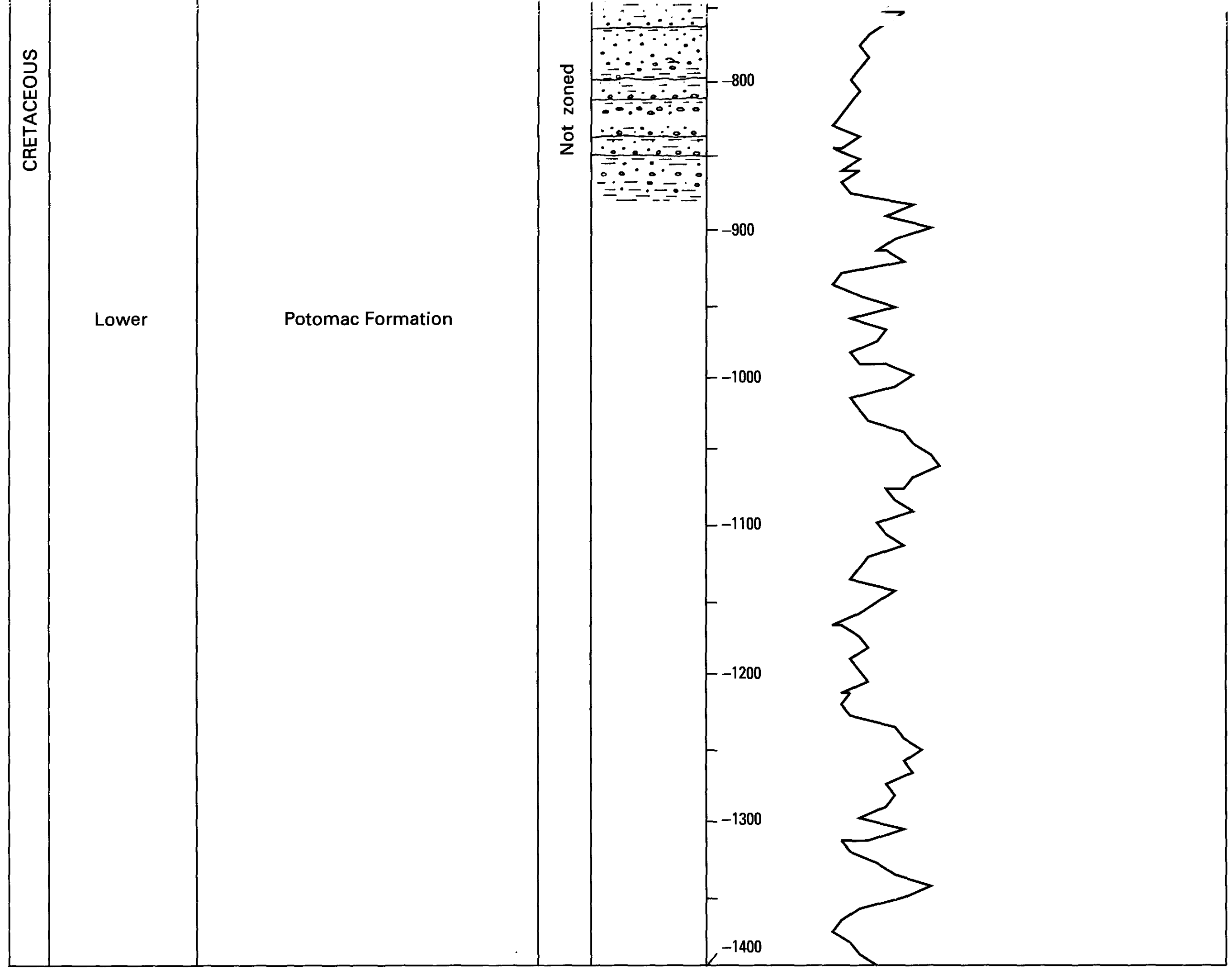


* Woodstock Member. ** Paspotansa and Piscataway Members. *** Pollen zone IV.

Figure 18.5. Lithic and gamma-ray logs of the Fentress corehole. The explanation shows patterns used in figures 18.5–18.7. Total depth of the Fentress borehole was 2,000 ft. Coring was continuous til 1,124 ft; the lowest cored interval is at 1,873 ft. Lithology is shown only for cored intervals. Foraminifera zones P4–P20 are extrapolated from dinocyst data from L.E. Edwards (written commun., 1988–89) and mollusk data from L.W. Ward (oral commun., 1988); zones N13–N14 are from benthic Foraminifera data provided by T.G. Gibson (oral commun., 1990); and zones N16–N23 are extrapolated from Gibson (1983).

DISMAL SWAMP COREHOLE





* Potapaco Member. ** Paspotansa Member. *** Pollen zone IV.

Figure 18.6. Lithic and gamma-ray logs of the Dismal Swamp corehole. See figure 18.5 for explanation. Total depth of the Dismal Swamp borehole was 2,000 ft. Coring was continuous til 1,113 ft. The lowest core was from 1,853 ft below sea level. Basement was hit at −1,850 ft. Foraminifera zones P1c–P5 and N16 are from T.G. Gibson (oral commun., 1987); zones P6a and P6b are extrapolated from dinocyst data from L.E. Edwards (written commun., 1987); and zones N17–N23 are extrapolated from Gibson (1983).

and clays, which are oxidized red, purple, brown, and yellow.

In the Fentress corehole, the lowermost 50 ft of nearshore and deltaic deposits is overlain by 115 ft of laminated to thick-bedded, olive-gray to dark-gray silt, clay, and fine to coarse sand containing variable amounts of glauconite, mica, shells, microfossils, wood, burrows, and pyrite that overall indicate a marine origin. This section contains numerous fining-upward sequences, which form a cyclic pattern. The sandier, coarser grained beds at the bases of sequences are generally more glauconitic and shelly than the overlying sediments. Many of the sandier shell beds and shell hashes (representing storm deposits) are cemented by calcium carbonate. In the Dismal Swamp corehole, 128 ft of these cyclic marine beds was penetrated, and their former proximity to a shoreline is indicated by the abundance of finely disseminated lignitic material scattered throughout the section. The presence of the mollusks *Inoceramus arvanus* and *Exogyra woolmani* in the marine section indicates a late Cenomanian age (N.F. Sohl, oral commun., 1988) and suggests correlation with unit E of Brown and others (1972), sequence 2 of Gohn (1988), and pollen zone IV (R.J. Litwin, written commun., 1988).

This marine unit is truncated and overlain by dark-gray, very fine grained, thinly laminated, very micaceous clayey silt to muddy sand containing abundant wood fragments that is 21.5 ft thick in the Dismal Swamp corehole and 33.9 ft thick in the Fentress corehole. The very abundant mica imparts a greasy feel to the core and cuttings and enhances recognition of the unit in well cuttings. These characteristics suggest a deltaic-lagoonal depositional environment.

Glauconitic sand unit.—The glauconitic sand unit is a loose sand that was found only in the Dismal Swamp and Fentress coreholes, where its thickness is 54.5 and 59 ft, respectively. The small amount of recovered core, the drill cuttings, and the geophysical logs indicate that most of the unit is fine to very coarse glauconitic quartz sand that was deposited in a marine environment. In the Dismal Swamp corehole, the basal 0.7 ft of this unit was recovered along with the contact with the underlying deltaic-lagoonal beds that make up the top of the upper Cenomanian unit. This contact is an erosional unconformity where light-green to olive-gray, poorly sorted, pebbly, muddy sand sharply overlies dark-gray, finely laminated, very micaceous (greasy feel), lignitic clayey silt. The basal bed of the glauconitic sand unit contains rip-up clasts from the underlying micaceous silt. The age of this unit has not been determined, but it overlies strata of late Cenomanian age and underlies strata of Danian age; therefore, it is latest Cenomanian to earliest Danian (probably pre-Danian) in age.

Red beds.—The red beds, like the glauconitic sand unit, appear to have a limited distribution near the Norfolk arch axis. The thickness ranges from 53.4 ft in the Dismal Swamp corehole to 91.3 ft in the Fentress corehole. The unit consists of a wide variety of lithologies, including homogeneous gray and green to mottled bright red, purple, yellow, orange, and brown sections of interbedded oxidized clay, silty clay, silty fine sand, and pebbly coarse sand. Some beds contain scattered mica, carbonaceous material, wood chunks, mud cracks, and rootlets. Several horizons are paleosols and are generally associated with the top of a fining-upward sequence. The bases of the fining-upward sequences are generally marked by a muddy pebbly sand that truncates the silty or sandy clay of an underlying paleosol. In the Dismal Swamp corehole, the base of the red beds is 363 ft below sea level, where a gray to multicolored, interbedded, micaceous clayey sand to sandy clay sharply overlies the glauconitic sand unit, which consists of dark-green to gray-green, fairly well sorted, fine to coarse quartz sand containing small amounts of glauconite and phosphate. The age of the red beds has not been determined, but it too must be latest Cenomanian to earliest Danian. In this report, the glauconitic sand unit and red beds together are tentatively correlated with unit D of Brown and others (1972) and sequence 3 of Gohn (1988).

Eastern Shore of Virginia

The Upper Cretaceous section younger than pollen zone III is thinner in the Eastern Shore area than it is in the southeastern Virginia area. In the Eastern Shore area, it consists of marine to deltaic deposits. In the Jenkins Bridge corehole, upper Cenomanian beds 73 ft thick are overlain by lower Campanian beds 48 ft thick (G.S. Gohn, written commun., 1988). Reworked marine microfauna and clasts of Maastrichtian age in the lower upper Eocene channel fill or subaqueous debris flows of the Exmore and Kiptopeke coreholes indicate that thin Maastrichtian shelf deposits are locally present in the Eastern Shore area.

LOWER TERTIARY DEPOSITS

Lower Tertiary deposits consist primarily of glauconitic, clayey sands and silts. Microfauna indicate Paleocene, Eocene, and Oligocene units of marine origin. In southeastern Virginia, previously unrecognized equivalents of the lower Paleocene Brightseat Formation, upper Paleocene Aquia Formation, upper Paleocene and lower Eocene Marlboro Clay, and lower Eocene Nanjemoy Formation were found (L.E. Edwards, T.G. Gibson, and L.W. Ward, written and oral commun., 1988). In this area (see cross section *A–A'*, fig. 18.4), the formations are relatively thin and are bounded by unconformities.

The Brightseat Formation is a muddy glauconitic sand only a few feet thick. It is identified only in the Dismal Swamp corehole on the basis of Foraminifera (T.G. Gibson), calcareous nannofossils (L.M. Bybell), and dinoflagellates (L.E. Edwards) (figs. 18.4 and 18.6).

The Aquia Formation ranges in thickness from 15 to 40 ft. It consists of shelly, glauconitic sand; phosphatic pebbles or very coarse glauconitic quartz sand marks the base of the unit. Dinocysts indicate that the entire unit in the Dismal Swamp corehole and the upper part of the Aquia in the Fentress corehole are equivalent to the outcropping Paspotansa Member of the Aquia (L.E. Edwards, written commun., 1987). The brachiopod *Oleneothyris harlani* indicates that the lower part of the Aquia in the Fentress corehole is equivalent to the outcropping Piscataway Member (L.W. Ward, oral commun., 1988).

The Marlboro Clay consists of 10 to 16 ft of light-gray kaolinitic clay and is gradational with the underlying Aquia beds.

The Nanjemoy Formation ranges from 3.5 to 23 ft in thickness and consists of dark-olive-gray to olive-black, clayey, silty, fine to coarse, glauconitic quartz sand. It has a sharp, burrowed contact with the underlying Marlboro Clay. In the Dismal Swamp corehole, the Nanjemoy Formation includes Beds A and B (defined by Ward, 1984, 1985, and Powars, 1987) of the Potapaco Member (L.E. Edwards, written commun., 1988). In the Fentress corehole, the Nanjemoy Formation consists of 8 ft of clayey silty very fine sand of the Woodstock Member, which is younger than the Potapaco Member. This unit is overlain by 15 ft of coarser glauconitic sand containing dinocysts that suggest a middle or late Eocene or early Oligocene age (L.E. Edwards, written commun., 1989). The middle and late Eocene dinocysts are probably reworked into lower Oligocene deposits.

The lower Tertiary deposits in the Eastern Shore area appear to consist mainly of strata younger than those in the southeastern Virginia and Inner Coastal Plain areas. The Eastern Shore section includes lower upper Eocene, upper Eocene, and lower and upper Oligocene glauconitic sands and silts. The lower Tertiary deposits were not cored in the Jenkins Bridge corehole; however, well cuttings and geophysical logs suggest that they consist of glauconitic, clayey and silty sands and probably include upper Eocene as well as lower and upper Oligocene and lower Miocene(?) deposits similar to those in the Exmore corehole (figs. 18.4 and 18.7). Further evidence for the presence of these units in the Jenkins Bridge corehole is the presence of these units to the north in the Hammond oil test hole (see fig. 18.1, C.W. Poag, written commun., 1988).

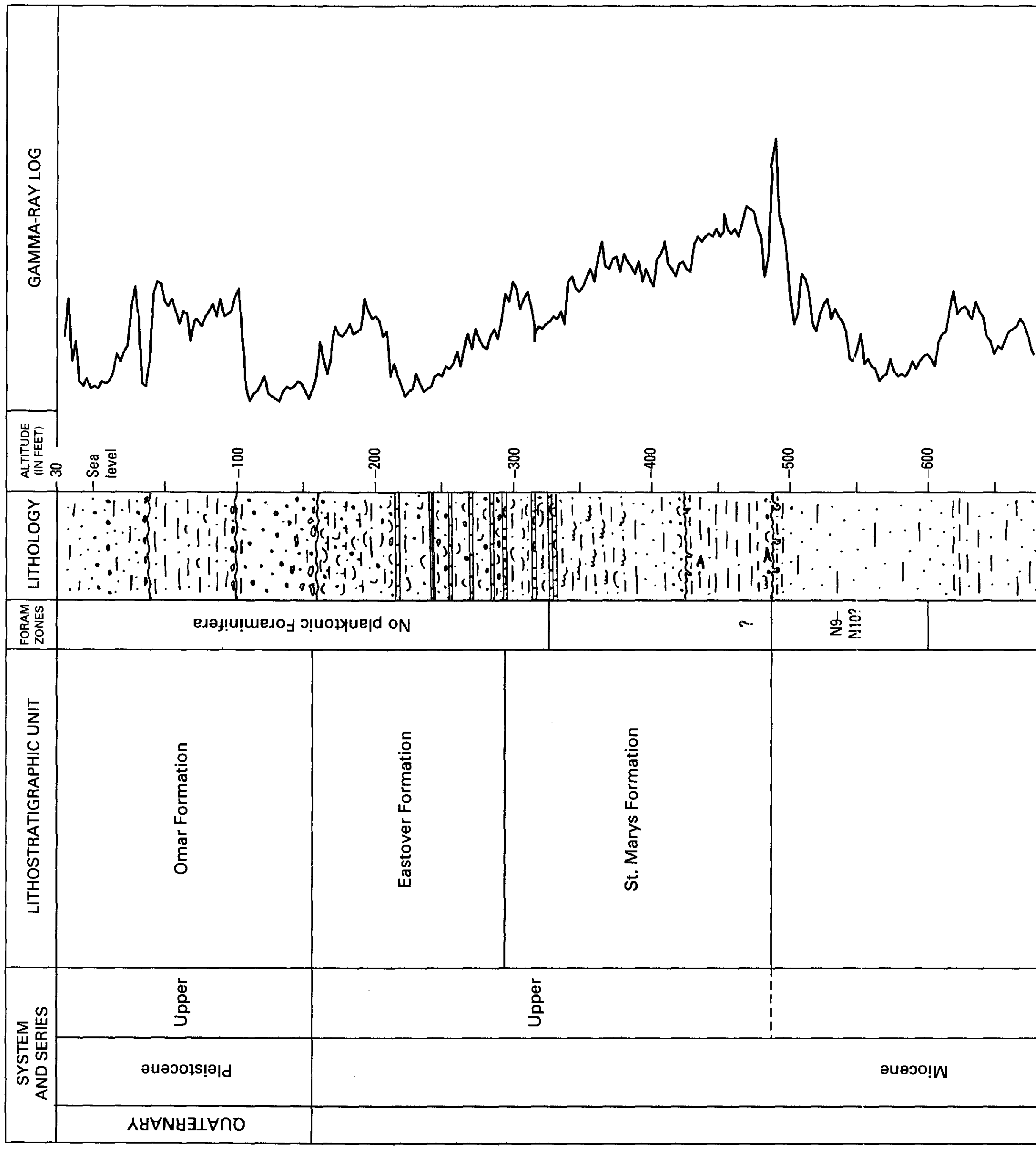
The lower upper Eocene beds (dated on the basis of foraminiferal assemblages studied by C.W. Poag, oral commun., 1991) are herein informally named the Exmore beds because they were discovered in the Exmore corehole; they were also found in the Kiptopeke corehole. They consist of as much as 221 ft of gray, clayey and silty, glauconitic sand containing abundant clasts and reworked fauna and flora from Lower Cretaceous (Albian), Upper Cretaceous (Cenomanian, Santonian, Campanian, and Maastrichtian), Paleocene, and lower and lower middle Eocene deposits (L.E. Edwards, C.W. Poag, and R.J. Litwin, written commun., 1988). The clasts become larger and more abundant downward in the section, where boulder-sized material is common. Many clasts consist of friable, unconsolidated material that could not have traveled far. These observations support an inferred origin for these unusual beds as either a subaqueous channel fill or a debris flow at the base of a paleoshelf or fault scarp. A channel-fill origin is harder to explain because the coreholes are more than 30 mi apart; perhaps they hit the same channel (which may trend north), or perhaps they penetrated two separate channels. A multiple-channel origin is supported by data from north and south of the Virginia study area. Poag's (1985) seismic-reflection data for the New Jersey continental shelf, slope, and rise reveal the presence of numerous early, middle, and late Eocene channels along the upper rise and slope. These channels are within a unit that has been traced to the shoreline. Popenoe's (1985) seismic-reflection data for the North Carolina continental shelf indicate the presence of a large Eocene channel.

The lower Tertiary deposits thicken southward from 129 ft in the Jenkins Bridge corehole to 371 ft in the Exmore corehole to 480 ft in the Kiptopeke corehole. This thickening to the south is due to the presence of thicker deposits of upper Eocene deposits. This thickening and the abrupt thinning and truncation of these deposits at and just south of the mouth of Chesapeake Bay indicate deposition in a subbasin bounded on the south by a paleoshelf or fault scarp.

The upper Eocene deposits (Chickahominy Formation) are 99 ft thick in the Exmore corehole and 259 ft thick in the Kiptopeke corehole and consist primarily of olive-gray, compact, glauconitic, micaceous, clayey silt and silty clay, which contain abundant microfauna (enough to give a white-speckled appearance), abundant finely crystalline iron sulfide, and scattered shell fragments. The lower part of the section coarsens downward to a very fine to fine sand and contains pebbles and reworked microfauna from Upper Cretaceous through middle Eocene deposits (L.E. Edwards and C.W. Poag, written commun., 1988). The contact between the Exmore beds and the overlying Chickahominy Formation appears to be gradational.

Lower Oligocene deposits were found in the Virginia Coastal Plain for the first time and are, herein, informally referred to as the Delmarva beds. These Delmarva beds consist of as much as 41 ft of olive-gray to grayish-olive, micaceous, clayey, silty, very fine, glauconitic quartz sand containing scattered patches of pyrite and marcasite. This sand contains a fairly abundant microfauna, and many reworked specimens from Upper Cretaceous through middle Eocene deposits are concentrated in the basal part of the section (L.E. Edwards and C.W. Poag, written commun., 1988). These beds coarsen upward to a very fine to fine sand. The base of the Delmarva beds is marked by an erosional unconformity representing a hiatus estimated to

EXMORE COREHOLE



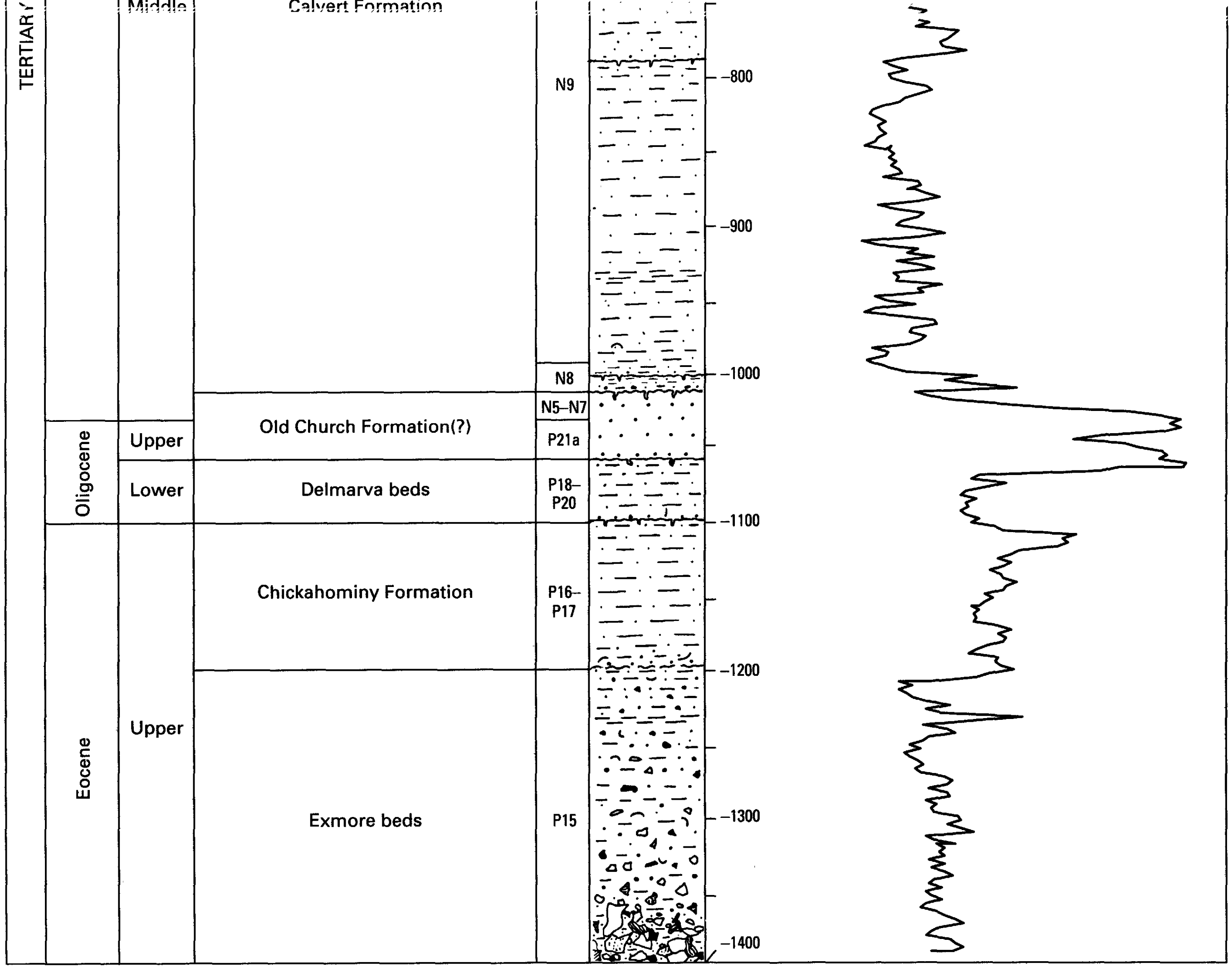


Figure 18.7. Lithic and gamma-ray logs of the Exmore corehole. See figure 18.5 for explanation. Total depth 1,396 ft. Foraminifera zones from planktonic Foraminifera data from C.W. Poag (written commun., 1987–91).

have lasted 0.5–8 m.y. (C.W. Poag, written commun., 1988).

The upper Oligocene and lower Miocene(?) deposits found exclusively in the Exmore corehole appear to include strata that are equivalent to and younger than the Oligocene(?) deposits west of Chesapeake Bay (Old Church Formation of Ward, 1984, 1985, and Mixon and others, 1989). These corehole deposits consist of about 45 ft of dark-olive-gray to greenish-black, glauconitic and phosphatic, fine to very coarse sand containing lesser amounts of quartz. The sand has an olive-brown clay-silt matrix. The section is noncalcareous except for about the top 10 ft, which contains a fairly abundant microfauna. In the core, the sharp burrowed contact between the upper Oligocene and lower Miocene(?) beds and lower Oligocene beds is probably an erosional unconformity. Foraminifera data (C.W. Poag, written and oral commun, 1990) suggest that the lower 17 ft of this glauconitic and phosphatic sand is late early Oligocene in age and that the upper 23 ft is early Miocene in age. In contrast, dinocyst data (L.E. Edwards, written commun., 1988) indicate that the same section is late Oligocene in age. Except for the increase in calcareous material in the top 10 ft, lithically this section appears very homogeneous and is, therefore, defined as one unit and is tentatively assigned to the Old Church Formation.

UPPER TERTIARY DEPOSITS

The upper Tertiary deposits consist of Miocene and Pliocene clays, silts, and very fine to very coarse sands, are variably diatomaceous, glauconitic, and shelly, and contain some indurated beds (cemented by calcium carbonate). Microfauna and macrofauna indicate a marine to restricted-marine origin. The upper Tertiary deposits (fig. 18.4) generally form a northeastward-thickening wedge of sediments that range in thickness from 285 ft in the Dismal Swamp corehole to 1,022 ft in the Jenkins Bridge corehole. However, in the Kiptopeke corehole, all the stratigraphic formations are found at their lowest elevations; this low area suggests a correlation (possibly structural) with the late early Tertiary subbasin and the eastward extension of Cederstrom's (1945b) James River-Hampton Roads fault zone. The section includes the Calvert Formation (lower and lower middle Miocene), the St. Marys Formation (upper middle and upper Miocene), the Eastover Formation (upper Miocene), the Yorktown Formation (lower and upper Pliocene), and the Chowan River Formation (upper Pliocene); the Calvert Formation accounts for most of the thickening. The Choptank Formation (middle middle Miocene), which is found west of Chesapeake Bay, appears to be absent from Virginia's Eastern Shore and southeastern Virginia. Foraminifera data (C.W. Poag, written commun., 1990) indicate that the only lower Miocene beds are 23 ft of glauconitic, phosphatic sand in the Exmore corehole, which are interpreted as the upper part of the Old Church Formation. Dinocyst data for the same beds suggest a late Oligocene age.

Diatom data (G.W. Andrews, written commun., 1987–88) and Foraminifera data (C.W. Poag and T.G. Gibson, written and oral commun., 1987–89) indicate that the Calvert Formation throughout the section is middle Miocene in age. The Calvert Formation is absent from the Dismal Swamp corehole, is 12 ft thick in the Fentress corehole, thickens to 458 ft in the Kiptopeke corehole, and gradually thickens to 550 ft in the Jenkins Bridge corehole. The Calvert thins abruptly at and just south of the mouth of Chesapeake Bay. The thicker Calvert sections contain as many as 10 fining-upward sequences. At the base of each sequence is a light- to dark-olive-gray, clayey and silty, very fine to fine sand that commonly contains medium to coarse sand with scattered quartz and phosphate pebbles, phosphate and (or) glauconitic chips and grains, and very sparse to abundant shells. These basal sands grade upward into very fine sandy clay silt and diatomite. The basal contacts of the sequences are generally sharp and burrowed. Diatoms (G.W. Andrews, written commun., 1988) indicate that the Calvert Formation in the Exmore corehole includes 9 ft of the Fairhaven Member, 213 ft of the Plum Point Marl Member, and 297 ft of the Calvert Beach Member. There the Calvert Beach Member has more sand than the rest of the Calvert Formation.

Along the line of section, the St. Marys Formation is the oldest Tertiary unit that extends across the entire study area and overlies older units over the Norfolk arch. In the Dismal Swamp corehole, Foraminifera-rich sandy clays of the St. Marys Formation (upper Miocene part) unconformably overlie glauconitic sandy limestone of the Nanjemoy Formation (lower Eocene) (T.G. Gibson, oral commun., 1990). The St. Marys consists of up to 189 ft of muddy, very fine sand and sandy clay and silt containing scattered shells, fairly abundant iron sulfide, and finely disseminated organic material. At the northern end of the study area, the St. Marys is found at the same elevation in both the Exmore and Jenkins Bridge coreholes and can be divided into a lower clayey silty section about 60 ft thick and an upper sandy shelly section 100 to 129 ft thick. To the south, the entire section is like the lower finer grained section. An erosional unconformity at the base of the St. Marys Formation is indicated by (except in the Dismal Swamp corehole) shelly, poorly sorted, woody, pyritic, very fine to coarse sand of the St. Marys abruptly overlying olive-gray clayey silt of the Calvert Formation (Calvert Beach Member).

The Eastover Formation (upper Miocene) extends across the entire study area and ranges in thickness from 69 ft in the Dismal Swamp corehole to 256 ft in the Kiptopeke corehole. The Eastover consists of dark-gray to bluish-gray to greenish-gray, muddy fine sand interbedded with finer and coarser grained beds. The Eastover is sparsely to abundantly shelly, contains some shell hashes and indurated beds, and is, in part, glauconitic and micaceous. Macrofauna and microfauna indicate a shallow-water, marine to restricted-marine depositional environment. The mollusk *Isognomon maxillata* is a common species and is abundant in the upper half of the Eastover section in the Exmore corehole. In the Eastern Shore area, the upper half of the Eastover appears to be coarser grained than the lower half and may represent a regressive phase.

The Yorktown Formation (lower and lower upper Pliocene) extends across the entire study area except where it has been cut out by Pleistocene channeling, as in the Eastville and Exmore paleochannels (Mixon, 1985; Colman and Mixon, 1988; D.S. Powars and R.B. Mixon, unpub. data). As seen in cross section *A–A′* (fig. 18.4), the Yorktown deposits differ from all other units by having a nearly horizontal base along the line of section. The Yorktown includes bluish-gray, greenish-gray, and dark-greenish-gray, very fine to coarse sand, in part glauconitic and phosphatic. The formation is commonly very shelly (locally becoming a bioclastic sand) and contains an abundant microfauna. The sands are commonly interbedded with gray and blue-gray sandy and silty clay. The basal part of the Yorktown coarsens downward into a pebbly, glauconitic, and phosphatic quartz sand sharply overlying and burrowed into the underlying Eastover Formation. The contact is an erosional unconformity.

The Chowan River Formation (upper Pliocene) was reported (Gibson, 1983) in the Moores Bridge well (labeled NOR–T–12 in figs. 18.1 and 18.4) at depths from about 90 to 115 ft and consists of blue-green fossiliferous clayey fine sand. It is also reported (Johnson and others, 1987) locally only in a few borrow pits (Gomez and Yadkin pits) and in the subsurface of southeastern Virginia, where it is as much as 51 ft thick in eastern Virginia Beach. It consists of interbedded silty fine sand, clayey silt, and bioclastic sand. Exposure of the Chowan River Formation in relatively shallow borrow pits less than 5 mi south of the NOR–T–12 well suggests either a steep gradient between them (possibly caused by structural effects) or incorrect correlation. Several whole mollusks that washed out of the Kiptopeke pilot hole above 200 ft have been identified by L.W. Ward (written commun., 1990) as restricted to the Chowan River Formation. They probably came from a dark-greenish-gray, interbedded shelly to very shelly, silty fine sand and muddy fine sand containing a few scattered medium to coarse grains; the sands are present from −48 to −72 ft elevation. A bioclastic silty fine to medium sand found in the Fentress corehole from −51 to −65 ft elevation is correlated with the Chowan River Formation.

QUATERNARY DEPOSITS

The Quaternary beds include fluvial, estuarine, marginal-marine, and nearshore shelf sediments deposited during the Pleistocene glacial-interglacial regime. The Eastville and Exmore paleochannels are Pleistocene channels, representing ancient Susquehanna River courses, cut during sea-level lowstands (glacial intervals). Erosion of the Eastville paleochannel is thought by Colman and Mixon (1988) to have taken place about 150,000 yr ago. These workers report that the Exmore paleochannel was incised around 250 Ma or, possibly, 450 Ma; these dates are based, respectively, on uranium-series and amino-acid-racemization data of Mixon and others (1982), Szabo (1985), and Wehmiller and others (1988). The Eastville and Exmore paleochannel fills are transgressive sequences consisting, from bottom to top, of coarse fluvial gravel and sand, estuarine muds, and lower estuarine to open-bay sand and sandy mud that are shelly in part. The channel fills are overlain by nearshore-shelf and barrier-spit deposits representing deposition during interglacial high stands. See Mixon (1985) and Mixon and Powars (1985) for a more comprehensive study of the uppermost Cenozoic deposits in the Eastern Shore area.

CONCLUSIONS

1. Our new data indicate that both Virginia's Eastern Shore and the southeastern Virginia area appear to have undergone subsidence during the Early Cretaceous (Albian) and early Late Cretaceous (late Cenomanian) and relative subsidence during at least part of Paleocene, Eocene, Oligocene, Miocene, and Pliocene time.
2. The Upper Cretaceous and Tertiary strata in the Outer Coastal Plain of Virginia record a complex depositional, erosional, and tectonic history for the northward transition from the Norfolk arch to the Salisbury embayment. A zone of abrupt thinning and truncation of Upper Cretaceous and Tertiary formations at and just south of the mouth of Chesapeake Bay may be fault controlled.
3. Thick Upper Cretaceous strata of marine and fluvial-deltaic origin are truncated northward across the mouth of Chesapeake Bay. The Upper Cretaceous marine and deltaic beds younger than pollen zone III (upper Cenomanian beds, equivalent to Raritan Formation of New Jersey) are widespread and are as much as 200 ft thick in southeastern Virginia. They overlie the Potomac Formation, which contains a barren pollen interval (possibly equivalent to pollen zone III) that overlies several hundred feet of strata that contain zone II pollen. In contrast, at the northern end of the Eastern Shore of Virginia in the Jenkins Bridge corehole, the upper Cenomanian beds are only 73 ft thick and are overlain by 48 ft of lower Campanian marine deposits.
4. In southeastern Virginia, red beds (paleosols) as much as 100 ft thick and glauconitic sands as much as 60 ft thick are found between Upper Cretaceous beds of marine to deltaic origin and Danian beds (Brightseat Formation). Extensive development of these red paleosols on Cretaceous units indicates that the Norfolk arch area in

southeastern Virginia underwent relative uplift and prolonged subaerial exposure during some of the time after the late Cenomanian and before the Danian.

5. Thin remnants of Paleocene beds (the Brightseat and Aquia Formations and the lower part of the Marlboro Clay) and lower Eocene beds (the upper part of the Marlboro Clay and the Nanjemoy Formation) are present farther southward (as far south as the Dismal Swamp corehole) than was previously recognized.
6. In the southern and central part of Virginia's Eastern Shore, the Tertiary section includes previously undescribed lower upper Eocene beds that contain abundant cobble- to boulder-sized clasts and reworked Foraminifera and dinocysts from older Paleogene and Cretaceous strata. These beds appear to represent multiple paleochannel fills or an extensive subaqueous debris flow as much as 221 ft thick. These strata truncate older Paleogene and Cretaceous units. The lower upper Eocene deposits are conformably overlain by very fossiliferous upper Eocene shelf deposits as much as 257 ft thick. Both units thicken toward the southern tip of the Eastern Shore and are truncated across the mouth of Chesapeake Bay; the truncation suggests a subbasin bounded on the south by a shallow paleoshelf and (or) a fault scarp.
7. The upper Eocene deposits are overlain by previously undescribed lower and upper Oligocene and lower Miocene(?) glauconitic sands as much as 86 ft thick. The lower and upper Oligocene and lower Miocene(?) deposits are truncated southward near the southern end of the Eastern Shore. However, a couple of feet of the lower Oligocene unit is present in the Kiptopeke corehole, and 15 ft is present in the Fentress corehole; thus, the unit is present only along the coast.
8. The middle Miocene part of the Calvert Formation thickens abruptly northward across the mouth of Chesapeake Bay and maintains a thickness of about 500 ft across the northern part of the study area; this thickness suggests substantial subsidence of the Eastern Shore area during this time.
9. The middle and upper Miocene St. Marys Formation is the oldest unit to extend across the entire study area and is overlain by the upper Miocene Eastover Formation. Lower and lower upper Pliocene marine deposits extend uniformly across the entire area, exhibiting little or no tilting between arch and basin areas, and are overlain, and locally deeply incised, by Quaternary deposits of fluvial to marine origin.

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19. Stratigraphic and Structural Controls on Ground-Water Flow in the Coastal Plain Aquifers, North-Central Charles County, Maryland[1]

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INTRODUCTION

Erosional truncation, facies changes, and sediment overlap control the hydrostratigraphic relations among the aquifers that supply ground water to the Waldorf area of Charles County, Md. Deformation of the Coastal Plain aquifers by faulting associated with the Brandywine fault system may also, in part, control the availability and flow of ground water in north-central Charles County.

Waldorf lies in the north-central part of Charles County, Md., about 15 mi southeast of Washington, D.C. (fig. 19.1). The term "Waldorf area" as used here defines an area within about a 5-mi radius of the center of Waldorf.

The Maryland Geological Survey and the Water Resources and Geologic Divisions of the U.S. Geological Survey (USGS) have recently engaged in several cooperative research efforts in southern Maryland. This paper draws upon data gathered and results obtained from the following cooperative projects: a study of the hydrogeology and water-supply potential of the aquifers in the Waldorf area of Charles County, the drilling and analysis of a stratigraphic corehole in Waldorf, and the mapping of the surficial geology of Charles County.

The hydrogeologic framework and a conceptual model of the aquifers at Waldorf were developed on the basis of local and regional stratigraphic and structural relations among the formations and aquifers. Lithologic data, geophysical logs, and micropaleontological analyses of cores were used to correlate the aquifer and formational boundaries in the subsurface. The hydrogeologic framework was the "architectural blueprint" used to construct the digital ground-water-flow model of the aquifers in the Waldorf area. The physical relations and hydraulic properties of the aquifers were represented numerically. These numerical values were input to the ground-water-flow-model computer code written by McDonald and Harbaugh (1984). The flow model incorporates seven layers to simulate the seven water-bearing units that lie above the Lower Cretaceous Arundel Formation at Waldorf. The model was calibrated for transient conditions from 1900 to 1985 by

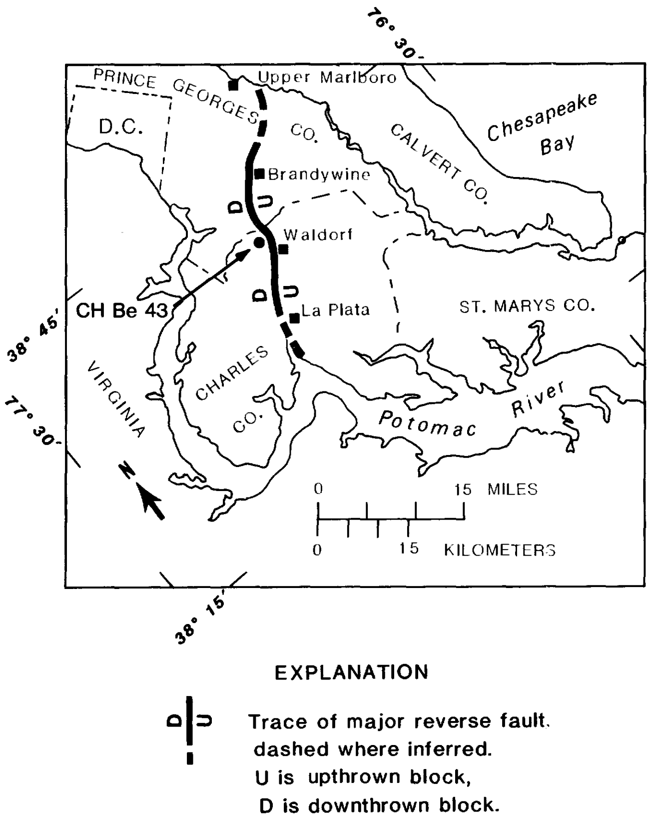

Figure 19.1. Trace of major reverse fault within the Brandywine fault system in Prince Georges and Charles Counties, Md.

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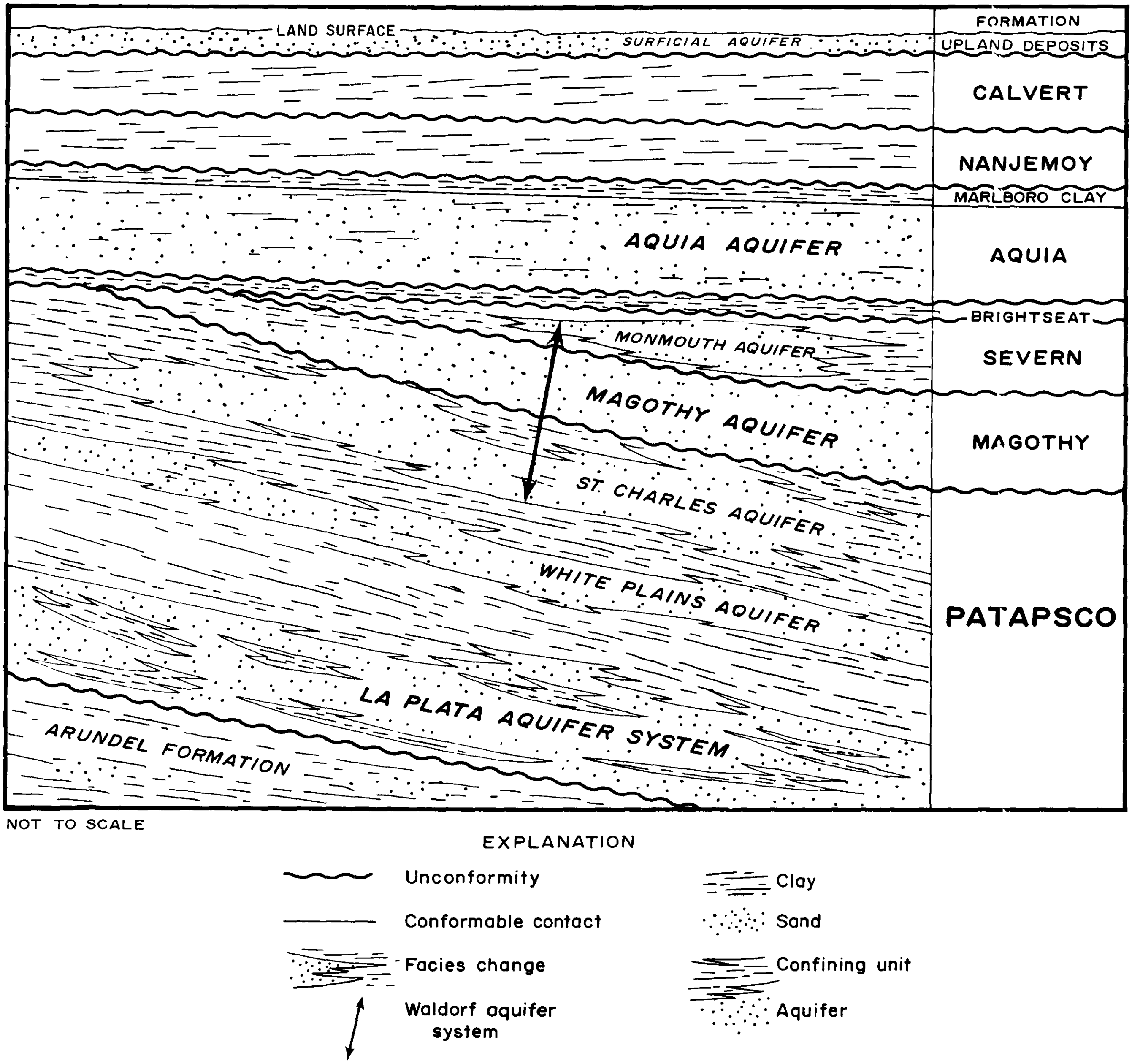

Figure 19.2. Stratigraphic relations among the formations, aquifers, and confining units in the Waldorf area of Charles County, Md. Effects of faulting are not shown.

comparison of simulated water levels with about 130 hydrographs for observation wells throughout southern Maryland.

STRATIGRAPHIC RELATIONS AND THEIR SIMULATION

The two major hydrologic units that supply the Waldorf area's water needs are the Waldorf and the La Plata aquifer systems (fig. 19.2). The stratigraphically higher of the two aquifer systems is the Waldorf, which consists of three hydraulically connected aquifers. The uppermost aquifer of this system—the Monmouth—consists of a coarse quartz sand that probably is a nearshore facies of the marine, Upper Cretaceous Severn Formation. The middle aquifer—the Magothy—is a sand facies of the eastward-thickening, fluviomarine, Upper Cretaceous Magothy Formation. The basal aquifer of the system—the St. Charles—consists of sands within the fluviodeltaic, Lower to Upper Cretaceous Patapsco Formation. The name "Waldorf aquifer system" (Wilson, 1986), as used in this paper, refers to the Monmouth, Magothy, and St. Charles aquifers only within about a 5-mi radius of Waldorf. A relatively minor aquifer—the White Plains—underlies the St. Charles and overlies the La Plata aquifer system in the Waldorf area (fig. 19.2).

Marine transgressions during the Late Cretaceous resulted in (1) erosional truncation of the Patapsco Formation and deposition of the overlying Magothy Formation and (2) erosional truncation of the Magothy Formation and deposition of the overlying Severn Formation. This beveling of underlying formations formed thin and leaky confining units and sand-on-sand contacts among the aquifers that compose the Waldorf aquifer system. Within 5 mi to the west and south of Waldorf, the Magothy and Severn Formations are completely eroded, and Paleocene sediments of the Brightseat and Aquia Formations overlap the Patapsco Formation.

The three aquifers of the Waldorf aquifer system were modeled as three separate model layers. High leakance values were used to simulate breaches in the confining units of the Waldorf aquifer system. The hydraulic connections among these aquifers result in transmissivities greater than 10,500 ft^2/d (feet squared per day) in parts of the Waldorf area for the full thickness of the Waldorf aquifer system. The hydrologic boundaries formed by erosional truncation of the aquifers that make up the Waldorf aquifer system were simulated with no-flow conditions.

The La Plata aquifer system is an interconnected series of fluviodeltaic sands within the lower part of the Patapsco Formation at Waldorf. These sands are separated by confining units that differ areally in thickness and effectiveness as confining units. In contrast to the Waldorf aquifer system, however, the La Plata aquifer system was modeled as a single layer. A northeastward trend of increasing sand thickness within the Patapsco Formation in southern Maryland is the result of facies changes within the Patapsco Formation (Hansen, 1969). As part of this trend, sand thicknesses of the La Plata aquifer system show a significant increase from the southern to the northern parts of the Waldorf area, and, as a consequence, transmissivities increase from less than 500 ft^2/d to greater than 3,000 ft^2/d northeastward through the Waldorf area. In this paper, the name "La Plata aquifer system" refers only to the lower Patapsco aquifers present in north-central Charles and southern Prince Georges Counties, Md.

STRUCTURAL CONTROLS AND THEIR SIMULATION

Stratigraphic correlations show that a northeast-striking series of en echelon faults, the Brandywine fault system (Jacobeen, 1972), extends through central Charles County. Mixon and Newell (1977) and Mixon and Powars (1984) attributed the origin of the Brandywine fault system to reversal of movement, in response to regional compressional stress, along a preexisting zone of crustal weakness that was probably associated with a normal fault along the western edge of an extensive Triassic-Jurassic(?) basin. Faulting occurred during the Cretaceous, Paleocene, Eocene, and possibly later times.

A major fault that strikes northeast through the western part of Waldorf (fig. 19.1) is most likely a continuation of Jacobeen's (1972) Danville fault, which is part of the Brandywine fault system. Like the Danville fault, this fault is interpreted as an up-to-the-east, high-angle reverse fault having as much as 250 ft of throw at the pre-Cretaceous basement. Seismic sections by Jacobeen (1972) show probable bed offsets extending up through the lower part of the Patapsco Formation. Folding is apparently the dominant structural effect of the basement faulting on the upper part of the Patapsco, the Magothy and Severn Formations, and the pre-Miocene (pre-Calvert) Tertiary formations. Additional subsurface faults associated with the Brandywine fault system—two northeast-trending and two east-trending faults—have been identified in northern Charles and southern Prince Georges Counties by Lucy McCartan.

Calibration of the digital ground-water-flow model of the Waldorf area was improved by lowering the transmissivity (aquifer thickness times horizontal hydraulic conductivity) of the aquifers in a narrow row of model cells that coincide with the trace of a major fault within the Brandywine fault system (fig. 19.3). The model transmissivities along the trace of this fault are 50 to 90 percent less than the values in the adjacent model cells. Either the aquifer thickness or horizontal hydraulic conductivity must be reduced to lower transmissivity. The following fault-related mechanisms could possibly lower aquifer transmissivity.

Partial or total offset of sand beds along fault planes may impede the regional flow of ground water. Stratigraphic correlations and seismic sections (Jacobeen, 1972; Potomac Electric Power Company, 1973) indicate that bed offsets within the La Plata aquifer system are likely. Bed offsets may juxtapose aquifers and confining units; this juxtaposition would form at least a partial hydraulic barrier within the aquifer system that is equivalent to a major and rapid facies change in the aquifer. Hansen (1971) noted that a facies change that occurs over a short distance can constitute a hydrogeologic barrier that impedes the flow of recharge from the updip regions of an aquifer.

Bed offsets in the aquifers that compose the Waldorf aquifer system are possible, but they are not as likely as bed offsets in the stratigraphically lower La Plata aquifer system. Outcrops of both the Nanjemoy Formation and Marlboro Clay in Charles County show faults (J.D. Glaser, Maryland Geological Survey, oral commun., 1988). Also, Dryden (1932) described possible faults in outcrops of the Pliocene upland deposits near the town of Upper Marlboro.

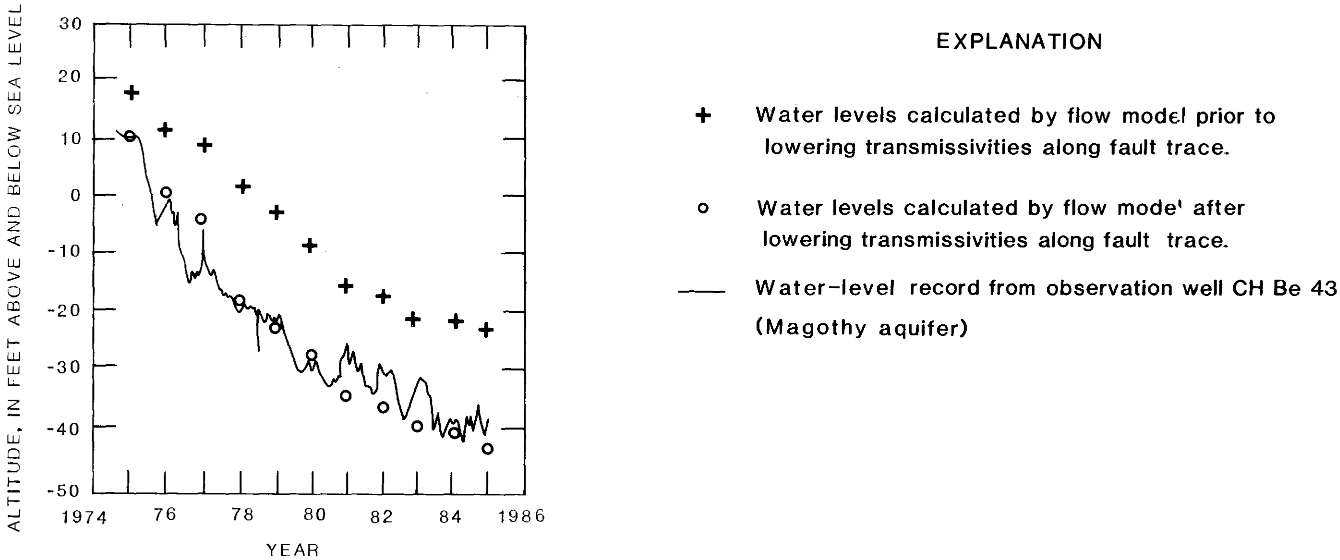

Figure 19.3. Comparison of model-calculated water levels and hydrograph from well CH Be 43, which is screened in the Magothy aquifer. The well location is shown in figure 19.1.

Two possible effects of faulting that could reduce aquifer thickness and cause reduced aquifer transmissivity are postdepositional erosion of upthrown areas or control of sedimentation patterns by faulting during deposition. The drill-hole data currently available are not sufficient to document any fault-related aquifer thinning. Near the boundary between Charles County and Prince Georges County, however, the Nanjemoy Formation thins by as much as 70 ft on the upthrown side of the fault shown in figure 19.1 relative to the downthrown side.

Compaction of sediments caused by faulting may reduce an aquifer's hydraulic conductivity by decreasing aquifer porosity. Some soft-sediment deformation and consequent alteration of aquifer fabric are probable within these faulted Coastal Plain sediments. Permeability tests of cores would be needed to quantify any effect of faulting on hydraulic conductivity.

CONCLUSIONS

Erosional truncation, sediment overlap, and facies changes among and within formations are key factors governing the flow and availability of ground water in the mid-Atlantic Coastal Plain aquifers. In the Waldorf area, these processes have resulted in a stratigraphically complex aquifer system—the Waldorf aquifer system—that consists of hydraulically connected sands in three different formations. Additionally, regional facies changes have significantly increased sand thickness and, consequently, transmissivity in the La Plata aquifer system at Waldorf.

Faulting of Coastal Plain sediments has the potential to affect aquifer transmissivity and influence ground-water flow by several processes. A hydraulic barrier may be formed by complete or partial juxtaposition of aquifers against confining units. Such a barrier is likely to exist in the La Plata aquifer system at Waldorf and is a possibility in the stratigraphically higher Waldorf aquifer system. Other possible causes of lowered transmissivity include thinning of aquifers as a result of fault-controlled sedimentation patterns or postdepositional erosion of uplifted blocks. Additionally, soft-sediment deformation and compaction of an aquifer during faulting may reduce transmissivity by lowering hydraulic conductivity through reduction of aquifer porosity.

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Southeastern Coastal Plains

20. Lithologic Changes in Fluvial to Marine Depositional Systems in Paleogene Strata of Georgia and Alabama

By Thomas G. Gibson[1]

INTRODUCTION

Most studies of Paleogene strata of the eastern Gulf of Mexico Coastal Plain have concentrated on west-central and western Alabama (Copeland, 1966; Jones, 1967; Toulmin, 1977) where the pioneering work by Smith, Johnson, and Langdon (1894) established many of the stratigraphic units. As part of the 1894 study, Langdon recognized some of their units along river sections in eastern Alabama. Little subsequent work was done in the Paleogene strata of eastern Alabama and westernmost Georgia until the study by Toulmin and LaMoreaux (1963). They considered the strata in eastern Alabama and western Georgia as a connecting link between the Gulf of Mexico and Atlantic Coastal Plains and recognized in the Chattahoochee River valley most of the stratigraphic units found farther west in Alabama. Shortly thereafter, Newton (1965, 1968) published maps of two of the counties in eastern Alabama along the Chattahoochee River. I began mapping Paleocene to middle Eocene strata from western Georgia (west of the Flint River) to eastern Alabama (east of the Pea River) in 1977 and continued through 1982.

Relatively high elevations of 500 to 650 ft are found in the Paleogene outcrop belt of eastern Alabama and western Georgia. Dissection by the Chattahoochee River and its tributaries, here flowing at elevations slightly above 100 ft, has resulted in high relief along the drainage system. The Paleogene strata have a southward dip of around 20–30 ft/mi; this gentle dip combined with the high relief results in an outstanding series of north-south exposures of each formation along about 30 mi. As the wide outcrop belt was in the transition area from marginal-marine to open-marine environments during the Paleogene, the change from nonmarine or marginal-marine sediments to open-marine facies is present in many of the formations. Furthermore, the larger outcrops may expose more than 90 ft of section in an area where the formations are generally less than 100 ft thick. Thus, several formational contacts may be seen within a single outcrop, and these exposures are a considerable aid in tracing lithologic changes. I used a U.S. Geological Survey (USGS) drilling rig equipped to take continuous cores in some critical locations in my mapping area in western Georgia and eastern Alabama during 1979 to 1982. The drill cores, commonly obtained in locations just downbasin from the outcrop exposures of a particular formation, were used to augment the outcrop sections. The coreholes usually penetrated through several formations and yielded a more complete thickness of some lithofacies and also better preserved paleontologic materials.

This paper summarizes the facies changes in nine formations; some of these changes were previously documented (Gibson, 1980, 1982a, b, c; Bybell and Gibson, 1985). As the facies changes typically involve the transition from marginal-marine environments in the north to open-marine environments in the south, I can locate the approximate present position and elevation of the shorelines for each Paleogene unit and note differences in their locations.

Acknowledgments

Laurel M. Bybell (USGS) helped greatly by providing calcareous nannofossil biostratigraphic data. I benefited from stimulating discussions with Juergen Reinhardt (USGS) on sedimentary structures and their use in paleoenvironmental interpretation.

FACIES CHANGES AND SHORELINE LOCATIONS

The study area extends from Clayton, Ala., on the northwest and Elba on the west to Blakely, Ga., on the southeast (fig. 20.1). Seven stratigraphic intervals containing nine formations were studied in this area (fig. 20.2). The Clayton Formation consists of marine facies over much of this area and contains marginal-marine strata only at high elevations near its northernmost outcrops. Limited

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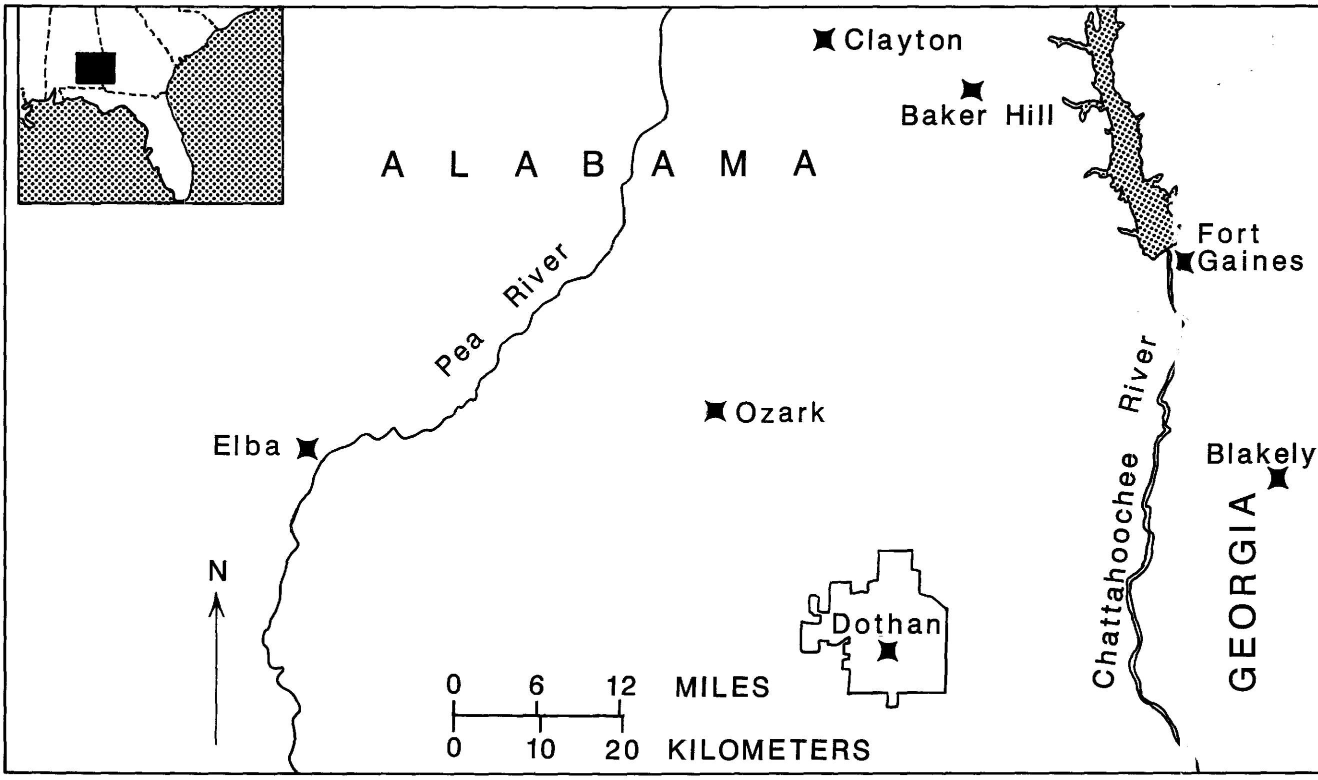

Figure 20.1. Chattahoochee River valley and surrounding study area in Alabama and Georgia.

exposures of the Porters Creek Formation are found only in upbasin localities and contain only marine facies; limited exposures of the Naheola Formation are found only in downbasin localities and contain only marginal-marine facies. The Nanafalia Formation contains marine facies, and its updip equivalent, the Baker Hill Formation, contains mostly marginal-marine facies. The Tuscahoma Formation is composed largely of marginal-marine facies except for one extensive marine sequence at the base. The Bashi Formation contains marine facies, and its updip equivalent, the Hatchetigbee Formation, contains mostly marginal-marine facies. The Tallahatta Formation contains both marine and marginal-marine facies.

The lithologies of the Paleogene units vary because of differences in the depositional environments in which they formed and also because of differences in the rate of supply of clastic sediments into the basin. Carbonate sediments are dominant in shallow-marine facies of the Clayton Formation; they were deposited during the early Paleocene when little clastic material was supplied to eastern Alabama. Carbonate sediments are essentially absent, however, from outcrop sections of the middle and upper Paleocene and lower Eocene units; these units were deposited when larger amounts of clastic material were supplied through deltaic input. Glauconitic sediments are common in the marine parts of the Tuscahoma Formation and throughout the Nanafalia and Bashi. Clay is the dominant clastic sediment in the Baker Hill Formation; it was deposited in marginal-marine environments. The fine sands of the Bashi and Hatchetigbee Formations rapidly thin downbasin; this thinning indicates that much of the sediment supply was trapped in shallow-marine environments, possibly because of a rapid rise in sea level during the time of deposition (Gibson, 1980). The Tuscahoma represents the major time of significant clastic input into this area; it thickens to nearly 200 ft of interlaminated clay and silt in downbasin outcrops (Toulmin and LaMoreaux, 1963).

In the following section of the paper, brief discussions of the lithologic facies in each formation are followed by the paleoenvironmental interpretations and inferences about the location and elevation of the shoreline. The ages of the formations and shoreline placements are shown in figure 20.2.

Clayton Formation

Downbasin outcrops of the Clayton along the Chattahoochee River near Fort Gaines consist almost entirely of locally sandy limestone of neritic origin (Toulmin and LaMoreaux, 1963). Upbasin, at elevations around 530 ft in the type area at Clayton, Ala., the lower part of the formation is dominated by medium-grained, well-sorted sand containing a low-diversity molluscan fauna. This basal

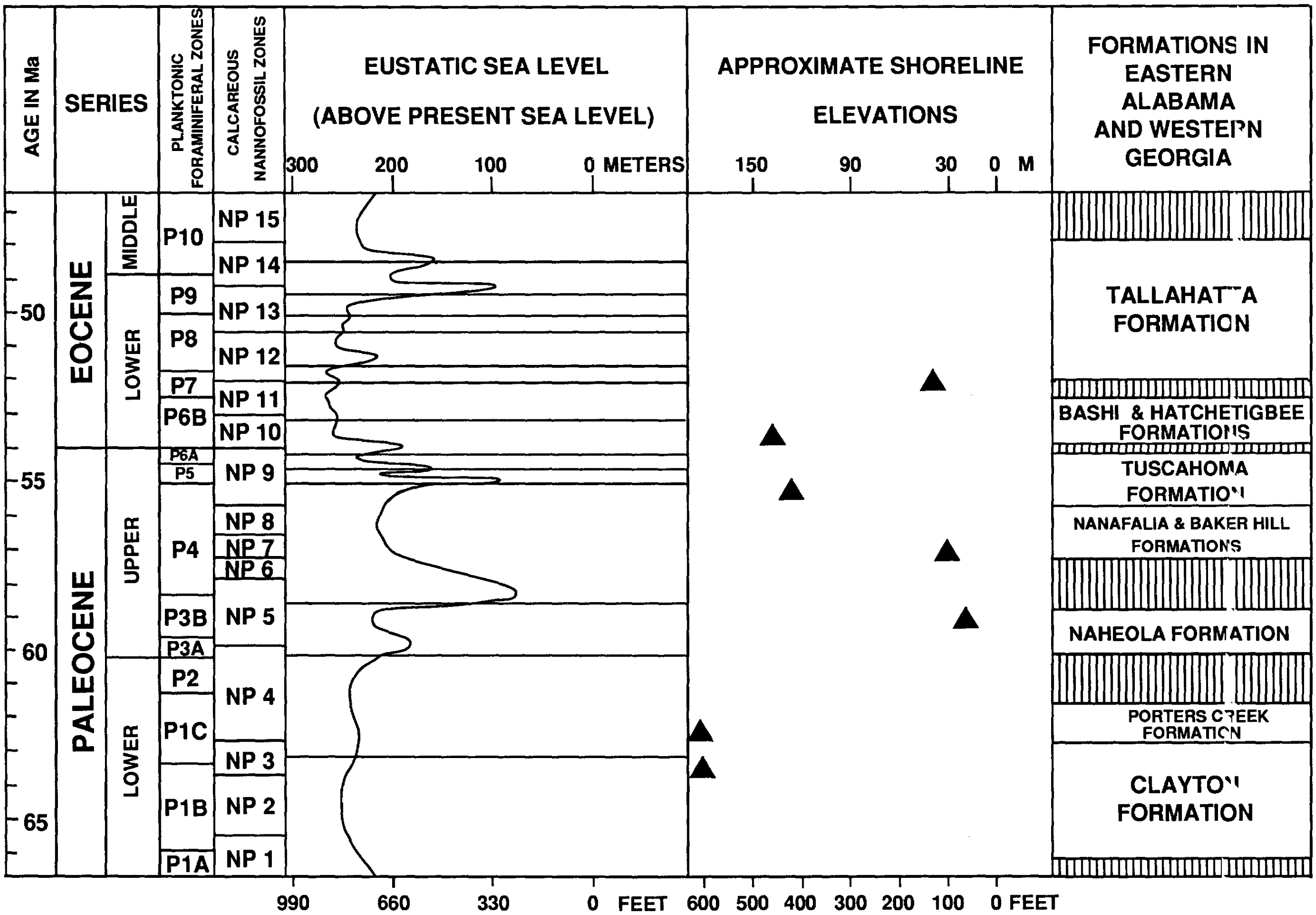

Figure 20.2. Correlation of Paleocene to middle Eocene formations and proposed shoreline elevations in the Chattahoochee River valley. Time scale, biostratigraphic zonation, and eustatic curves from Haq and others (1987).

sand is overlain by well-sorted sand containing shell banks largely composed of oysters and a clay bed with abundant leaf remains (Gibson, 1980). These deposits are suggestive of very shallow marine deposition with some restricted marine interfingering. To the northwest of Clayton at elevations between 600 and 650 ft, sand becomes the sole lithology; this sand may be crossbedded and contain *Ophiomorpha*. These outcrops suggest that the transition from marine to shoreface and marginal-marine environments during Clayton deposition occurred someplace around a present elevation of 600 to 650 ft.

Porters Creek Formation

The only exposures of the Porters Creek Formation in the study area are near Clayton, Ala. Apparently, most of this unit was eroded during the significant period of low sea level that followed its deposition. The only facies exposed is a waxy, dark, silty clay (Gibson, 1980). The lower part of this unit, exposed at 545 ft elevation, contains a foraminiferal assemblage indicative of neritic environmental depths of 120 to 180 ft or possibly slightly deeper. This interpretation would indicate a shoreline at elevations of 650 ft or slightly higher, similar to or slightly higher than that existing during deposition of the Clayton Formation.

Naheola Formation

The Naheola is also very restricted in its present distribution, being found only in sinkholes developed in the top of limestone of the Clayton near Fort Gaines. The micaceous fine sand and carbonaceous clay were deposited in brackish waters (Gibson and others, 1980) and indicate that the shoreline was at a lower elevation than the 110 ft at which these deposits are found. These beds were originally placed in the Gravel Creek Sand Member of the Nanafalia Formation (Gibson, 1980), but subsequent pollen studies suggest that they belong to the underlying Naheola Formation (N.O. Frederiksen, oral commun., 1988).

Nanafalia and Baker Hill Formations

The fossiliferous, glauconitic sands of the Nanafalia Formation are found in the Chattahoochee River valley at Fort Gaines at elevations of about 120 ft. The coarse clastic sediment, along with large lignitic logs and a dominantly oyster molluscan fauna, suggests that these deposits formed in very shallow marine environments. In outcrops and coreholes within 1–2 mi to the north of Fort Gaines, strata correlative with the glauconitic sands of the Nanafalia are gravelly sand containing large-scale crossbeds. Within another mile northward, the strata are largely kaolinitic and bauxitic clay interbedded with crossbedded medium sand. These crossbedded sand and kaolinitic clay beds were called the Baker Hill Formation (Gibson, 1982a). The Baker Hill Formation continues for about 20–30 mi to the northwest of Fort Gaines. In its more upbasin localities, it consists of gravelly sand in unidirectional large-scale crossbeds. The northward change from shallow-marine glauconitic sand of the Nanafalia through the crossbedded gravelly sand of the Baker Hill, considered to be shoreface deposits, to the kaolinitic clays, determined to be of marginal-marine origin (Gibson and others, 1980), occurs at elevations of about 120–130 ft which are, therefore, the approximate shoreline elevations.

Tuscahoma Formation

The basal, transgressive, gravelly, clay-clast-bearing glauconitic sand of the Tuscahoma is found at elevations around 435 ft south of Baker Hill. The lithology and the occurrence of abundant *Venericardia* and oysters suggest that these deposits are part of a shoreface complex and that the shoreline was nearby. A rapid upward change to interbedded silt and clay deposited in marginal-marine and then in nonmarine environments (Gibson and others, 1980) and a large increase in thickness downbasin suggest rapid progradation during latest Paleocene time.

Bashi and Hatchetigbee Formations

The basal fossiliferous, glauconitic sand of the Bashi Formation changes upbasin to crossbedded sand of the Hatchetigbee Formation. The Hatchetigbee outcrops of fine to medium sand containing multidirectional, small-scale crossbeds and clay drapes change northward to gravelly medium sand containing larger, unidirectional crossbeds. The latter deposits are interpreted as very nearshore bar sands and suggest a nearby shoreline location at elevations around 465 ft. As many as six depositional cycles can be observed in the Bashi along the Pea River near Elba and suggest significant fluctuations in sea level during deposition of this formation. The basal cycle discussed here represents the most extensive marine transgression of the Bashi.

Tallahatta Formation

The upbasin deposits of the Tallahatta consist of clay-clast-bearing, highly crossbedded, gravelly sand. Beds in the lower part of the formation do not show indications of marine deposition until elevations of 160 ft are reached northwest of Blakely. There, strata of slightly glauconitic medium sand contain fragmented shell material and suggest that the shoreline was nearby.

PALEOGENE SHORELINES

As biostratigraphic placement of these strata was given by Gibson and others (1982), the shoreline elevations listed here can be compared (fig. 20.2) with the worldwide eustatic curves in the Cenozoic cycle chart of Haq and others (1987). Times of concurrence between the relative highstands of sea level observed in this region and the eustatic sea-level curves are seen in the Clayton, Porters Creek, lower Tuscahoma, and Bashi Formations. Relative sea-level stands indicated here by the Nanafalia and Tallahatta Formations are lower than the eustatic sea-level curves of Haq and others (1987). These differences could be explained by local to regional upwarping or downwarping in the Chattahoochee area or by the need for better calibration in the eustatic curves. If uplifting or downwarping between or within the times of deposition of the formations in this area influenced the relative shoreline elevations, then comparison of these shoreline differentials with those indicated by other well-dated sequences of this age can give some history of vertical movements in this area. If only slight uplifting or downwarping occurred in this area during the Paleocene-middle Eocene, then this area can be extremely useful in determining the magnitude of sea-level changes that are difficult to determine from the seismic record alone.

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21. Correlation, Age, and Depositional Framework of Subsurface Upper Santonian and Campanian Sediments in East-Central South Carolina

By Gregory S. Gohn[1]

INTRODUCTION

The subsurface Cretaceous section of the South Carolina Coastal Plain consists of clay-, silt-, and sand-dominated units that were deposited in a variety of paleoenvironments during the Cenomanian through the Maastrichtian (Valentine, 1982; Colquhoun and others, 1983; Owens and Gohn, 1985; Sohl and Owens, 1991). Herein, the stratigraphy and depositional history of the upper Santonian and Campanian parts of this section in east-central South Carolina are described briefly.

In the present study area (fig. 21.1), the upper Santonian and Campanian sections display obvious lateral and vertical facies changes within cyclic, unconformity-bounded stratigraphic units. Unit contacts typically juxtapose nearshore sands and silts at the top of one unit beneath deeper water clay-silts or glauconitic sands at the bottom of the overlying unit. These contacts can be traced throughout the study area despite lateral facies changes, and they provide a basis for defining five informal stratigraphic units that are both lithostratigraphic (mappable on the basis of lithology) and allostratigraphic (unconformity bounded). These units were deposited during transgressive-regressive cycles; each has deeper water marine beds at the base and relatively nearshore marine deposits at the top. The lithofacies patterns within these units are typical of the deltaic and delta-related sections that characterize the Cretaceous of the Atlantic Coastal Plain (Owens and Gohn, 1985). Regionally, the lithologies, stratigraphic positions, and ages of the five units suggest at least partial correlation with the deposits historically assigned to the outcropping Black Creek Formation of the Carolinas (Swift and Heron, 1969; Sohl and Owens, 1991).

The drill-hole data base for the present study area is very limited, particularly when compared to the data available for hydrocarbon-producing basins (for example, see the papers selected by Roy, 1980). Most water wells in east-central South Carolina bottom in Paleocene or Eocene aquifers, and only a few municipal and industrial wells penetrate the deeper Cretaceous beds. Electric and gamma logs typically are the only material available for study from many water wells, although additional geophysical logs, drillers' logs, and cuttings are available for some wells. The water-well data are supplemented, however, with data from a few stratigraphic and (or) hydrologic test holes. Two continuously cored stratigraphic test holes in the western part of the area (fig. 21.1), USGS-Clubhouse Crossroads 1 (CC1) and USGS-St. George 1 (StG1), are the cornerstones of the present study.

Despite its shortcomings, the existing data base is adequate for regional assessments of thickness and lithology trends although details are generally lacking except locally. In addition, depositional facies can be inferred from the geophysical logs for individual wells and then extrapolated to infer regional facies trends.

SECTIONS IN CLUBHOUSE CROSSROADS 1 AND ST. GEORGE 1

The continuously cored sections in CC1 and StG1 provide most of the direct lithologic information for the studied sections and nearly all the biostratigraphic information. Both holes were drilled into dominantly fine grained, calcareous marine beds of the upper Santonian and Campanian units; therefore, they provide a considerable amount of biostratigraphic data, much of which has been published for CC1 (Hazel and others, 1977; Hattner and Wise, 1980). The section in CC1 has been included in most regional stratigraphic and geohydrologic studies of the area (for example, studies by Brown and others, 1979; Colquhoun and others, 1983; Gohn, 1988).

Five stratigraphic units have been identified as summarized in table 21.1. In CC1, units 1, 4, and 5 are individually about 100 to 150 ft thick and consist entirely of marine clay-silts (fig. 21.2). Each unit grades upward from silty clays at the bottom (distal prodelta) to slightly coarser clayey silts and muddy fine sands (proximal prodelta) at the top. Eastward from CC1, the upper parts of these units become distinctly coarser grained, and well-sorted delta-

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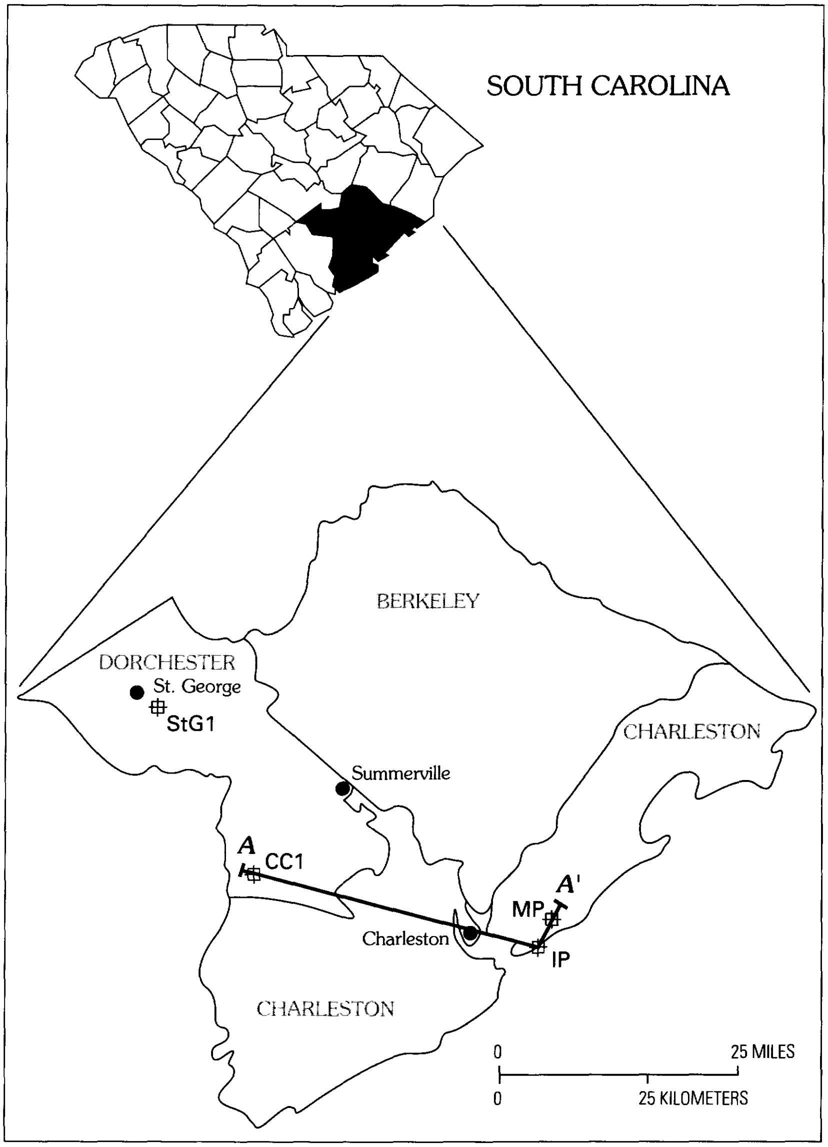

Figure 21.1. Drill holes in east-central South Carolina mentioned in this report. Cross section *A–A'* is shown in figure 21.4. Abbreviations for drill-hole names: StG1 = USGS-St. George 1, CC1 = USGS-Clubhouse Crossroads 1, IP = Isle of Palms water well, and MP = Morgan Point water well.

front sand bodies dominate the upper part of each unit. Sands do occur at the top of unit 3 at CC1; there unit 3 consists of clay-silts in its lower half (prodelta) and well-sorted quartz sand containing *Ophiomorpha* at the top (delta-marginal barrier?). Unit 2 consists of clayey glauconite-quartz sand in which calcareous nodules are common. Unit 2 is interpreted herein as erosionally truncated; nearshore beds at the top of the unit have been removed, and only the deeper water marine lower part of the unit remains.

In StG1 (fig. 21.3), units 3 and 4 consist of prodelta clay-silts. However, unit 5 is represented by a relatively thin nearshore section of poorly sorted clayey sands and sandy clays, much of which contains a restricted ostracode fauna. Units 1 and 2 cannot be traced lithologically (or faunally) from CC1 to StG1 and are considered to be absent from StG1.

Planktic faunas and floras in the two corehole sections define the Santonian and Campanian ages of these sediments (table 21.1). At CC1, the presence of *Eprolithus floralis* and a member of the *Aspidolithus parcus* group within unit 1 (table 21.1), as well as the fossils in unit 2, suggests that the Santonian-Campanian boundary can be placed near 1,700 ft within unit 1. The last appearance of *E.*

Table 21.1. Upper Santonian, Campanian, and lower Maastrichtian datums in the USGS-Clubhouse Crossroads 1 and USGS-St. George 1 wells, South Carolina

[Fossil abbreviations following species names: cn, calcareous nannofossils; m, mollusks; os, ostracodes; and pf, planktic Foraminifera. Depths in feet below the surface. For USGS-St. George 1, C.W. Poag and W.A. Bryant (both of the U.S. Geological Survey) identified the planktic Foraminifera, and P.C. Valentine (USGS) identified the calcareous nannofossils. The fossil data from USGS-Clubhouse Crossroads 1 are principally from Hazel and others (1977) and Hattner and Wise (1980). Units 1 and 2 are missing from St. George 1]

	CC1	StG1
Lower Maastrichtian datums:		
Lowest occurrence of *Globotruncana aegyptiaca* (pf)	804	1,000
Highest occurence of *Flemingostrea subspatulata* (small form) (m)	1,030	not studied
Lowest occurrence of *Exogyra costata* (m)	1,066	not studied
Campanian datums:		
Top of unit 5	**1,072**	**1,031**
Lowest occurrence of *F. subspatulata* (small form) (m)	1,090	not studied
Highest occurrence of *Exogyra ponderosa* (m)	1,162	not studied
Top of unit 4	**1,214**	**1,082**
Highest occurrence of *Fissocarinocythere gapensis* (os)	1,263	1,095
Highest occurrence of *Haplocytheridea insolita* (os)	1,263	1,095
Lowest occurrence of *H. insolita* (os)	1,338	1,168
Top of unit 3	**1,344**	**1,177**
Highest occurrence of *Fissocarinocythere pittensis* (os)	1,410	1,260
Highest occurrence of *Ventilabrella glabrata* (pf)	1,457	not found
Lowest occurrence of *F. gapensis* (os)	1,480	1,280
Lowest occurrence of *Quadrum sissinghii* (cn)	not found	1,320
Lowest occurrence of *Ceratolithoides aculeus* (cn)	1,500	1,320
Lowest occurrence of *Rugoglobigerina rugosa* (pf)	1,504	1,320
Top of unit 2	**1,544**	**missing**
Highest occurrence of *Marthasterites furcatus* (cn)	1,560	not found
Lowest occurrence of *Ventilabrella glabrata* (pf)	1,600	not found
Top of unit 1	**1,644**	**missing**
Lowest occurrence of *Aspidolithus parcus* group (cn)	1,690	1,320
Upper Santonian datums:		
Highest occurrence of *Eprolithus floralis* (cn)	1,713	not found
Sole occurrence of *Heterohelix reussi* (pf)	1,713	not found
Top of Middendorf Formation	**1,754**	**1,325**

floralis is generally placed at the Santonian-Campanian boundary or within the late Santonian (Perch-Nielsen, 1985, top of Zone CC16). The presence of *Marthasterites furcatus* with a member of the *Aspidolithus parcus* group indicates an early Campanian age (Zone CC18 of Perch-Nielsen, 1985) for unit 2 at Clubhouse Crossroads.

The presence of *Quadrum sissinghii*, *Ceratolithoides aculeus*, and *Rugoglobigerina rugosa*, as well as the absence of *M. furcatus*, in unit 3 in both cores indicates an early to middle Campanian age for unit 3. This distribution also suggests that units 1 and 2 found in CC1 are missing at St. George. In addition to the nannofossil data, ostracode data are useful; the interval containing the overlap of the ranges of the ostracodes *Fissocarinocythere pittensis* and *F. gapensis* occurs at the same stratigraphic position within unit 3 in both drill holes.

The planktic fossils in units 4 and 5 provide less resolution than those in units 2 and 3, although they are compatible with a Campanian age. The highest occurrences of the ostracodes *Fissocarinocythere gapensis* and *Haplocytheridea insolita* in unit 4 (with *Antibythocypris elongata* in CC1) indicate a late Campanian age for that unit (Hazel and Brouwers, 1982). *Haplocytheridea insolita* is restricted to unit 4 in both wells. The Campanian-Maastrichtian boundary is placed at the top of unit 5 in CC1 on the basis of the lowest occurrence of *Exogyra costata* just above the top of unit 5 and the distribution of the small form of *Flemingostrea subspatulata* (see Sohl and Christopher, 1983). The Maastrichtian species *Globotruncana aegyptiaca* first occurs about 30 ft above the top of unit 5 in StG1. On the basis of its stratigraphic position, unit 5 is necessarily of late Campanian age.

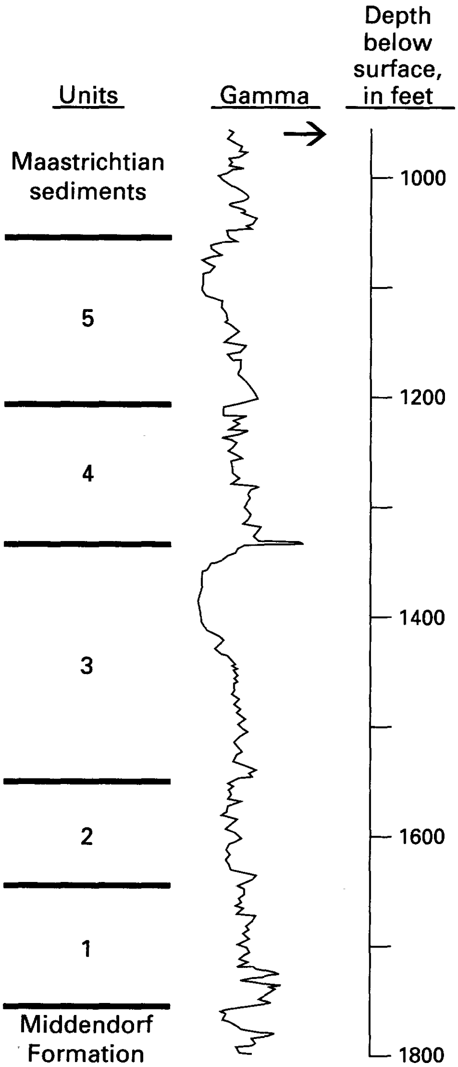

Figure 21.2. Gamma log for the studied section in USGS-Clubhouse Crossroads 1, Dorchester County, S.C. The boundaries of Santonian–Campanian unit 1 and Campanian units 2 through 5 are shown.

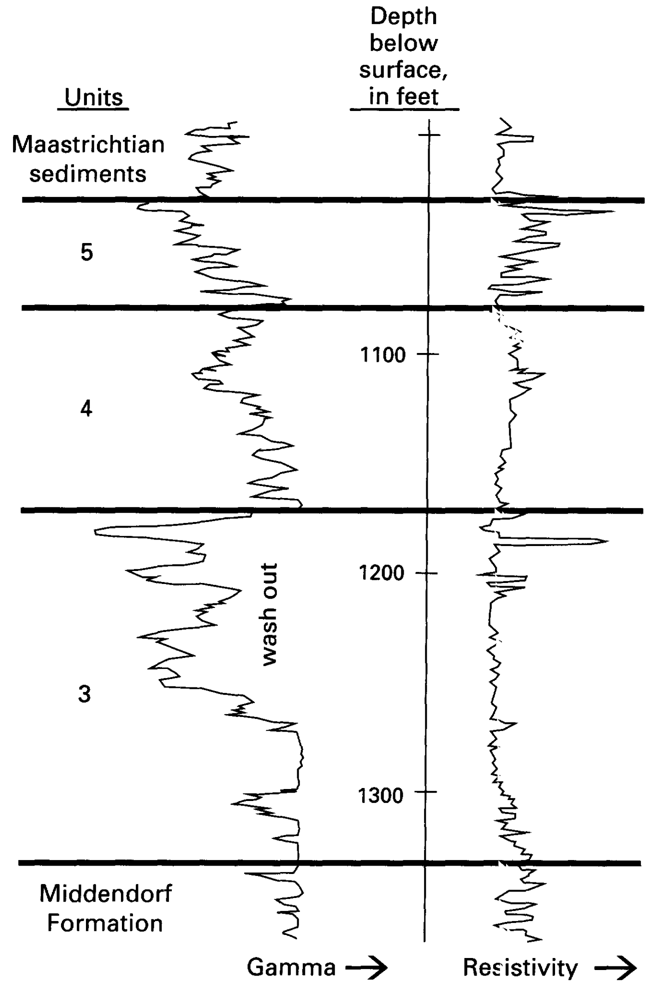

Figure 21.3. Gamma and electrical resistivity logs for the studied section in USGS-St. George 1, Dorchester County, S.C. The boundaries of Campanian units 3 through 5 are shown.

EXAMPLE OF REGIONAL FACIES DISTRIBUTION

The distribution of depositional lithofacies in unit 4 is described in this section as a representative example of the facies patterns found in the studied sequences. As seen in figures 21.3 and 21.4, unit 4 is thickest in the eastern part of the study area, as represented by the Isle of Palms (IP) and Morgan Point (MP) wells, and thins to the west and northwest. Similarly, the collective thickness of well-sorted sand bodies within unit 4, as interpreted primarily from geophysical logs, also is greatest in the east and decreases to the west. Well-sorted sands are absent from unit 4 in the CC1 and StG1 sections.

The geophysical logs for the IP and MP wells show a strong differentiation of sediments into clay- and sand-dominated intervals (fig. 21.4). In these wells, unit 4 coarsens upward from a basal clay-silt into two or more sand bodies separated by thinner clay-silt layers. The sands tend to be well sorted and to have sharp upper and lower contacts.

The sand-dominated sections are interpreted to represent delta-front and proximal prodelta deposition. The lower, finer grained parts of the sections represent the initial progradation of the prodelta above a basal transgressive unconformity. The well-sorted sands having sharp contacts are delta-front deposits, possibly barriers formed by wave redistribution of river-supplied sediments (see examples in Roy, 1980). To the west, the fine-grained sections lacking major sand bodies represent prodelta deposits on the margin of the delta lobe(s).

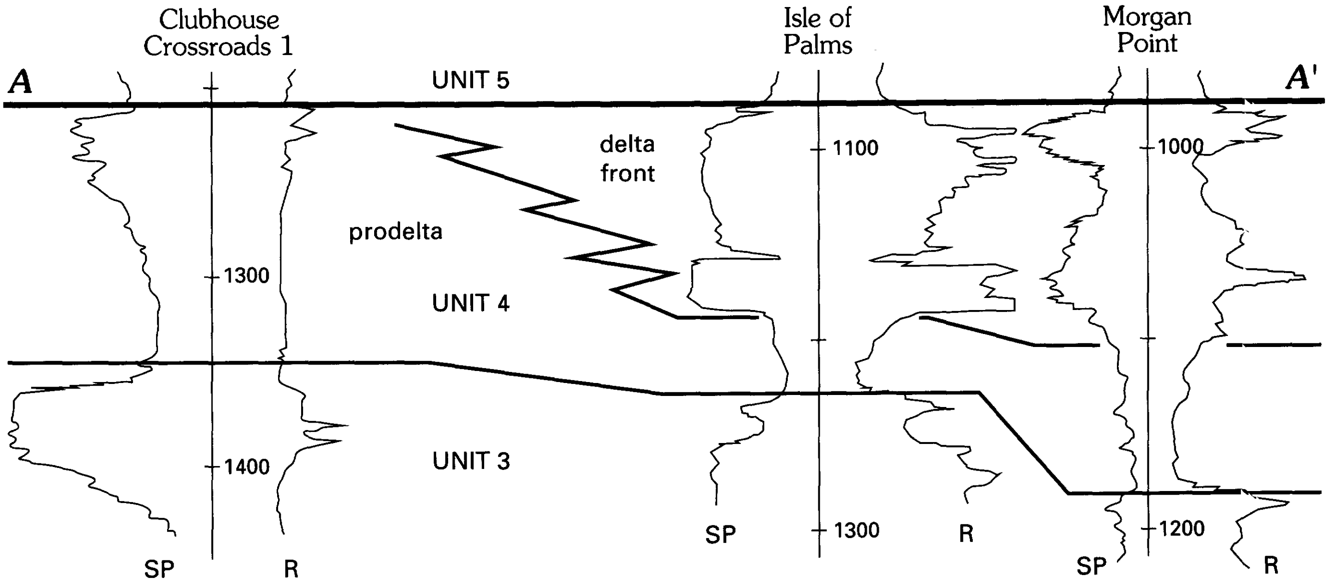

Figure 21.4. Distribution of facies in unit 4. Location of the section is shown in figure 21.1. Electric logs are shown for each well: SP = spontaneous potential log, R = resistivity log. Depths are in feet.

SUMMARY

Although generalized herein, a relatively complete picture of the depositional history of the studied section has evolved through the integration of sedimentary facies analysis with modern biostratigraphic zonations. Five unconformity-bounded, lithologically distinct units containing diagnostic late Santonian and Campanian fossils can be traced in the subsurface of east-central South Carolina. Each unit contains a mosaic of sedimentary facies that represents initial marine transgression and subsequent delta progradation. The recognition of five stratigraphic units within an interval that is traditionally assigned to only one formation (Black Creek) suggests a greater stratigraphic complexity than is indicated by the traditional nomenclature.

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22. Magnesium-Rich Clay Minerals in Tertiary Carbonate Rocks of Southwestern Florida[1]

By Lucy McCartan,[2] A.D. Duerr,[3] *and* R.M. Hawkins[2]

INTRODUCTION

This report describes our research on the clay mineralogy of rocks in De Soto County, Fla., sampled in two holes (fig. 22.1). We examined and sampled the uppermost 440 ft of the 829-ft-long core from the Regional Observation and Monitor-Well Program (ROMP) 17 corehole, and we obtained cuttings from the 835-ft-deep Dees well about 10 mi southeast of the ROMP 17 corehole.

The ROMP 17 and Dees wells are stratigraphic control sites and ground-water monitoring wells drilled by the Southwest Florida Water Management District (SWFWMD) and the U.S. Geological Survey (USGS), respectively. Prior to the present study, the clay minerals in these wells had not been examined in detail. This report gives some of the preliminary results of a cooperative study to determine the ground-water resources, ground-water quality, and mineralogy of the surficial units and subsurface units in the area.

Clay minerals in the ROMP 17 core are restricted almost exclusively to dolomitic rocks. They occur in discrete beds and are disseminated in semi-indurated to well-indurated dolomitic rocks. Two of the clay minerals, palygorskite and sepiolite, are magnesium rich; also, a poorly crystalline phase referred to informally as "RE" in this report, has both refractory and expandable layers and may contain magnesium. Limestones in the same core have no clay minerals.

The clay-mineral distribution is interesting for three reasons. First, sepiolite and palygorskite are highly adsorbent; both are used commercially to adsorb organic pollutants. We infer that naturally occurring sepiolite and palygorskite function as ground-water filters. Where they occur in discrete beds or where they compose more than 15 percent of the total rock, they also function as leaky confining beds. Second, the clear pattern of clay minerals in dolomite and not in limestone in the ROMP 17 core is dissimilar to the mixed mineralogy in a suite of cuttings from the Dees well. The mixing of minerals in the Dees well appears to be an artifact of the drilling process. Finally, Mg-rich clay is a significant clue to the depositional and postdepositional history.

Acknowledgments

A.D. Duerr was partly supported by the SWFWMD. The SWFWMD is also responsible for the Regional Observation and Monitor-Well Program. We thank the Florida Geological Survey for access to the ROMP 17 core and for samples. Jonathan Arthur arranged for the core sampling, and Thomas Scott provided a detailed, unpublished lithologic log of the 829-ft-long core. We appreciate comments by Ruth Deike, John Hosterman, and Milan Pavich (all of the USGS) on earlier drafts of this report.

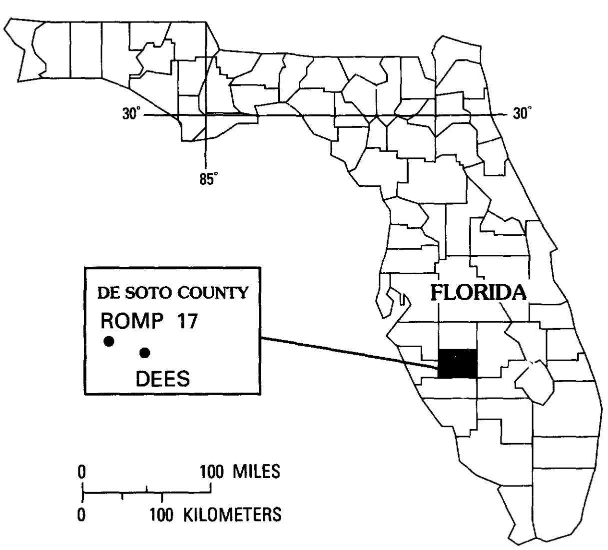

Figure 22.1. Locations of the ROMP 17 corehole and the Dees well, De Soto County, Fla.

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[1]Prepared in cooperation with the Florida Geological Survey and the Southwest Florida Water Management District.

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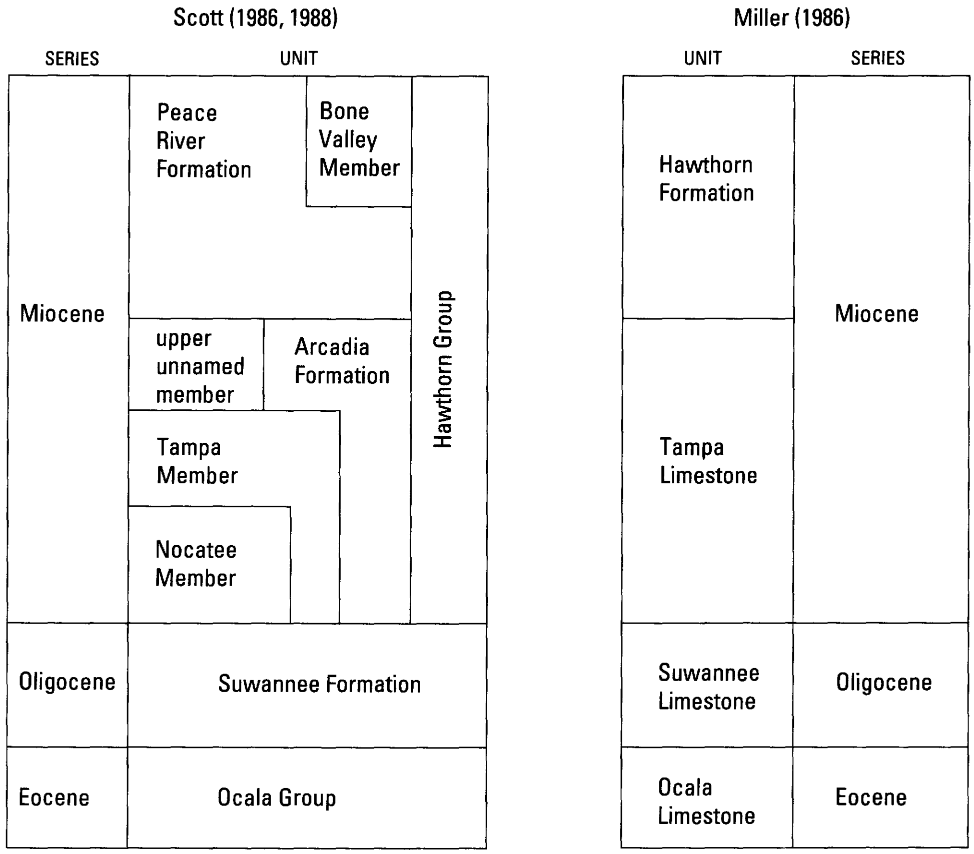

Figure 22.2. Stratigraphic nomenclature for the upper Eocene–Miocene series in southwestern Florida.

STRATIGRAPHY

The part of the ROMP 17 core that was studied sampled four units—the Hawthorn Formation (Miocene), the Tampa Limestone (Miocene), the Suwannee Limestone (Oligocene), and the Ocala Limestone (Eocene)—according to an unpublished log by Thomas M. Scott (Florida Geological Survey, 1983), which followed the nomenclature of Miller (1986). Scott's published stratigraphic nomenclature for the ROMP 17 core (in Johnson, 1986; see also Scott, 1986, 1988) is as follows:

- Hawthorn Group, 26–439 ft
  - Peace River Formation, including the Bone Valley Member, 26–99 ft
  - Arcadia Formation, 99–439 ft
    - upper unnamed member, 99–208 ft
    - Tampa Member, 208–292 ft
    - Nocatee Member, 292–439 ft
- Suwannee Formation, 439–829 ft

The simpler scheme of Miller (1986) is used in this report. The two schemes are compared in figure 22.2.

Geophysical logs indicate that the same formations (according to Miller's (1986) nomenclature) are in both the ROMP 17 and Dees wells. The contacts are deeper in the Dees well because the regional structure dips toward the south.

METHODS AND MINERALOGY

The upper 440 ft of the ROMP 17 core (fig. 22.3) was examined, and small samples were taken for X-ray diffraction (XRD) analysis and optical microscopy. X-ray diffractograms of the core samples (fig. 22.4) were compared with those of cuttings from the Dees well (fig. 22.5). Core samples were lightly crushed and attached to glass slides with amyl acetate. This method was necessary because of the limited amount of material available from the core. Cuttings from the Dees well were disaggregated or lightly crushed in deionized water to make a slurry. The clay-sized fraction (containing particles smaller than 2 μm) was separated by centrifuging the slurry. The clay-sized particles were then sedimented onto a glass slide, air dried, and X-rayed with CuK_{α_1} radiation. Because clay and nonclay minerals are in both suites of samples, the effects of the different procedures appear to be minimal.

Untreated samples that had diffraction peaks lower than 8.8 Å (10° 2Θ) were treated and then X-rayed again.

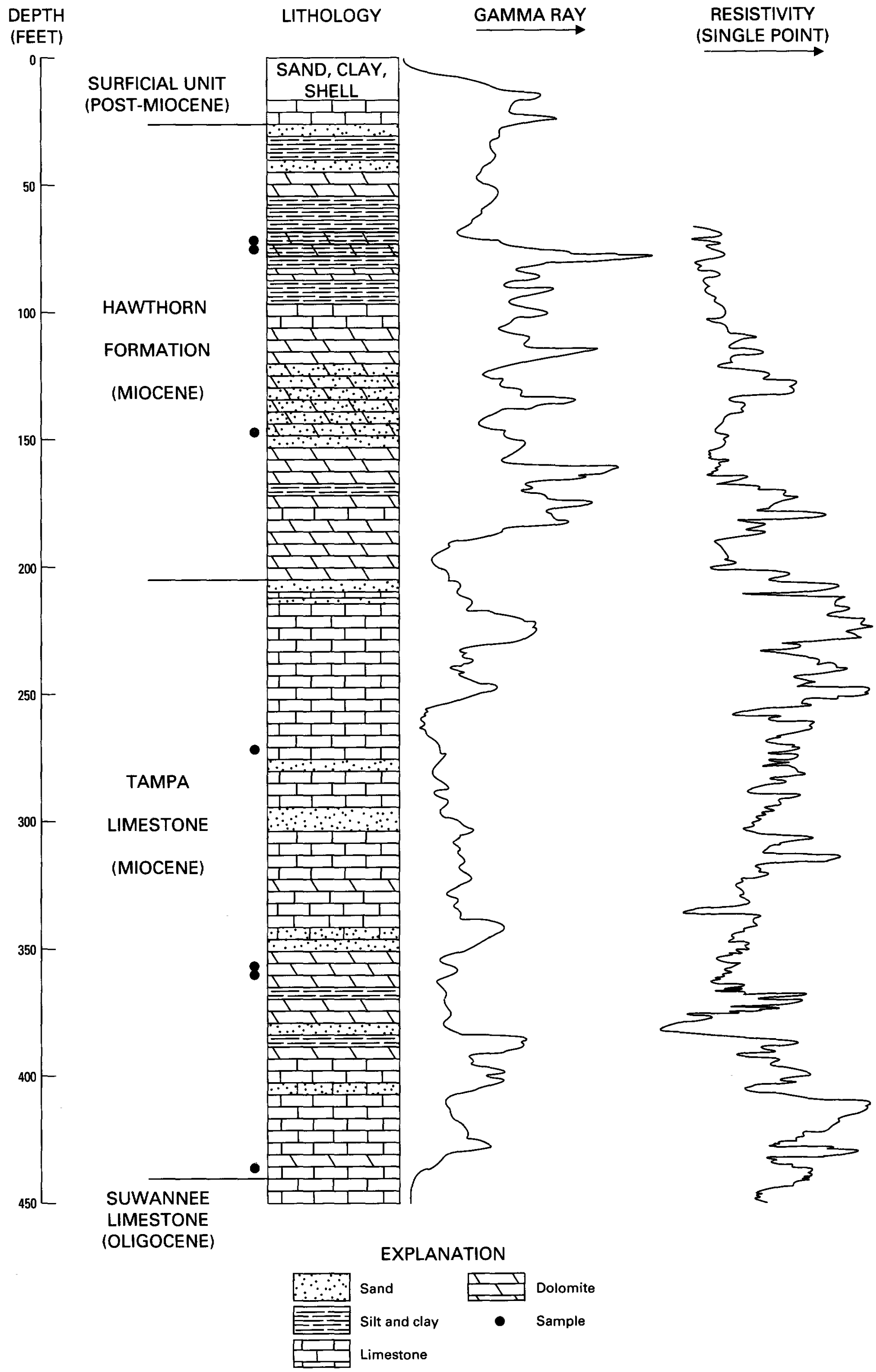

Figure 22.3. Lithologic, gamma-ray, and resistivity logs and stratigraphy of the ROMP 17 corehole. Gamma-ray and resistivity logs are from the files of the Southwest Florida Water Management District (SWFWMD). Lithology is from the SWFWMD and from core samples. Sand, silt, and clay in this core typically contain carbonate minerals. Formational contacts are from an unpublished log (dated 1983) provided by Thomas M. Scott, Florida Geological Survey; nomenclature is from Miller (1986).

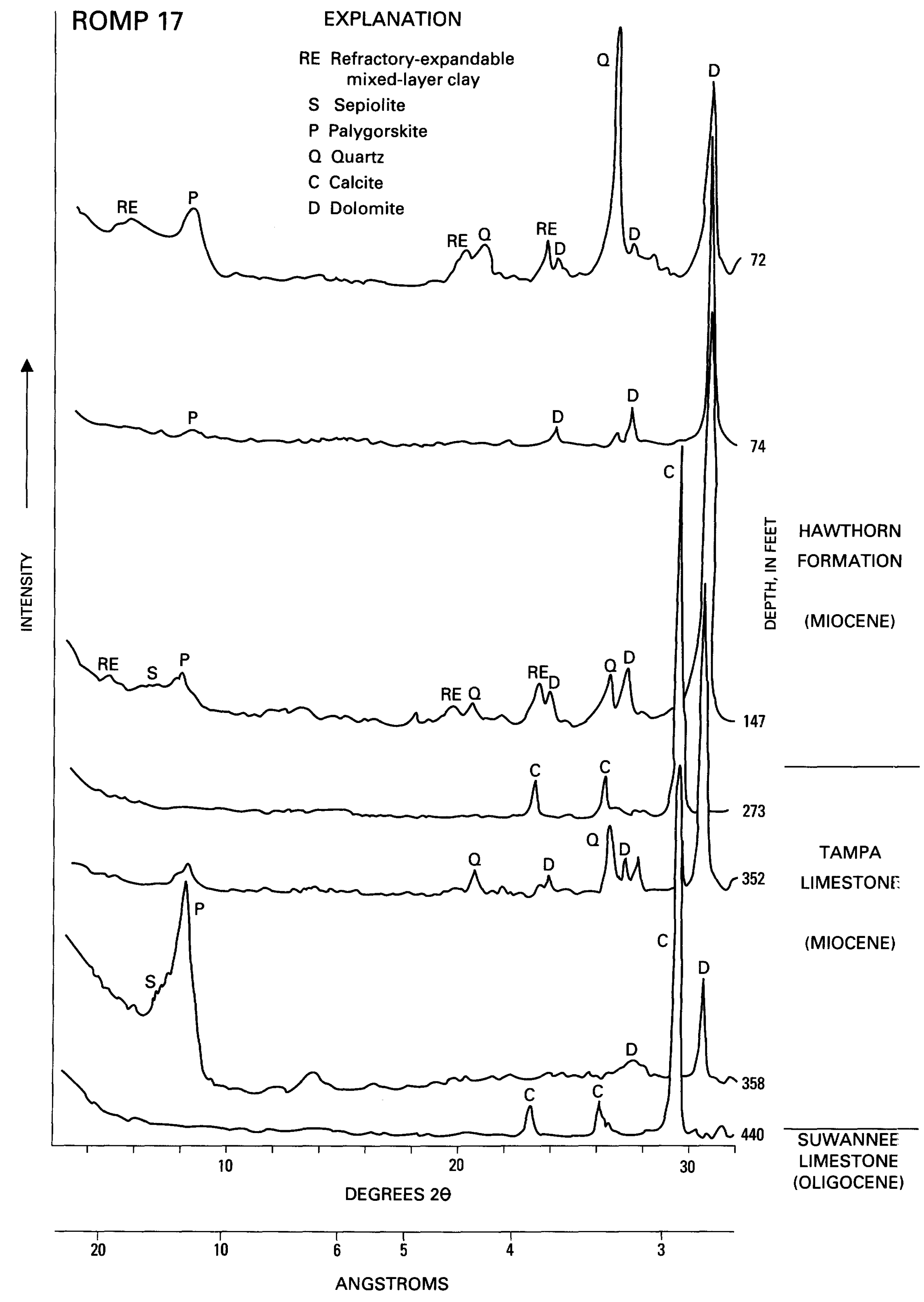

Figure 22.4. X-ray diffractograms of minerals in lightly crushed samples from the ROMP 17 core. Samples with diffraction peaks lower than 8.8 Å (10° 2Θ) were glycolated. See figure 22.1 for location of well.

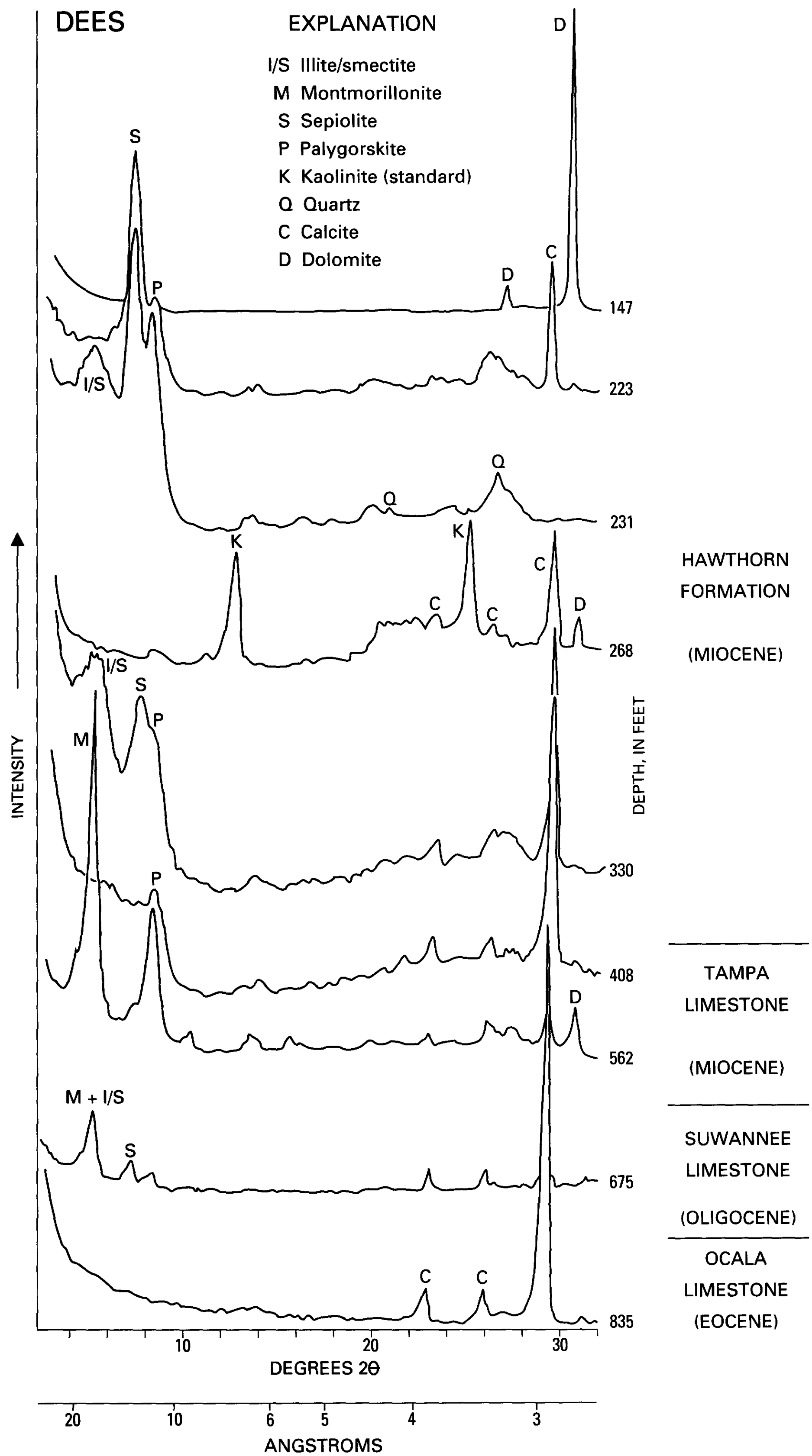

Figure 22.5. X-ray diffractograms of minerals in the clay-sized fraction (less than 2 μm) of cuttings from the Dees well. Samples with diffraction peaks lower than 8.8 Å (10° 2Θ) were glycolated. See figure 22.1 for location of well.

One slide was heated to 350 °C for 1 hour; another slide was vapor glycolated overnight and X-rayed again.

After slides were heated, diffractograms showed that all the peaks in the range from 10 to 35 Å (8.8°–2.5° 2Θ) for untreated samples became somewhat broader except those of illite/smectite mixed-layer clay and montmorillonite. These peaks collapsed to about 10 Å (8.8° 2Θ). Because of the large proportion and good crystallinity of montmorillonite (drilling mud) in the samples from −562 ft and −675 ft in the Dees well, the 10-Å peak is distinct. Heating of the small proportion of poorly crystalline illite/smectite in samples from −231 ft and −330 ft in the Dees well produced a low 10-Å peak that is masked under the large shoulder of the combined sepiolite and palygorskite peaks. The shifting of the peaks is due to the escape of water loosely held in the structure and the subsequent collapse of basal planes to the illite d-spacing (10 Å). Retention of some, but not all, of the intensity of sepiolite, palygorskite, and RE peaks reflects the presence of both refractory layers and water layers.

Increase of basal-plane spacing due to absorption of ethylene glycol, from about 14.7 Å (6° 2Θ) to about 17.7 Å (5° 2Θ) occurred only in illite/smectite mixed-layer clay and montmorillonite (fig. 22.5). Minerals were identified by comparing peaks on the diffraction traces from untreated samples with those listed in the Joint Committee on Powder Diffraction Standards (1974, 1981) data book. Illite/smectite was distinguished from montmorillonite on the basis of peak shape and the height of the low-angle saddle (Reynolds and Hower, 1970). Illite/smectite in these samples has a broad, low, irregular peak and a high saddle on the lower angle side of the peak. Montmorillonite has a sharp, narrow peak and a low saddle. All the minerals except quartz permit significant ionic substitution, and so minor discrepancies between the published X-ray patterns and those in figures 22.4 and 22.5 are assumed to be a result of differences in ionic substitution.

Minerals from three units in the ROMP 17 core between −17 ft and −440 ft are identified on the XRD trace (fig. 22.4). Cuttings from the Dees well between −147 and −835 ft were X-rayed (fig. 22.5); the Ocala Limestone was sampled in both wells, but data are available at present only from the Dees well. Mineralogy of the cuttings is far more mixed than that in the core; there appears to be little partitioning of Mg-bearing minerals in Dees well samples. Nevertheless, the cuttings gave at least a general picture of the most important minerals—low-Mg calcite, dolomite, palygorskite, and sepiolite.

ADSORPTION BY PALYGORSKITE AND SEPIOLITE

Palygorskite ($Mg_5Si_8O_{20}(OH)_2(H_2O)_4 \cdot 4H_2O$) and sepiolite ($Mg_8Si_{12}O_{30}(OH_2)_4(H_2O)_4 \cdot 8H_2O$) are similar fibrous clays. Their large capacity to adsorb various compounds is related to their very fine grained fibrous habit. Because their structures have channels, it was once assumed that they would absorb molecules in the channels. However, Barrer (1978) has shown that, although water and ammonia may enter the channels after heat treatment, other materials such as nitrogen and alcohols are confined to the outer part of the structures. Moreover, fibrous clays selectively adsorb certain hydrocarbons. Palygorskite has long been used as an adsorbing medium for organic fluids (Patterson, 1974), and surface-modified sepiolite is also now available commercially. Research is continuing into modifications that enhance the selectivity and volumetric aspects of sepiolite sorption (Alvarez and others, 1987; Pérez-Castells and others, 1987).

The degree to which untreated fibrous clays in natural environments filter ground water and modify ground-water chemistry is not well known. In areas such as southwestern Florida, where potable ground-water supplies are being depleted, an understanding of the behavior of such minerals is essential.

COMPARISON OF CORE SAMPLES AND CUTTINGS

The ROMP 17 corehole and the Dees well are about 10 mi apart, and they intersect the same units. Yet the mineralogic pattern appears to be significantly different in the two wells. In the ROMP 17 core, nearly pure limestone consisting of low-Mg calcite alternates with dolomite that contains palygorskite, sepiolite, quartz, and RE, a phase that has both expandable and refractory layers. The cuttings from the Dees well are a mixture of drilling mud (termed "bentonite," but mineralogically almost pure montmorillonite, at −562 and −675 ft), dolomite, calcite, kaolinite, palygorskite, sepiolite, and illite/smectite. The only "pure" limestone in the Dees well occurs in the Ocala Limestone; perhaps by that depth, the recirculation of clays from higher units and drilling mud had ceased (water was used below −675 ft instead of drilling mud). However, the sample at −147 ft in the Dees well is a clay-free dolomite that has no counterpart in the ROMP 17 core.

Some variation between wells 10 mi apart is to be expected. Depositional lithofacies in carbonate banks change over short distances (see section below on paragenesis). However, it is equally likely that the mixed mineralogy of the cuttings is an artifact of the drilling process. More work on cores from southwestern Florida will enable us to clarify this point. Until then, relating the mineralogy of cuttings to ground-water quality is probably inadvisable.

PARAGENESIS

Partitioning of Mg-rich carbonate minerals and Mg-rich clay minerals in Florida is unlikely to be accidental. Such partitioning also occurs in Cretaceous deposits in

Texas (Deike, 1991). The distribution of dolomite and calcite may be related genetically to the distribution of clay minerals. Similarly, the diagenetic processes that led to dolomitization of the primary biogenic calcium carbonates (Friedman, 1964; Shinn, 1972; Plummer, 1977; Land, 1982; Plummer and others, 1983; Thayer and Miller, 1984) may have been partly responsible for the presence of magnesium in the clays (Weaver and Beck, 1977; Fisher, 1988).

The sequence as a whole is typical of modern deposits in carbonate banks and sabkhas (Miller, 1986), which accumulate in warm, shallow marine water and adjacent supratidal flats. However, modern formation of Mg-rich clay has been reported only in association with submarine volcanic vents (Bowles and others, 1971). The Mg-rich clay minerals in southwestern Florida may have resulted from localized concentration of magnesium through processes such as evaporation of seawater (Weaver and Beck, 1977; Fisher, 1988), dissolution of volcanic ash derived from the Caribbean (Case and Holcombe, 1980) or the west (Gilluly, 1965), or recrystallization of high-Mg calcite (Chave, 1954) or dolomite (Plummer, 1977; Plummer and others, 1983).

CONCLUSIONS

Palygorskite and sepiolite are Mg-rich clays that tend to adsorb organic molecules in ground water that flows through units containing them. Their distribution may be an important consideration in the assessment of ground-water quality and pollution potential in southwestern Florida.

Lithology, mineralogy, and, therefore, stratigraphy are more accurately represented by core samples than by cuttings. Projects need to be designed with this difference in mind.

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23. Celestine ($SrSO_4$) in Hardee and De Soto Counties, Florida[1]

By Lucy McCartan,[2] L.N. Plummer,[2] J.W. Hosterman,[2,3] E. Busenberg,[2] E.J. Dwornik,[2] A.D. Duerr,[4] R.L. Miller,[4] *and* J.L. Kiesler[4]

INTRODUCTION

Celestine ($SrSO_4$) has been identified in cuttings from two drill holes (Newbern Grove and Dees) in Hardee and De Soto Counties, south-central Florida (fig. 23.1), where it is found in small quantities in the Hawthorn Formation (Miocene), the Suwannee Limestone (Oligocene), and the Avon Park Formation (Eocene). Celestine,[5] a member of the barite group, is the principal ore of strontium. It is a relatively rare mineral that is found in some limestones, sandstones, and evaporite deposits and also is associated with volcanic rocks and metallic veins (Ford, 1932). Celestine also has been reported by Cook and others (1985) from the Avon Park Formation in the TR 18–2 drill hole in Hernando County, Fla. (fig. 23.1). Thermodynamic speciation calculations indicate that water in the Upper Floridan aquifer is in equilibrium with the celestine it contains in the Newbern Grove and Dees drill holes. This equilibrium suggests that celestine is widely distributed; so far, we have found it in only a few samples. The presence of celestine in substantial quantities would indicate unusual chemical conditions after and possibly during deposition of the host sediment.

This preliminary report describes the occurrence of celestine and ground-water characteristics in the study area. The paragenesis of celestine is discussed in two parts. The first part is a speculative account of the possible early diagenetic formation of celestine. The second part relates the modern distribution of celestine to hydrochemical facies. An understanding of the origin of celestine will aid in prediction of the evolution of water quality in south-central Florida.

Acknowledgments

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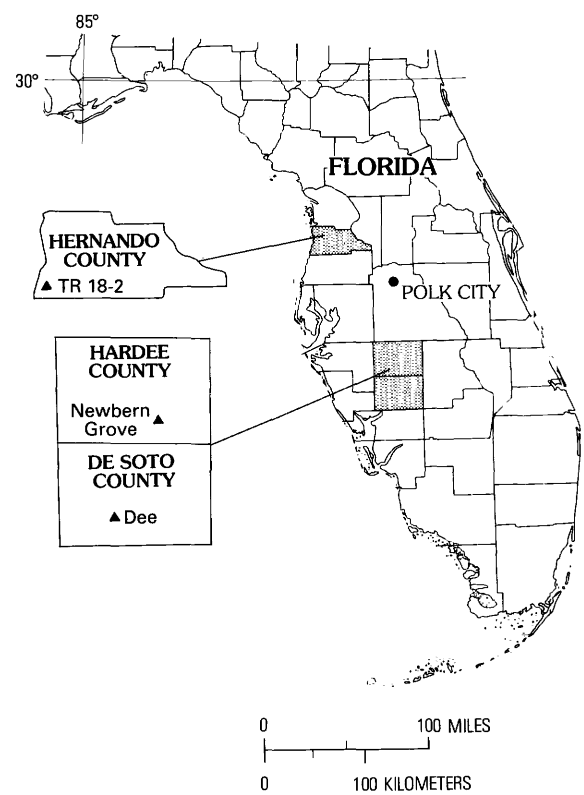

Figure 23.1. Locations of three Florida drill holes in which celestine has been found. Dees well and Newbern Grove well were drilled by the U.S. Geological Survey and are discussed in this report; TR 18–2 was described by Cook and others (1985).

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[5]The name celestine (Fleischer, 1987) is currently used by the U.S. Geological Survey.

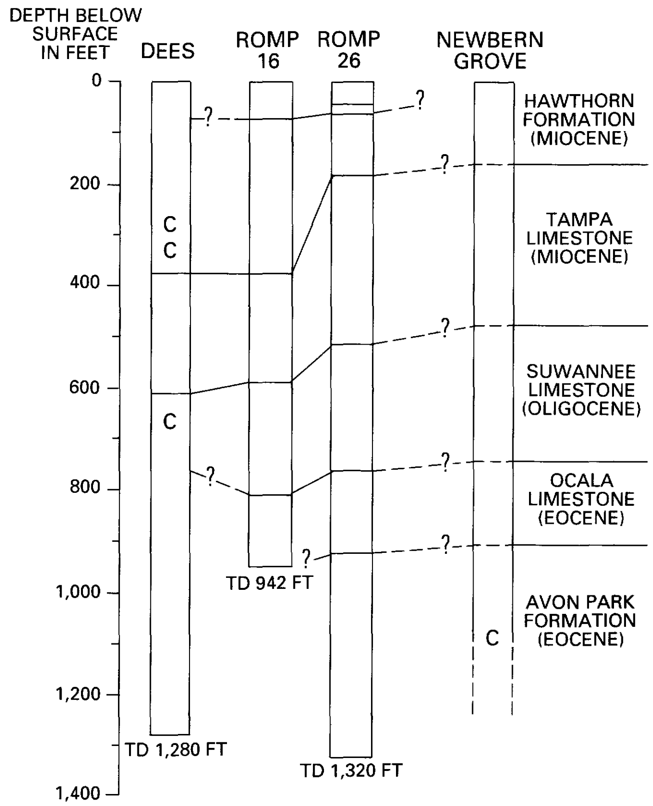

Figure 23.2. Stratigraphic columns showing depths at which celestine (C) was found in the Dees and Newbern Grove wells. Unpublished logs for ROMP wells 16 and 26 are from Thomas Scott and the files of the Florida Geological Survey, Tallahassee, Fla., and the U.S. Geological Survey, Tampa, Fla. Correlations are tentative. TD, total depth of well.

sample that was chemically analyzed. We appreciate reviews by Ruth Deike and Sam Altschuler, both of the USGS.

STRATIGRAPHY AND MINERALOGY

Celestine occurs as fine sand-sized grains in limestones, dolomites, and phosphatic sands in south-central Florida. Because samples that contain celestine were collected from several depths in the Dees and Newbern Grove drill holes (fig. 23.2), we know the general lithologic association, but the exact context of the sulfate cannot be determined without thin sections. The specific gravity (SG) of celestine is 3.9 and exceeds the specific gravities of calcite (2.7) and dolomite (2.9). Thus, celestine may lag behind both calcite ($CaCO_3$) and dolomite ($CaMg(CO_3)_2$) as cuttings move up the borehole during drilling, and samples are likely to contain cuttings of different densities from strata many feet apart.

Anhydrite ($CaSO_4$) and gypsum ($CaSO_4 \cdot 2H_2O$) were looked for but not found in the cuttings.[6] The major occurrence of sulfate minerals in south-central Florida is in the lower confining unit of the Floridan aquifer system, so-named because of its impermeability to ground water and its low porosity. The top of the lower confining unit in this area is more than 2,000 ft below the lowest celestine sample (Miller, 1986).

Celestine was discovered by optical microscopy in a bromoform (SG=2.85) concentrate of the fine sand fraction of disaggregated cuttings from a depth of 1,083 ft in the Newbern Grove well. It makes up about 99 volume percent of the heavy concentrate, about 50 volume percent of the fine sand fraction, and about 0.25 volume percent of the whole sample. Celestine was identified by comparison of peaks on a CuK_{α_1} X-ray diffraction trace (fig. 23.3) with those on a published standard celestine X-ray diffraction pattern (Joint Committee on Powder Diffraction Standards, 1986). The chemical composition was confirmed by X-ray dispersion on a scanning electron microscope (SEM). Celestine grains on the SEM appear as irregular, narrow, honeycombed plates, which typically have one or more smooth surfaces (fig. 23.4). Under the optical microscope, celestine grains are short, colorless, irregular prisms that have many voids or inclusions. The indices of refraction are about 1.62, the optic angle (2V) is fairly high, extinction is parallel, and birefringence is first-order white and gray. In at least one sample, collected from a depth of 675 ft in the Dees well, celestine is partly replaced by calcite.

CELESTINE AND GROUND-WATER CHEMISTRY

Water in the Upper Floridan aquifer in Hardee and De Soto Counties has anomalously high concentrations of Sr^{2+} (fig. 23.5, table 23.1) and SO_4^{2-}. The median concentration of Sr^{2+} in U.S. drinking water is 0.11 mg/L (Hem, 1985). The highest values reported in U.S. wells are 52 mg/L, in Wisconsin (Skougstad and Horr, 1963), and 57 mg/L, in Port Charlotte, Fla. (Steinkampf, 1982; see also Black and Brown, 1951). The maximum value reported in this study is 43 mg/L (table 23.1). The Sr^{2+} concentration increases rapidly from the regional potentiometric high at Polk City along a southwestward ground-water flow path (Back and Hanshaw, 1970) that includes wells at Zolfo Springs, ROMP 26, and ROMP 10 (figs. 23.6 and 23.7). Along the flow path, Sr^{2+} concentration peaks at the Zolfo Springs well at 31.0 mg/L and then decreases along the flow path toward the southwest, whereas SO_4^{2-} concentration continues to increase (fig. 23.7). High concentrations of Sr^{2+} also are present southeast of the flow path in De Soto County. The Sr^{2+} concentration shown for ROMP 10 in figure 23.5 is 34.0 mg/L; in figure 23.7, it is 19.0 mg/L. The two

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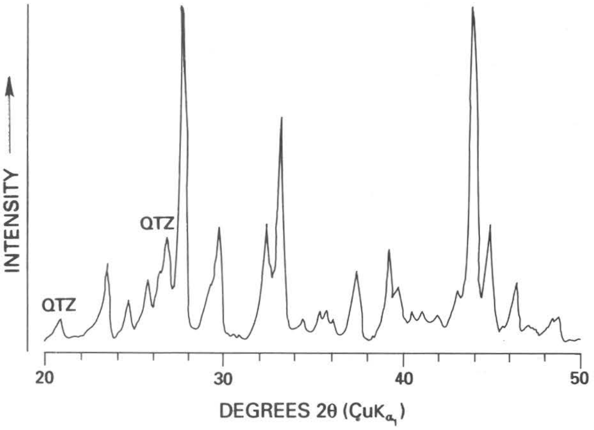
[6]Barite ($BaSO_4$), the major component of some drilling fluids, was also looked for but not found. The primary drilling fluid used in the two drill holes was water.



D-spacings (in degrees 2 theta and angstroms) and relative peak intensities (I/I_o) for celestine from a depth of 1,083 ft in the Newbern Grove well, Florida, and for standard celestine

[Data for standard celestine are from the Joint Committee on Powder Diffraction Standards (1986, card 5–593). Other reflections (angstroms/intensity) in the range from 2° to 50° 2θ listed in the standard file that did not appear on the trace of the Florida sample: 4.23/11, 3.29/98 (masked by the quartz peak in the Florida sample), 2.73/63, 2.39/7, 2.38/17, 2.16/7, 2.14/25, 2.00/48, 1.95/15. In addition, the card states that there are "10 other reflections" that are not listed. Dot leaders mean no reflection reported]

Newbern Grove celestine			Standard celestine		
D-spacing		I/I_o	D-spacing		I/I_o
°2θ	Å		°2θ	Å	
16.3	5.4	2	...	...	...
20.9*	4.3*	6*	...	...	...
23.4	3.8	22	23.6	3.77	35
24.6	3.6	11	24.9	3.57	2
25.7	3.5	11	...	...	...
26.4	3.4	3	26.0	3.43	30
26.7*	3.3*	19*	...	...	...
27.7	3.2	99	28.1	3.18	59
29.8	3.0	32	30.1	2.97	100
32.5	2.8	22	...	...	...
33.3	2.7	67	33.5	2.67	49
34.4	2.6	5	34.7	2.58	6
35.4	2.54	9	...	...	...
36.1	2.51	7	...	...	...
37.4	2.49	20	...	...	...
39.3	2.29	27	...	...	...
40.6	2.22	9	40.0	2.25	18
41.2	2.19	9	...	...	...
42.0	2.20	8	40.9	2.21	5
44.0	2.06	100	44.3	2.04	55
44.9	2.02	30	44.4	2.04	57
46.4	2.00	17	45.2	2.01	40
48.3	1.88	7	...	...	...
48.9	1.86	7	49.1	1.86	7

*Peak for internal quartz standard.

Figure 23.3. X-ray diffractogram of celestine in cuttings from a depth of 1,083 ft, Newbern Grove well, Hardee County, Fla. Celestine was crushed and attached to glass slide with amyl acetate. Quartz (26.7° and 20.9° 2Θ) is an internal standard. Table lists peaks from Newbern Grove celestine and standard celestine.

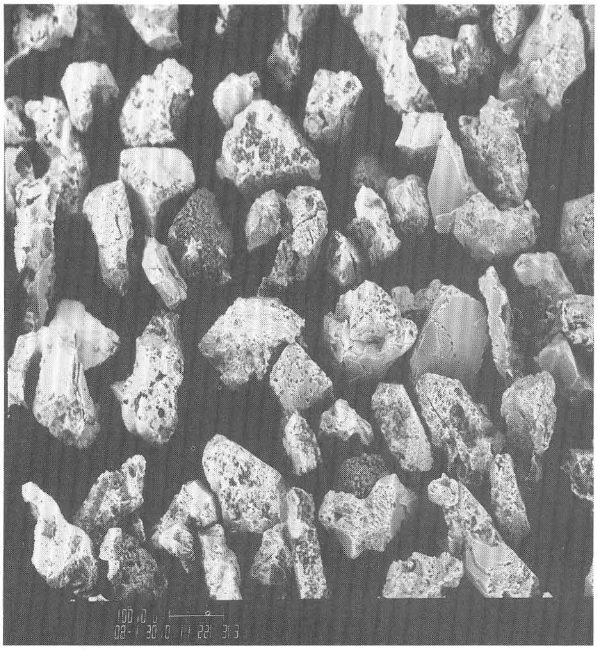


Figure 23.4. Scanning electron micrograph of celestine grains in cuttings from a depth of 1,083 ft in the Newbern Grove well, Hardee County, Fla. Grains are approximately 0.1 mm across.

values were obtained at different times. The lower measurement was made after permitting the well to flow for several hours and is considered the more accurate of the two.

Thermodynamic speciation calculations made by using the computer program WATEQF (Plummer and others, 1976) indicate that ground water in the Upper Floridan aquifer is initially undersaturated with celestine and gypsum at the Polk City potentiometric high (fig. 23.7). As the water moves southwestward along the flow path, the saturation indices of both minerals increase; apparently, the water reaches equilibrium with celestine near Zolfo Springs and beyond but never reaches equilibrium with gypsum along the flow path (fig. 23.7). These calculations indicate that celestine initially dissolves along the flow path. Dissolution of Sr-bearing gypsum further increases the dissolved Sr content. As the ground water reaches saturation with celestine near Zolfo Springs, further dissolution of gypsum causes precipitation of celestine and causes the observed decrease in Sr^{2+} concentration (fig. 23.7). The data suggest that the present ground-water diagenetic environment is dissolving celestine in some places and redepositing it at points further down the hydrologic gradient in the Upper Floridan aquifer system. Aragonite is not present in the section and is not providing Sr^{2+} in the modern ground-water system in the study area. Recrystallized shells that were originally aragonite indicate that aragonite may have been an early diagenetic source of Sr^{2+}.

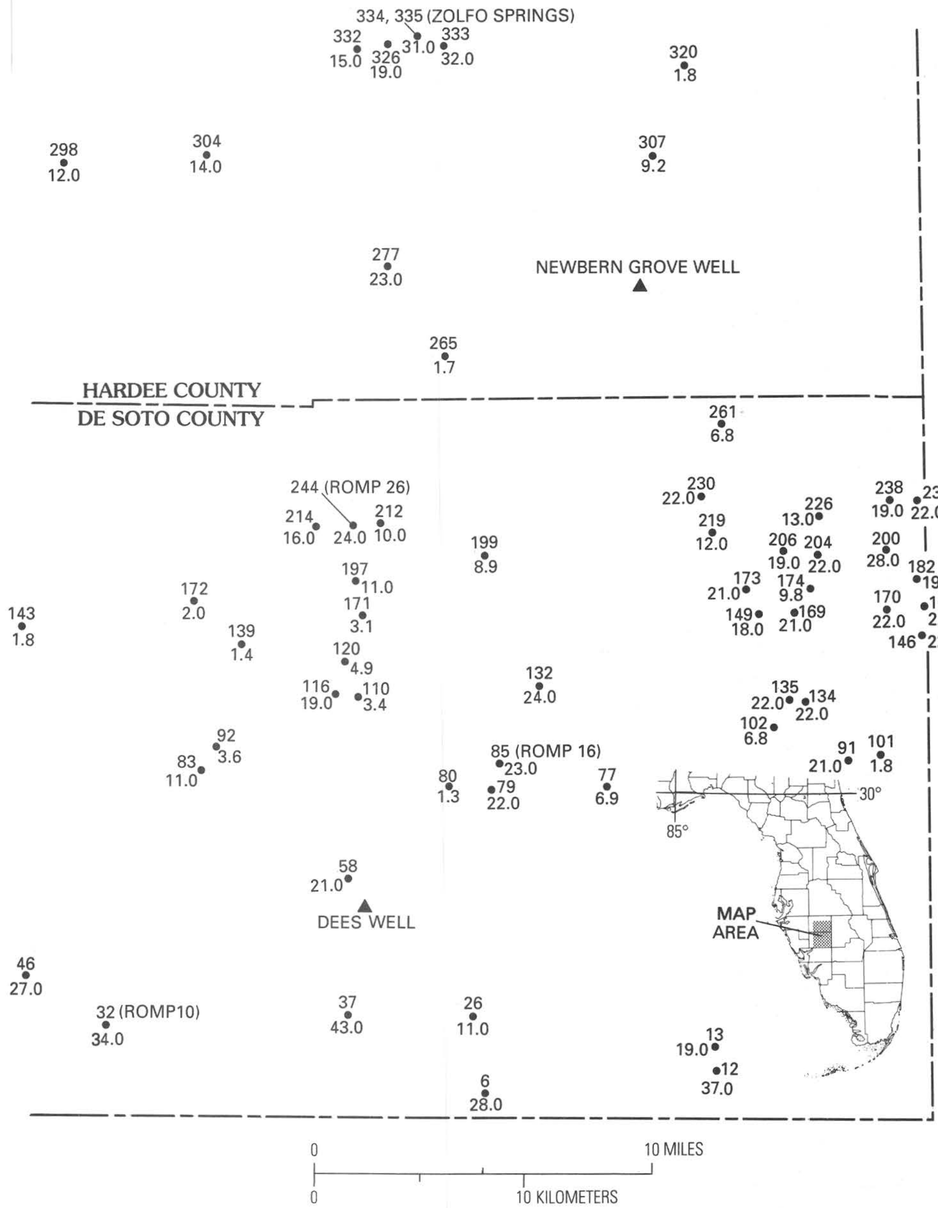


Figure 23.5. Concentration of Sr^{2+} in ground water in Hardee and De Soto Counties, Fla. Decimal numbers indicate maximum concentration of Sr^{2+}, in milligrams per liter (mg/L), for water in each well; other numbers refer to well localities (table 23.1).

DISCUSSION

The present distribution of celestine in south-central Florida, as inferred from the mineral samples and the presence of high Sr^{2+} concentrations in ground water of the Upper Floridan aquifer, is a result of early and late diagenetic processes. In the early processes, the major ions that presently control the ground-water chemistry were trapped in solid phases (minerals). Subsequently, as freshwater was added in the recharge area and sea level changed, the distribution of celestine changed from an early diagenetic pattern to the present, late diagenetic pattern.

Early Diagenesis

The ions most likely involved in the generation of celestine during or soon after deposition of the host sediment are Ca^{2+}, Mg^{2+}, Sr^{2+}, SO_4^{2-}, and HCO_3^-. Clues to

Table 23.1. Concentration of strontium in ground water in Hardee and De Soto Counties, Fla.

[Data are from the files of the U.S. Geological Survey, Tampa, Fla. Only values measured after 1972 were used; many older measurements are inaccurate. NGVD, National Geodetic Vertical Datum]

Locality number (fig. 23.5)	Altitude of land surface (feet above NGVD of 1929)	Total depth of well (feet)	Concentration of dissolved strontium, Sr^{2+} (milligrams per liter)
De Soto County:			
6	36	900	28.0
12	42	2,160	37.0
13	42	1,210	19.0
26	47	460	11.0
32 (ROMP 10)	35	411	19.0, 34.0
37	47	1,189	43.0, 34.0
46	34	1,411	27.0
58	40	418	21.0
77	61	396	6.9
79	64	1,500	22.0
80	62	335	1.3
83	54	384	11.0
85 (ROMP 16)	54	942	23.0
91	81	1,478	21.0
92	73	335	3.6
101	?	291	1.8
102	?	?	6.8
110	65	353	3.4
116	?	?	1.2, 19.0
120	?	350	4.9
132	68	1,412	24.0
134	82	1,508	22.0
135	80	1,461	22.0
139	55	250	1.4
143	49	430	1.8
146	87	1,525	22.0
149	86	1,492	18.0
162	89	1,018	22.0
169	80	1,350	19.0, 20.0, 21.0
170	90	300	22.0
171	62	327	3.1
De Soto County:			
172	59	320	2.0
173	86	294	11.0, 8.8, 7.7, 10.0, 21.0
174	89	300	5.6, 9.8
182	91	1,426	19.0
197	20	208	11.0
199	80	1,300	8.9
200	90	1,443	28.0
204	91	1,271	19.0, 19.0, 22.0
206	89	1,421	19.0
212	34	257	10.0
214	62	893	16.0
219	90	895	12.0
226	89	1,343	13.0
230	90	765	22.0
237	93	1,361	22.0
238	91	1,505	19.0
244 (ROMP 26)	75	1,320	24.0
261	94	284	3.9, 6.8
Hardee County:			
265	75	1,345	1.7
277	54	1,196	23.0
298	86	1,280	12.0
304	78	350	14.0
307	75	343	9.2
320	83	140	1.8
326	55	326	19.0
332	90	321	15.0
333	65	1,002	3.5, 32.0
334 (Zolfo Springs)	65	1,002	31.0, 30.0
335 (Zolfo Springs)	65	933	28.0

the early diagenetic processes may be found by examining the difference between the concentration of these ions in seawater of normal salinity (35 parts per thousand) and their proportions in minerals in accumulating sediments, where the amount of calcite ($CaCO_3$) is equal to the amount of dolomite ($CaMg(CO_3)_2$), and both are more abundant than gypsum ($CaSO_4 \cdot 2H_2O$). All three minerals are much more abundant than celestine ($SrSO_4$). The differences are due in part to intrabasinal biologic and climatic controls and to the contribution of magnesium by extrabasinal sources such as volcanic ash.

Biogenic $CaCO_3$ in the form of calcite and aragonite shells accounts for most or all of the carbonate as deposited originally. A small amount of Sr^{2+}, typically less than 1 mole percent (Plummer and Busenberg, 1987), commonly enters the aragonite structure during the precipitation of shell material. Thus, the likely starting materials are $Ca_{1-x}Sr_xCO_3$ (where x=0 to 0.03) and seawater. The presence of algal laminae and macroscopic Foraminifera at several levels in the area indicates episodes of hypersalinity (Applin and Applin, 1944). As brine from sabkhas adjacent to the submerged carbonate platform moves downward through the accumulating calcium carbonates, magnesium reacts with $CaCO_3$ to form dolomite (Land, 1982). During intervening periods of lower salinity, Sr^{2+} can be released as aragonite recrystallizes to low-Mg calcite (Weiss and Wilkinson, 1988). The paucity of aragonite in pre-late Pleistocene shell deposits suggests that most Sr^{2+} is released from aragonite soon after deposition (Harris and Matthews, 1968; Halley and Harris, 1979). In the hot, arid

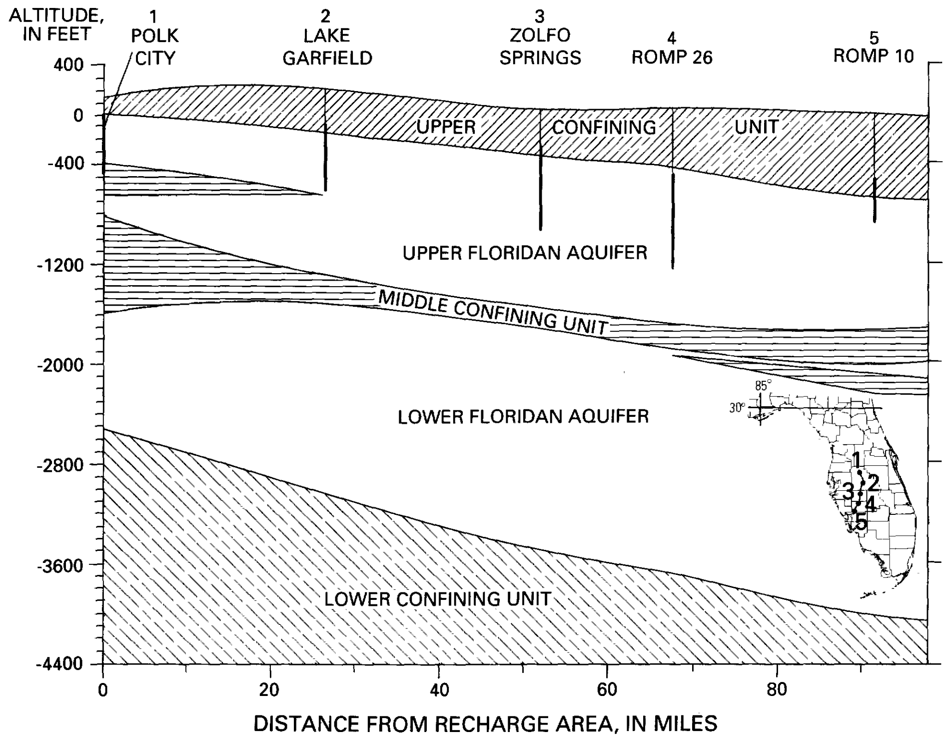


Figure 23.6. Geologic cross section southward from Polk City, Fla., along groundwater flow path. North is at the left. Ionic concentration data from Miller (1986); concentration increases are shown by thickening of lines representing wells. Upper and middle confining units are discontinuous and leaky in places, including Polk City. Polk City is near the center of the potentiometric high that marks the recharge area.

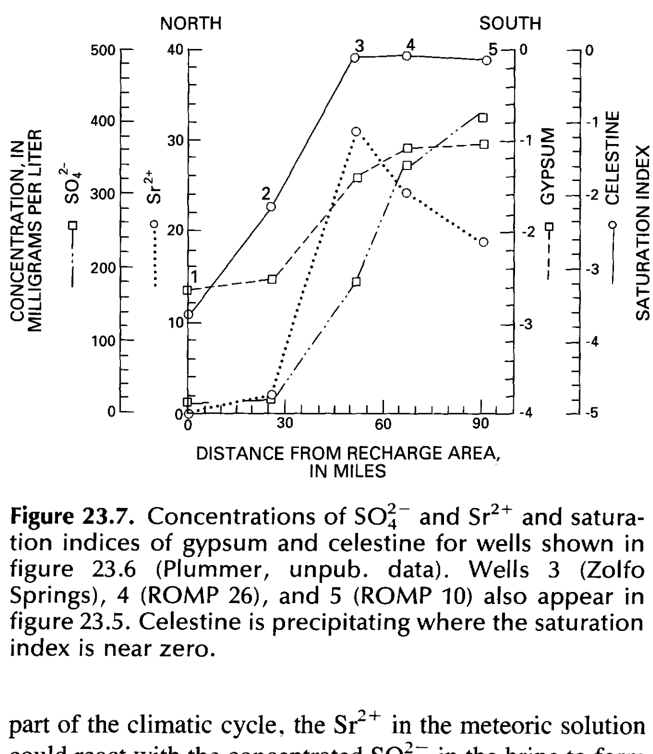


Figure 23.7. Concentrations of SO_4^{2-} and Sr^{2+} and saturation indices of gypsum and celestine for wells shown in figure 23.6 (Plummer, unpub. data). Wells 3 (Zolfo Springs), 4 (ROMP 26), and 5 (ROMP 10) also appear in figure 23.5. Celestine is precipitating where the saturation index is near zero.

part of the climatic cycle, the Sr^{2+} in the meteoric solution could react with the concentrated SO_4^{2-} in the brine to form celestine if Ca^{2+} were not also available.

Under most circumstances, Ca^{2+} is, in fact, abundantly available. Seawater and brine originally have five times as much Mg^{2+} as Ca^{2+} (Hood and Pytkowicz, 1974), but as Mg^{2+} is used up in early dolomitization, the brine becomes enriched in Ca^{2+}. This enrichment promotes the precipitation of gypsum rather than celestine. However, if Mg^{2+} is introduced to balance the excess Ca^{2+}, then Ca^{2+} and Mg^{2+} can combine with HCO_3^- to form interstitial dolomite. Actually, dolomite can be precipitated if the Mg^{2+}/Ca^{2+} ratio is greater than about 3 (Pierre and others, 1984), but in the Florida Keys, dolomite is precipitated where interstitial brine has a Mg^{2+}/Ca^{2+} ratio of 40/1 (Shinn, 1972). Dolomitization would remove Ca^{2+} and enhance the possibility of $SrSO_4$ precipitation.

The richest sources of Mg^{2+} are volcanic rocks. Rivers draining basic volcanic rocks in Australia supply abundant Mg^{2+} to the Coorong Lagoon (table 23.2), where dolomite is precipitated (von der Borch, 1976). Although we cannot find evidence linking south-central Florida rivers with volcanic source rocks in the Cenozoic, some mafic ash from Eocene through Miocene volcanic events in the Caribbean (Case and Holcombe, 1980) and western North America (Gilluly, 1965) probably was spread by wind and water over the carbonate banks and sabkhas of the Florida peninsula during the Eocene through Miocene. Mg^{2+} would

Table 23.2. Some chemical and mineralogical characteristics of south-central Florida and three other gypsum-bearing localities

[Y, abundant; y, present but not abundant; N, not present]

	South-central Florida	Coorong Lagoon, Lake Alexandrina, Australia	Ojo de Liebre Lagoon, Baja California, Mexico	Southern U.S. Atlantic Coastal Plain
Setting	Subsurface carbonates; Eocene–Miocene.	Alkaline lake; carbonate substrate; Holocene.	Shoreline bay; noncarbonate substrate; Holocene.	Subsurface mixed carbonates and phosphate-siliciclastic rocks; Eocene–Miocene.
References	This report; Miller, 1986; Thayer and Miller, 1984.	von der Borch, 1976.	Pierre and others, 1984.	Gohn, 1988; Weaver and Beck, 1977.
Celestine	Y	N	N	N
Gypsum	Y[1]	Y	Y	y
Calcite	Y	y	y	Y
Aragonite	y[2]	Y	N	y[2]
Dolomite	Y	Y	Y	Y
Halite	N	N	Y	N
Sr^{2+} highly concentrated interstitially.	Y	N	Y,[3] N	N
Main Sr^{2+} source	Brine, aragonite[4]	Holocene aragonite substrate.	Pleistocene aragonite in highlands.	Brine, aragonite.
Mg-clays	Y	N	N	Y
Main Mg^{2+} source	Seawater	Volcanic source rocks.	Seawater	Seawater.
Inferred secondary source of Mg^{2+}.	Volcanic ash	None	None	Volcanic ash.

[1]Gypsum is mainly 2,000 ft or more below lowest celestine sample.
[2]In pre-late Pleistocene deposits, most aragonite has recrystallized to calcite.
[3]Wide variations laterally.
[4]Aragonite is inferred to have been the original mineralogy of many mollusk shells that are now calcite or are represented by empty molds.

be released easily from the highly unstable glass in the ash as it reacted with the brine. Both dolomite and celestine could then precipitate from Mg-rich brine. Other evidence that may be explained by the interaction of Mg-rich ash and brine is the presence of the Mg-rich clays palygorskite ($(MgAl)_5(SiAl)_8O_{22}(OH)_2 \cdot 8H_2O$) and sepiolite ($Mg_4Si_6O_{15}(OH)_2 \cdot 6H_2O$) in cuttings from the Dees well, in cores from nearby wells (McCartan, Trommer, Owens, and Hawkins, article 24, this volume), and at many other localities in Florida, Georgia, and South Carolina (Weaver and Beck, 1977) (table 23.2).

Late Diagenesis

Petrologic studies that reveal multiple episodes of dolomitization in Eocene and Paleocene rocks in central Florida (Thayer and Miller, 1984; see also Randazzo and Hickey, 1978) are applicable to De Soto and Hardee County rocks. These complexities are due to processes such as climatic changes (brine produced in arid times followed by freshwater flushing during humid times), rises in sea level (flushing of brine or freshwater by normal seawater), and changes accompanying rising temperature and pressure due to burial.

Another indicator of diagenesis is the difference between the present chemical composition of subsurface minerals and the chemical composition of experimentally grown minerals. Our preliminary analysis of a gypsum sample from the middle confining unit in a well in St. Petersburg, about 60 mi west of Lake Garfield (fig. 23.6), reveals 0.08 mole percent strontium. Kushnir's (1980) gypsum grown at 30 °C from seawater concentrated to three times normal salinity contains 0.18 mole percent strontium. The strontium deficit in the St. Petersburg gypsum may be a result of recrystallization of gypsum in equilibrium with less saline water (Ichikuni and Musha, 1978) or primary precipitation in a brine relatively depleted in Sr^{2+}.

The distribution of hydrochemical facies reflects modern diagenetic processes acting on previously deposited minerals. Using the hydrochemical data of Back and Hanshaw (1970), Plummer (1977) and Plummer and others (1983) showed that the predominant diagenetic reaction in the study area is dedolomitization, that is, the net dissolution of dolomite and the precipitation of calcite driven by gypsum dissolution. Accompanying the dedolomitization reaction are many other diagenetic reactions that occur to a lesser extent, including sulfate reduction (Plummer and others, 1983) and dissolution and reprecipitation of celestine (this study). Currently, celestine is being dissolved upgradient and reprecipitated at points further down the hydrologic gradient to the southwest as $CaSO_4$ continues to dissolve. If sea level does not rise, all celestine eventually will be flushed seaward.

Kushnir (1986) considered several models for the origin of celestine and the evolution of the dissolved Sr content of ground water in contact with $CaSO_4$-bearing rocks. In one model (mechanism A), water rich in NaCl and Sr^{2+} (several hundred milligrams per liter) precipitates $SrSO_4$ upon contact with $CaSO_4$-bearing rocks. This reaction probably occurs deep in sedimentary basins but is currently not occurring in the Upper Floridan aquifer. In Kushnir's (1986) second model (mechanism B), water having low NaCl content and low Sr^{2+} content (approximately 1 mg/L) flows into a zone of Sr-bearing $CaSO_4$. Release of Sr^{2+} by dissolution of $CaSO_4$ causes a rapid rise in dissolved Sr content, and if celestine saturation is reached prior to $CaSO_4$ saturation, further dissolution or recrystallization of $CaSO_4$ causes precipitation of celestine and a subsequent lowering of dissolved Sr content. Our observations in Hardee and De Soto Counties, Fla., confirm that Kushnir's mechanism B is operating downgradient; however, the presence of celestine in the rocks upgradient of the $CaSO_4$-bearing zone must be explained by another mechanism. Relatively dilute ground water attains high dissolved Sr content (fig. 23.5) and approaches celestine saturation before reaching the $CaSO_4$ zone. Therefore, dissolution or recrystallization of $CaSO_4$ is not the mechanism whereby Sr^{2+} concentration is increased in the upgradient part of the flow path. Other possible mechanisms for this part of the flow path include dissolution or recrystallization of aragonite (Kinsman, 1969) and dissolution of preexisting celestine. Along the flow path of figure 23.6, the water chemistry suggests that the $CaSO_4$ zone begins between Lake Garfield and Zolfo Springs (fig. 23.7). The precipitation of celestine, which begins when ground water enters the $CaSO_4$ zone, causes a decrease in dissolved Sr content downgradient of Zolfo Springs (fig. 23.7).

FACTORS AFFECTING CELESTINE DEPOSITION AND NONDEPOSITION IN OTHER AREAS

Celestine is reported in notable amounts from one other Coastal Plain locality, in central Texas (Pearson and Rettman, 1976; Deike, 1991). It is in the Edwards Limestone (Cretaceous), where it coexists with calcite, dolomite, gypsum, and fluorite. Thin sections of the Edwards Limestone reveal that celestine fills bioclast molds in units composed of secondary dolomite and that it also occurs as late diagenetic precipitates along fractures. The celestine appears to have been introduced, or at least reprecipitated, well after lithification of the host sediment. In this respect, it is similar to the Avon Park Formation occurrence in Hernando County, Fla. (Cook and others, 1985). Similarly, Sr^{2+}-rich ground water in Ohio is associated with celestine-bearing evaporitic rocks, and adjacent shales do not have high Sr^{2+} levels (Feulner and Hubble, 1960). However, ground water approaching celestine saturation (maximum Sr^{2+} concentration=15 mg/L and SO_4^{2-} concentration typically greater than 1,300 mg/L) is present in the Madison Limestone (Mississippian) in Montana, South Dakota, and Wyoming, but celestine has not been found in cores from the area (Busby and others, 1991). Because of the apparent lack of celestine in a variety of other settings where its components are abundant (table 23.2), the presence of celestine in south-central Florida cannot be explained readily by reference simply to burial, climatic or sea-level changes, and the release of Sr^{2+} during recrystallization of aragonite and gypsum. Further petrologic and geochemical studies may yield new evidence of the special conditions necessary for the formation of celestine.

CONCLUSIONS

The distribution of celestine in south-central Florida is a result of early concentration and trapping of Sr^{2+}, SO_4^{2-}, Mg^{2+}, HCO_3^-, and Ca^{2+} in solid phases (minerals) in carbonate and evaporite deposits and the subsequent dissolution and reprecipitation of celestine by ground-water processes.

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24. Mineralogy of a Sand-Filled Sinkhole Complex in Pinellas County, Florida[1]

By Lucy McCartan,[2] J.T. Trommer,[3] J.P. Owens,[2] *and* R.M. Hawkins[2]

INTRODUCTION

A sinkhole complex in Pinellas County, Fla., has formed in the Tampa Limestone of Miocene age (figs. 24.1 and 24.2). It is filled mainly with quartz sand from the overlying unit, which is presumably of Pleistocene age. As much as 15 ft of leached limestone separates the surficial sand from the less weathered limestone below. Thirty-four holes 12–74 ft deep were drilled in 1987–88 to establish the stratigraphy in the area and to install ground-water monitoring wells. Six of these have been studied in detail. In this report, we describe split-spoon and auger samples from these drill holes and provide a preliminary interpretation of their mineralogical significance.

Acknowledgments

This study was conducted by the U.S. Geological Survey (USGS) in cooperation with the Pinellas County Sewer System, which provided some support for Trommer. Reviews by Craig Hutchinson and David Soller (both of the USGS) led to improvements in the report.

METHODS

Split-spoon samples are taken by pushing a hollow steel or brass pipe with a plastic liner ahead of a drill bit attached to auger stems. Therefore, split-spoon samples generally are more representative of the geologic unit from which they are taken than the material on the auger flights. For this reason, the two types of samples are distinguished in this report. Drill holes WP, FL3, T3, and T2 were sampled by the split-spoon method; only auger samples are available from 2ndFL and Surf2.

Subsamples were selected from the plastic split-spoon liners and bags of auger samples for separation and mineralogical analyses. Ten to 20 cm^3 of coarse-grained material (sand) and about 1 cm^3 of fine-grained material (silt and clay) were used for mineralogical analyses. Clay-sized minerals were extracted from all samples by mixing them with deionized water. The muddy water was decanted and centrifuged to permit recovery of the fraction containing particles finer than 2 µm. Clay was then sedimented on glass slides, air dried, and X-rayed with CuK_{α_1} radiation. Minerals in the clay fraction were identified by comparison of peaks between 2.5° and 12.5° 2Θ with those in the standard data book (Joint Committee on Powder Diffraction Standards, 1974, 1981). Untreated samples yielding peaks between 2.5° and 8.8° 2Θ were vapor glycolated overnight and X-rayed again. Glycolation causes the main peak for illite/smectite mixed-layer clay to shift from 6° to about 5° 2Θ; this shift enables us to discriminate between this mineral and dioctahedral vermiculite, whose main peak remains at about 6°.

Sand-sized material was wet sieved to remove clay and to separate the fine sand fraction (80–200 mesh; 0.18–0.07 mm); the sand was then dried at 110 °C. Light and heavy fine sand fractions were then separated in bromoform (specific gravity of 2.85). The mineral grains were identified in index oil mounts examined under a petrographic microscope. Nonopaque heavy minerals (figs. 24.3 and 24.4, table 24.1) were characterized as nonresistant (hornblende, garnet, monazite, and sphene), intermediate (staurolite, sillimanite, kyanite, and andalusite), and resistant (zircon, tourmaline, rutile, and spinel) on the basis of their observed resistance to chemical dissolution and physical abrasion (Owens and others, 1983). Heavy minerals were estimated to compose less than 5 percent of all the samples. Phosphate is abundant in some samples from 2ndFL and WP but was not quantified.

The proportion of carbonate minerals was determined by repeated treatment of selected whole subsamples with 10 percent hydrochloric acid. The weight of the dried sample after acid treatment was subtracted from the dry weight of the original sample.

STRATIGRAPHIC NOMENCLATURE

The Florida Geological Survey has suggested changes in the stratigraphic nomenclature for Miocene rocks in

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[1]Prepared in cooperation with the Pinellas County Sewer System.
[2]U.S. Geological Survey, Reston, VA 22092.
[3]U.S. Geological Survey, Tampa, FL 33634.

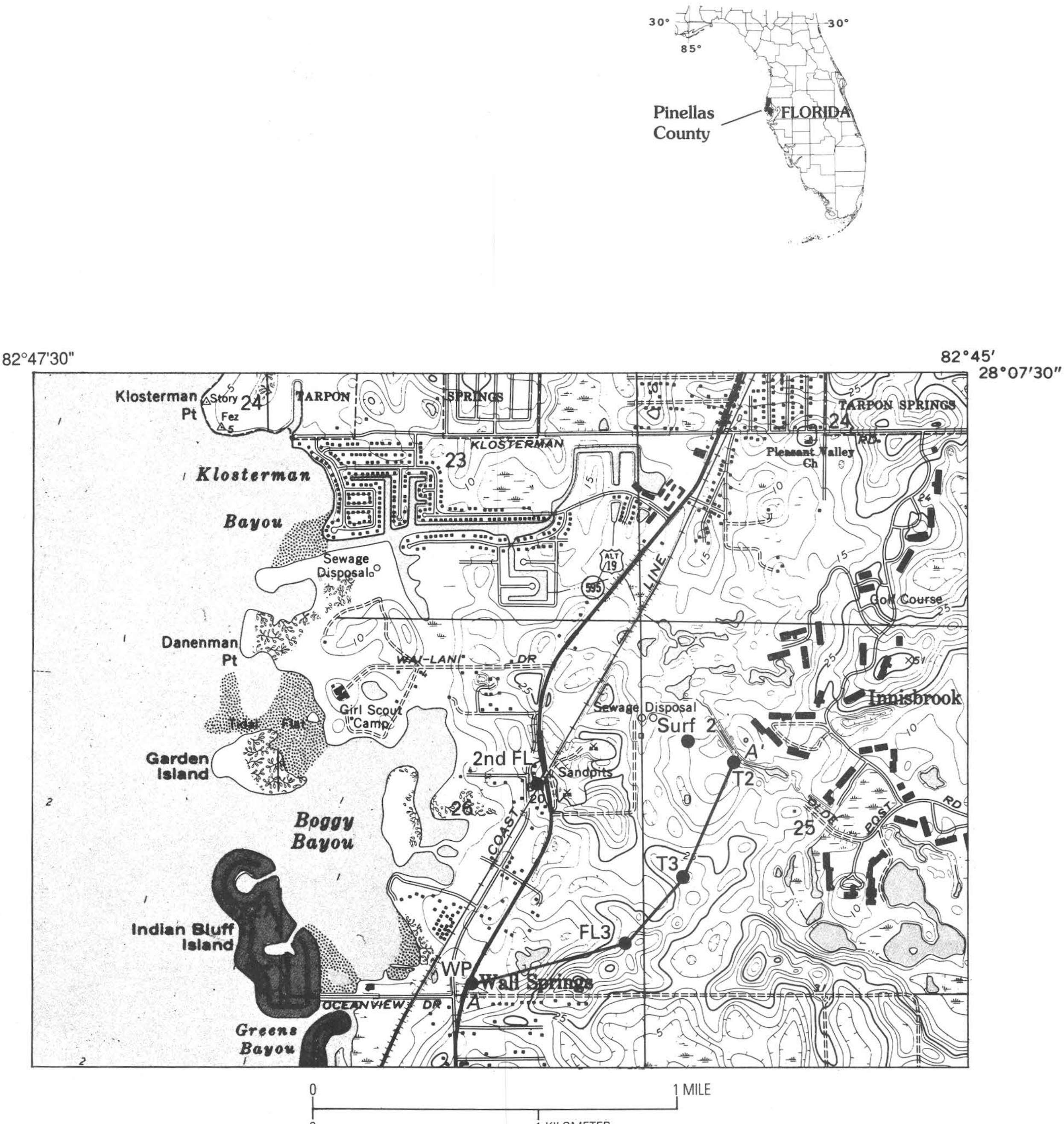

Figure 24.1. Map of part of a sinkhole complex in the study area, Pinellas County, Fla., showing topography (contour interval=5 ft), locations of drill holes from which split-spoon samples (WP, FL3, T3, and T2) and auger samples (2ndFL and Surf2) were taken, and line of stratigraphic section *A–A'* (fig. 24.6). Figure 24.2 depicts the same part of the Dunedin 7.5-minute quadrangle shown here.

southwestern Florida. These changes reflect Scott's (1986, 1988) correlation of cores and cuttings. The simpler scheme of Miller (1986) is used in this report. The two schemes are compared in figure 24.5.

MINERALOGY

The noncarbonate fraction of the Tampa Limestone is mainly quartz, but it contains as much as 20 percent

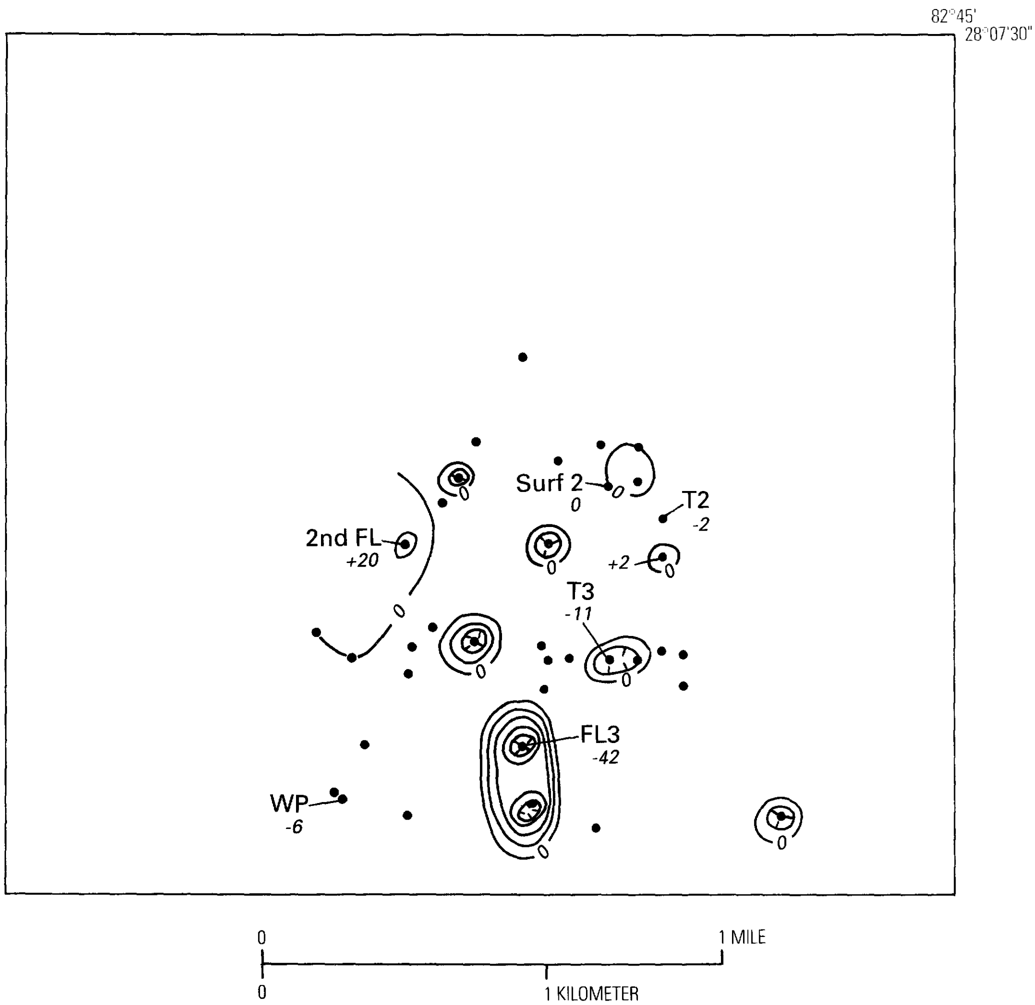

Figure 24.2. Structure contours of base of surficial sands in part of the northeast corner of the Dunedin, Fla., 7.5-minute quadrangle. Numbers are feet above or below sea level; contour interval is 10 ft. Tick marks on inner contour line indicate a depression. Other sinkholes (a few feet in diameter) were suggested by ground-penetrating radar surveys conducted in Pinellas County from 1986 to 1988 (J.T. Trommer, unpub. data). Unlabeled black dots are control drill holes (not discussed in this report) for which the depth to the base of surficial sand is known. At points not enclosed in contour lines, the surficial sand may be as much as 9 ft thick.

feldspar. A very small proportion of the total volume is clay minerals, and the predominant phase is illite/smectite. The most abundant heavy minerals are ilmenite, zircon, tourmaline, rutile, and garnet.

The mineralogy of the overlying sand is more mature than that of the Tampa Limestone. There is very little feldspar, and the resistant heavy minerals are more abundant in the sand than in the limestone. Quartz is the most abundant mineral in the clay-sized fraction of the surficial sand, and kaolinite is the next most abundant mineral. Illite, illite/smectite, and vermiculite are present in small amounts.

The intervening limestone residuum has some characteristics of both the fresher limestone below and the sand above. It has slightly less feldspar than the limestone but more than the sand. Similarly, the residuum has a slightly lower proportion of garnet and other fairly nonresistant heavy minerals than the limestone but has a larger proportion than the sand. The main clay-sized minerals in the residuum are illite/smectite in WP and T2, kaolinite in T3, and calcite in 2ndFL.

Fine Sand Fraction

The high proportion of staurolite, sillimanite, kyanite, and garnet in the heavy-mineral suite of the fine sand fraction (table 24.1) indicates that the terrigenous component of the Tampa Limestone and the overlying sand was derived originally from a metamorphic source, such as the

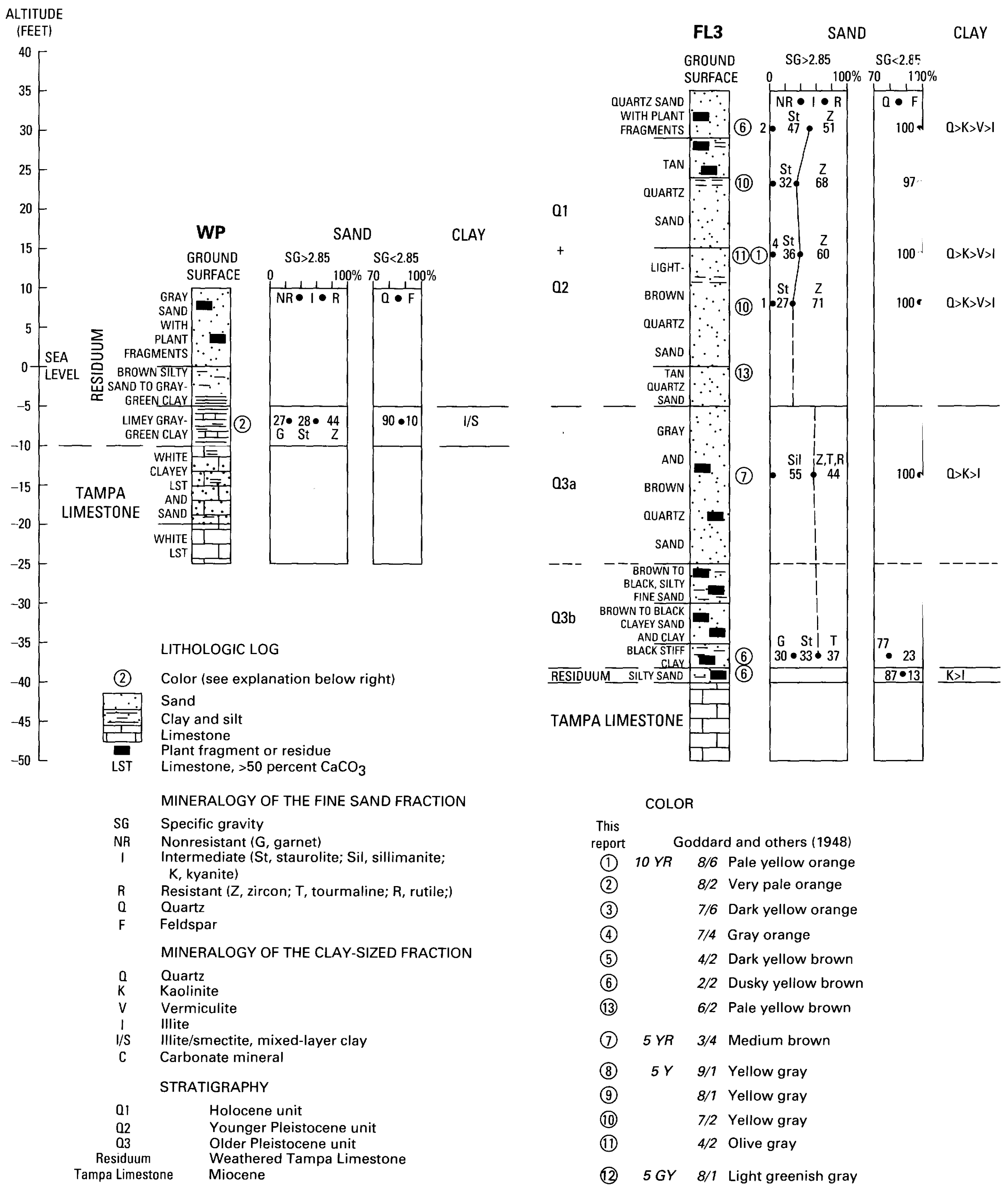

Figure 24.3. Lithology and mineralogy of drill holes WP and FL3, Pinellas County, Fla. Locations are shown in figure 24.1. Columns labeled sand and clay convey the dominant minerals present in fine sand (approximately 0.25–0.06 mm) and clay (clay is <2 μm) size categories. The nonopaque heavy sand minerals are further divided into nonresistant (to chemical weathering), intermediate, and resistant groups. The three figures in a line represent the total percentage of each heavy-mineral group within the nonopaque minerals. Minerals indicated near each number have the highest frequency percentages. Quartz makes up more than 90 percent of the fine sand fraction of the surficial unit, but feldspar is as much as 23 percent of the noncarbonate sand in residuum and fresh limestone samples. The relative amount of each mineral in the clay-sized fraction is shown in the right column by letter symbols. Colors to the left of the graphic lithology column were described in the field; colors designated by circled numbers to the right were noted at the time of the laboratory analysis.

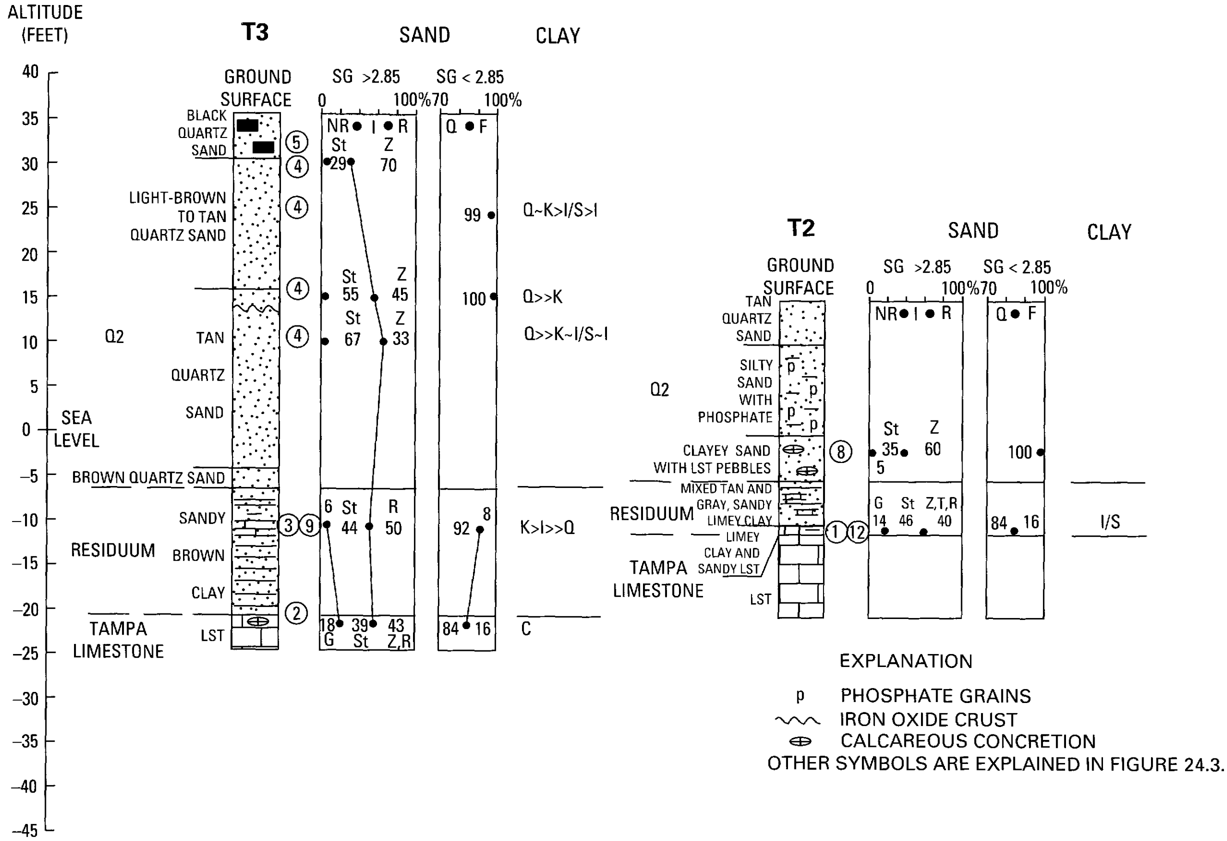

Figure 24.4. Lithology and mineralogy of drill holes T3 and T2, Pinellas County, Fla. Locations are shown in figure 24.1. Symbols are the same as those in figure 24.3.

Piedmont rocks of Georgia. Potassium feldspar, abundant only in the limestone and limestone residuum, indicates a granitic (potassium-rich) source. Together, the feldspar and heavy-mineral suite reflect a granitic-metamorphic terrane.

Ilmenite, the most abundant opaque mineral, typically is between a quarter and half of the total heavy-mineral fraction of the Pinellas County sediments. Ilmenite is found in most metamorphic, igneous, and sedimentary rocks, and so it is not useful as a provenance indicator. Similarly, quartz is common in most rocks of intermediate to felsic composition and contributes little to the discussion of provenance.

Phosphate minerals except monazite are included in the opaque heavy minerals column in table 24.1 although some grains are not opaque. Phosphate other than monazite in the surficial sand is reworked from either the Tampa Limestone or younger Miocene phosphatic units that are present east of the study area (Johnson, 1986; Scott, 1986). Monazite is a detrital mineral derived from the Piedmont of Georgia.

Resistant minerals are more abundant in the sands than in the limestone and residuum. This relative abundance suggests that the sand was derived mainly from the Tampa Limestone rather than directly from the Piedmont. During reworking, less resistant minerals were abraded and dissolved more rapidly than more resistant minerals. Another explanation, which seems much less likely, is that a sedimentary rock unit other than the Tampa Limestone was the source of the mineral suite that characterizes the surficial sands in Pinellas County.

Clay Fraction

Very little clay is present in the fresh limestone. Kaolinite is the most abundant clay mineral in the surficial sands and the residuum. Quartz, however, is the most abundant mineral (clay sized but not a clay mineral) in the clay-sized fraction of the surficial sands, and its origin is discussed below.

The presence of a minor amount of sepiolite in the 2ndFL well may reflect a high concentration of magnesium in the environment of deposition or in early postdepositional ground water (Weaver and Beck, 1977; Fisher, 1988). Judging from the vertical distribution of clay minerals in the drill holes, we infer that illite/smectite mixed-layer clay was the major detrital clay phase incorporated in the Tampa Limestone.

Table 24.1. Mineralogy of selected lithologic samples from the Pinellas County, Fla., sinkhole complex

[All values in frequency percent except those for calcite, which are in weight percent. Dot leaders (...), mineral not found; na, not analyzed. Heavy minerals have specific gravities greater than 2.85. Fine sand fraction: H, hornblende; G, garnet; M, monazite; Sph, sphene; St, staurolite; Sil, sillimanite; K, kyanite; A, andalusite; Z, zircon; T, tourmaline; R, rutile; Sp, spinel. Feldspar is all potassium feldspar. Clay-sized fraction: Q, quartz; V, dioctahedral vermiculite; K, kaolinite; I/S, illite/smectite mixed-layer clay; I, illite; C, calcite; H, halloysite; Sep, sepiolite]

Well	Unit	Depth below surface (feet)	Fine sand fraction																Clay-sized fraction (<2 μm)
			Nonopaque heavy minerals												Opaque heavy minerals	Light minerals			
			Nonresistant				Intermediate				Resistant								
			H	G	M	Sph	St	Sil	K	A	Z	T	R	Sp		Quartz	Feldspar	Calcite[1]	
Split-spoon samples																			
WP	Residuum..........	17	4	21	...	2	24	...	4	...	31	12	1	...	41	90	10	43	I/S
FL3	Q1?................	5–7	...	2	...	...	34	9	3	...	37	8	9	...	22	99	1	...	Q>K>V>I
	Q2................	10–12	...	...	...	...	23	8	1	1	46	10	12	...	26	97	3	...	na
	Q2................	22	3	1	...	...	26	7	2	1	37	7	16	...	22	100	...	...	Q>K>V>I
	Q2................	27	...	1	...	...	18	5	3	1	39	12	20	...	26	100	...	...	Q>K>V>I
	Q2................	35	...	...	...	...	...	...	...	...	...	...	...	...	...	99	1	...	Q>K>I
	Q3a................	48	...	...	1	...	20	31	2	2	15	18	12	...	27	100	...	...	na
	Q3b	72	...	28	2	...	24	...	5	4	11	18	8	...	35	77	23	...	na
	Residuum..........	74	...	...	...	...	...	...	...	...	...	...	...	...	...	87	13	64	K>I
T3	Q2................	6	...	...	1	...	20	3	5	1	53	3	14	...	31	100	...	...	na
	Q2................	10	...	...	...	...	...	...	...	...	...	...	...	...	...	99	1	...	Q=K>I/S>I
	Q2................	20	...	...	...	...	43	8	4	...	23	15	8	...	24	100	...	...	Q>>K
	Q2................	25	...	...	...	...	45	15	6	1	16	12	5	...	24	100	...	...	Q>>K=I/S=I
	Q2................	27	6	1	...	...	41	6	6	...	22	9	10	...	28	100	<1	...	na
	Residuum..........	47	5	1	...	...	30	...	12	2	10	10	31	...	45	92	8	...	K>I>>Q
	Tampa Limestone...	57	7	8	3	...	21	4	13	1	19	9	14	...	50	84	16	85	C
T2	Q2................	17	3	1	1	...	19	11	5	...	37	5	17	1	40	100	...	...	na
	Residuum..........	27	5	9	...	...	30	2	14	...	16	11	13	...	84	84	16	71	I/S
Auger samples																			
2ndFL	Residuum..........	3–6	4	3	...	...	30	7	4	2	32	13	4	...	23	93	7	...	K=I/S>I
	Residuum..........	6–9	...	...	...	...	29	11	5	3	40	6	7	...	41	90	10	...	C>H=I/S=K>Sep
	Unknown	12.5–14	5	2	...	...	38	3	7	1	34	6	4	...	38	94	6	36	I/S=K>I
	Unknown	14–15.5	3	28	24	...	13	2	5	...	5	13	7	...	52	82	18	44	I+Sep
Surf2	Q2................	11–12	...	1	...	...	22	7	5	...	41	6	17	...	25	92	8	...	K>I
	Q2................	12–12.4	...	...	...	...	...	...	...	...	...	...	...	...	...	86	14	...	K=I>I/S

[1]Calcite weight percent was determined by digestion of a measured subsample by HCl. Calcite was not counted with quartz and feldspar. Some dolomite may be present.

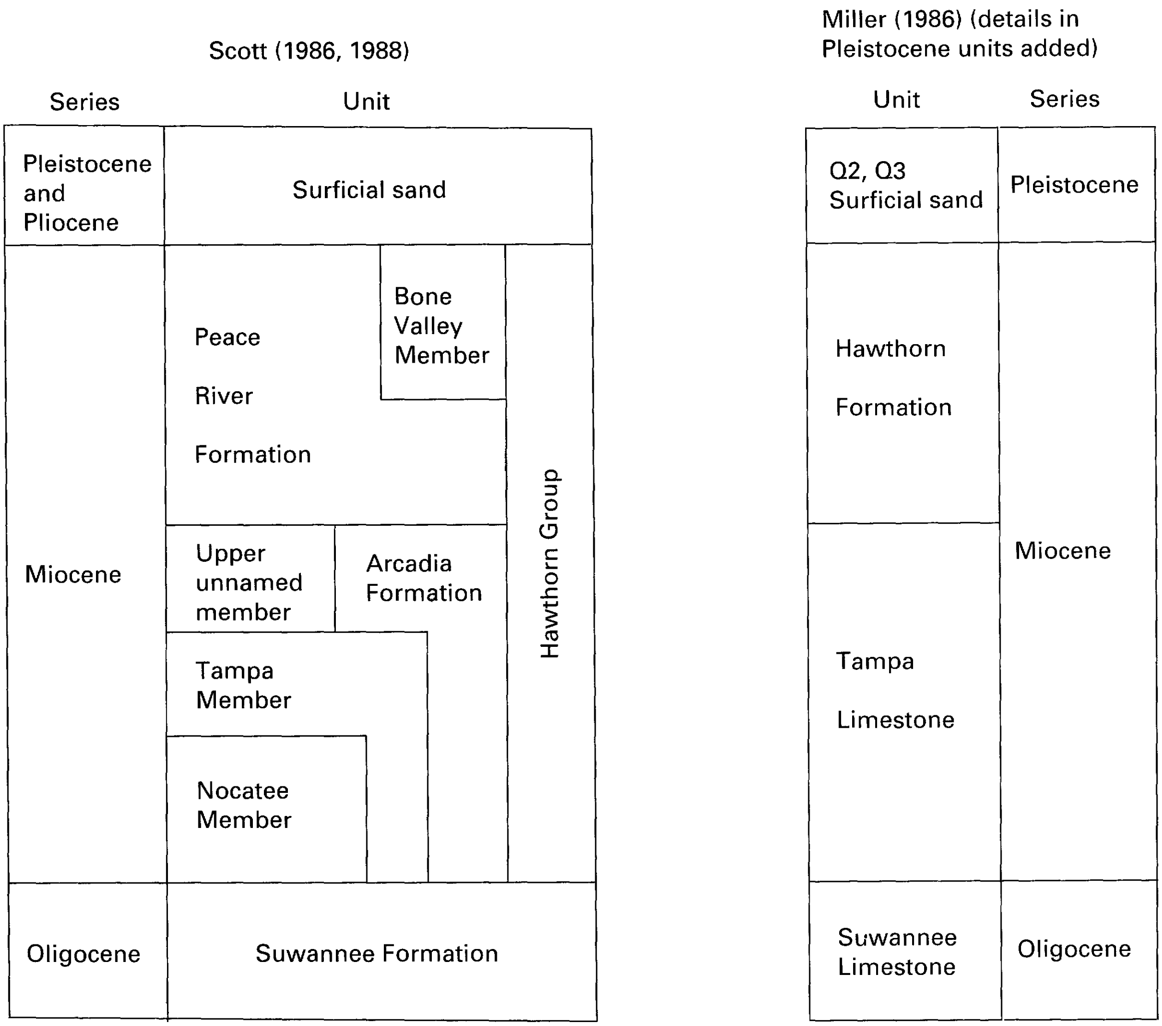

Figure 24.5. Stratigraphic nomenclature for Oligocene to Pleistocene units in Pinellas County, Fla.

Mineral Evidence of Weathering

This part of the discussion of the mineralogy of Pinellas County deposits is based on empirical studies of in-place weathering of Coastal Plain sediments along the Atlantic seaboard (Owens and others, 1983; McCartan and others, 1984). The mineralogy of surficial deposits in Pinellas County is similar to that in Atlantic Coastal Plain Quaternary deposits. The conclusions derived from the previous mineralogical studies follow: (1) if texture, composition, and depth are consistent in the section, then earlier deposits will have a lower proportion of nonresistant minerals than later deposits; (2) if sediment is of a single age, composition, and texture, then material nearer the surface will have a lower proportion of nonresistant minerals; and (3) abundant silt, clay, and authigenic cement tend to preserve the original mineralogy by impeding ground-water flow. An abundance of carbonate minerals would tend to buffer the ground water at pH=8.4 (Garrels and Christ, 1965), and this may retard leaching of minerals. This buffering may be the reason that garnet, feldspar, and illite/smectite are so abundant in the limestone and residuum.

The sand at the Pinellas site has very little fine-grained material, and so ground water flows freely through it and minerals undergo continuous alteration. Quartz is the most abundant mineral in the clay-sized fraction of the surficial sands. Clay-sized quartz is produced by physical disintegration of sand-size quartz grains due to chemical weathering (Pavich and others, 1989) and is typically found in sand dunes (Soller, 1988). Although some of the fine-grained quartz may be due to weathering processes at the site, most was probably deposited by wind and may have come from distant sources.

Most of the feldspar has altered to kaolinite and illite. Most of the small amount of illite/smectite has broken down into illite and smectite layers; all the smectite and some of the illite have altered to vermiculite. Garnet and other nonresistant heavy minerals have dissolved or have altered to complex oxides and hydroxides.

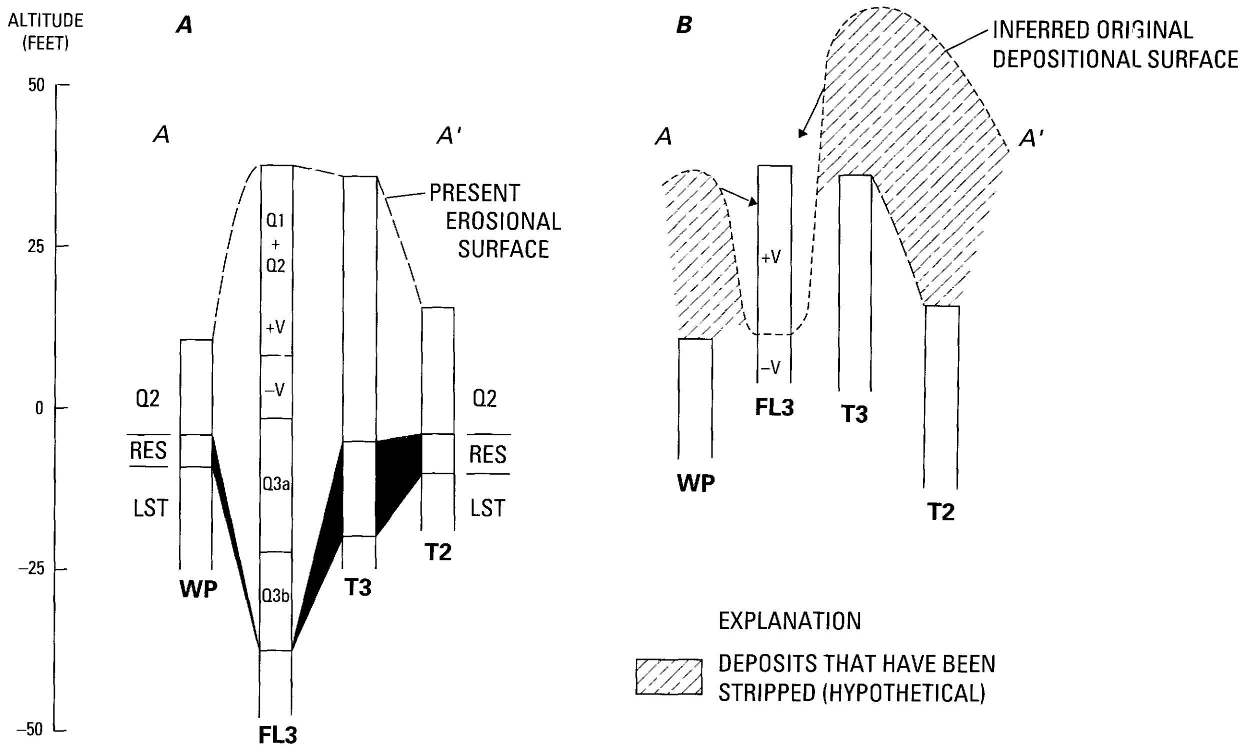

Figure 24.6. Stratigraphic section *A–A'* based on wells in which split-spoon samples were taken. See figure 24.1 for line of section. *A*, Stratigraphy inferred for the site: Q1, Holocene; Q2, Q3a, Q3b, Pleistocene; RES, residuum on Tampa Limestone; LST, Tampa Limestone (Miocene). V, dioctahedral vermiculite. *B*, The distribution of vermiculite, which is present only in the upper part of FL3, may be due to erosion of adjacent vermiculite-bearing deposits and infilling of the sinkhole axis at FL3, indicated by arrows.

The presence of dioctahedral vermiculite in FL3, a hole that penetrates the deepest part of a sinkhole (fig. 24.6*A*), is of particular importance. Although vermiculite is found down to at least 27 ft in FL3, it is not present in T3. Conversely, small amounts of illite/smectite were found at 10 and 25 ft below the surface in T3, but none is present in FL3, where the base of the sand is about 73 ft below the surface. Vermiculite is usually present in undisturbed weathering profiles more than 60,000 yr old (Owens and others, 1983; McCartan and others, 1984). The differences in FL3 and T3 clay mineralogy suggest that the sand at T3 is significantly less weathered than the sand in the bottom of the FL3 section and that the T3 section has been stripped of several feet of weathered sand (fig. 24.6*B*). Figure 24.6*B* extrapolates the area of eroded sands to T2 and WP although the clays from the surficial sand in those holes have not yet been studied. However, the least resistant of the heavy minerals, hornblende, is 3 percent of the heavy-mineral fraction at 17 ft in T2, which is comparable to the percentages at 22 ft in FL3 and 27 ft in T3. It may, therefore, be inferred that illite/smectite is also present. By analogy with T3, T2 also appears to have been stripped.

DISCRIMINATION OF PLEISTOCENE UNITS

Contacts at the top and base of the weathered limestone were determined in the field and are further supported by mineralogical studies. However, field observations alone were insufficient to delineate units within the surficial sand. Using all the available data (figs. 24.3, 24.4, and 24.6, tables 24.1 and 24.2), we conclude that the sand was deposited in beach and dune environments during one or more high stands of sea level in the Pleistocene and that some of the sand has been redistributed since the last sea-level high stand. The most significant mineralogical evidence supporting a pre-Holocene age is the presence of vermiculite.

The bulk of the sand can be assigned to one unit on the basis of mineralogy (figs. 24.3, 24.4, and 24.6*A*; tables 24.1 and 24.2). We designate that unit "Q2" to reflect our belief that it was deposited during the most recent pre-Holocene sea-level high stand and to distinguish it from Holocene deposits, which we designate "Q1." The only sediment that may be assignable to Q1 is the upper 5–8 ft of FL3; however, Q2 from the upper part of the weathered

Table 24.2. Selected mineralogical characteristics of stratigraphic units in the Pinellas County, Fla., sinkhole complex

[Abbreviations are defined in table 24.1. Semicolons separate the three classes of nonopaque minerals shown in table 24.1]

Age and unit	Fine sand fraction: Predominant nonopaque heavy minerals	Fine sand fraction: Potassium feldspar (percentage of non-carbonate fraction)	Predominant clay minerals
Late Pleistocene:			
Q2	St; Z	<3	K
Q3a	St,Sil; Z,T,R	<3	K
Q3b	G; St; T	23	unknown
Holocene to Miocene:			
Residuum on Tampa Limestone	St; Z,T,R	6–18	K, I/S
Miocene:			
Tampa Limestone	St; Z,R	16	unknown

section surrounding the deepest part of the sinkhole (at FL3) may have been eroded and washed into the depression in Q1 time. The absence of illite/smectite from what otherwise looks like the top of the Q2 sequence at FL3 may be due to reworking, which would permit more thorough oxidation of the minerals.

Between the depths of 40 and 74 ft in FL3 are two sand units that have heavy-mineral suites substantially different from that of Q2. In particular, the presence of abundant sillimanite at 48 ft, which is assumed to represent the 40- to 60-ft interval, is unique at the Pinellas County site. It is also unusual throughout the surficial sands of Florida (Martens, 1928, 1935). We designate this sillimanite-bearing sand "Q3a." The greater age signified by Q3 is justified only by its stratigraphic position. There is no firm evidence indicating whether Q3 was deposited during the same sea-level high stand as Q2 or during the previous one. The lowest and, therefore, oldest of the sand beds is between 60 and 74 ft in FL3; we have designated it "Q3b." Its sand mineralogy is similar to that of residuum and limestone at the site (tables 24.1 and 24.2). The absence of sillimanite indicates a change in sources between Q3b and Q3a; this change may have occurred over a long or a short time.

The mineralogy of samples from one auger hole, Surf2 (fig. 24.1), indicates that the unit sampled is Q2 (table 24.1). The other auger hole, 2ndFL, is unusual in that the clay-sized fraction of the sample from 6 to 9 ft below the surface contains abundant calcite and halloysite. Halloysite is one of the early weathering products of feldspar (Owens and others, 1983) and is related to kaolinite. Sepiolite, a Mg-rich fibrous clay, may also be present in small amounts. The most abundant opaque heavy mineral in this zone is pyrite, which was not found in other samples from the site. The material in the 2ndFl drill hole is probably in weathered Tampa Limestone (residuum); the zone from 6 to 9 ft was originally a clay-rich bed within the Tampa.

CONCLUSIONS

The preliminary results of this pilot study on the mineralogy of surficial sands in western Pinellas County have provided a basis for the interpretation of the provenance, depositional history, and weathering of the sands. The bulk of the surficial sand can be attributed to reworking of weathered limestone, but all the noncarbonate minerals in both the sand and the limestone were originally derived from metamorphic rocks in the Piedmont of Georgia. Weathering after deposition of the sand during the Pleistocene is mainly a result of oxidation caused by reaction of ground water with the minerals. The mineralogy of the Pinellas County site can now be compared with that of surficial sands at other Coastal Plain sites in Florida and elsewhere.

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25. Ion-Exchange Control of Radium-226 Activities in Ground Water of Sarasota County, Florida[1]

By Ronald L. Miller[2]

Water from wells in Sarasota County, Fla. (Kaufmann and Bliss, 1977; Miller and Sutcliffe, 1985), commonly exceeds the 5-pCi/L maximum contaminant level (MCL) set for Ra-226 (radium-226) plus Ra-228 in the National Drinking Water Regulations (U.S. Environmental Protection Agency, 1987). Concern that mining and processing of phosphate ore may increase radioactivity in the ground water of west-central Florida prompted several studies during the seventies (U.S. Environmental Protection Agency, 1973; Guimond and Windom, 1975; Irwin and Hutchinson, 1976; Kaufmann and Bliss, 1977). Kaufmann and Bliss (1977) studied Ra-226 in ground water of Sarasota County and surrounding counties and concluded (1) that elevated Ra-226 levels in ground water of Sarasota County result from natural enrichment processes, such as dissolution of minerals in the Hawthorn Formation or the migration from deep aquifers of mineralized ground water enriched in Ra-226, and (2) that occasional incidents of Ra-226 contamination of ground water in surrounding counties resulting from leakage from phosphate mines and chemical plants were likely to recur, but that the extent of contamination was not well defined because of monitoring deficiencies. A later study, however, revealed that radioactivity in ground water having low specific conductance did not migrate more than about a hundred meters from two phosphate chemical plants that were 14 and 34 yr old and from a 2-yr-old pond containing phosphate clay waste (Miller and Sutcliffe, 1984).

This report will compare the radioactivity found in ground water from phosphate mining areas and from Sarasota County and discuss the relation between the activity of dissolved Ra-226 and specific conductance. For this discussion, water having a specific conductance of 700 μS/cm at 25 °C or less will be called fresh. Water having a specific conductance of more than 700 μS/cm at 25 °C will be called high-conductance water.

Miller and Sutcliffe (1985) found that the highest Ra-226 activities in ground water of Sarasota County occur in the parts of the upper Hawthorn Formation and the overlying Tamiami Formation (both of late Tertiary age) that contain high-conductance water. High Ra-226 activities in ground water are associated with both of the two high-conductance water types found in Sarasota County: seawater and water containing a calcium magnesium strontium sulfate. They hypothesized that alpha-particle recoil ejects Ra-226 from the insoluble phosphate ore into ground water where the Ra-226 is subsequently scavenged by the abundant clays associated with phosphate deposits. Over time, these clays (montmorillonite, attapulgite, kaolinite, and illite, according to Boody and Barwood, 1982, p. 6) accumulate significant amounts of Ra-226 if ground water remains fresh. Where high-conductance water contacts clay, high concentrations of cations in the water compete with Ra-226 for ion-exchange sites on the clays and, thus, release more Ra-226 into the ground water. The high Ra-226 activities in ground water of the upper Hawthorn and the Tamiami Formations in Sarasota County probably occur because (1) high-conductance water moved into these formations recently with respect to the 1,600-yr half-life of Ra-226, or (2) the residence time of ground water is long enough for the alpha-recoil process to build Ra-226 activities in high-conductance ground water to their current levels. It seems likely that both processes are occurring and that their relative importance varies spatially and temporally.

Regression lines for Ra-226 activity and specific-conductance data (1) from the upper Hawthorn Formation in Sarasota County (Sutcliffe and Miller, 1981) and (2) from the Hawthorn Formation and mined land at an inland phosphate mine (Miller and Sutcliffe, 1982) were both significant at the 95-percent confidence level and were not significantly different from each other (fig. 25.1). These relations support the ion-exchange mechanism proposed above and can be used to explain why the phosphate-mining areas contain an abundance of uranium-238 decay series radionuclides but have lower levels of dissolved Ra-226 in ground water than parts of Sarasota County. In the

[1]Prepared in cooperation with the Florida Department of Environmental Regulation.

[2]U.S. Geological Survey, Tampa, FL 33634.

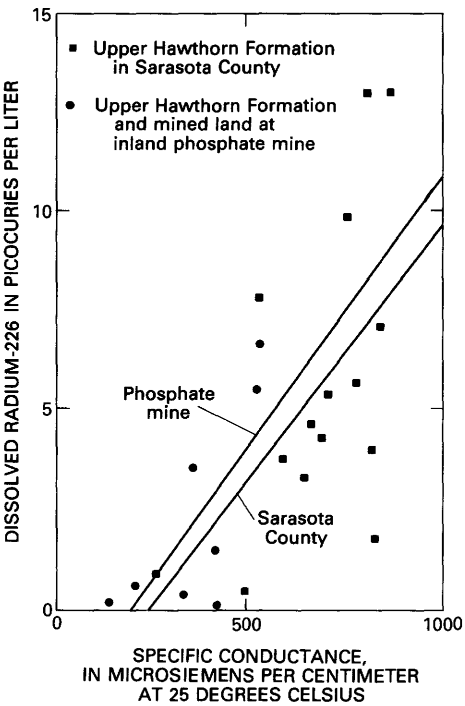

Figure 25.1. Comparisons of radium-226 activity and specific-conductance data in Sarasota County, Fla. (Sutcliffe and Miller, 1981), and an inland phosphate mine (Miller and Sutcliffe, 1982).

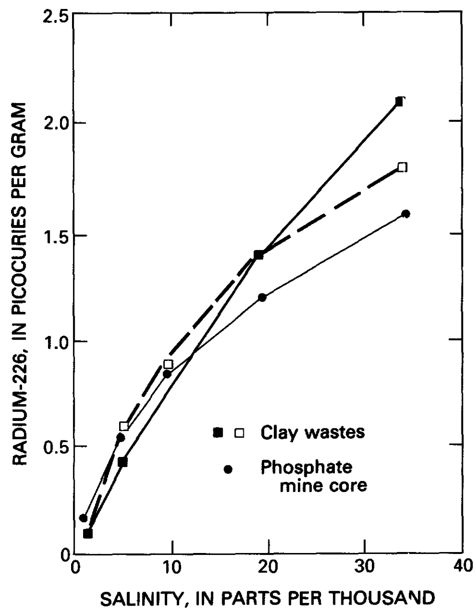

Figure 25.2. Release of radium-226 from three solid-phase samples.

phosphate-mining areas, the freshwaters of the Hawthorn Formation allow the clays to hold most of the Ra-226 released by alpha-particle recoil, whereas, in parts of Sarasota County, the high-conductance waters of the Hawthorn Formation allow more Ra-226 to remain in solution rather than adsorb to clays.

The ion-exchange release of Ra-226 from clays exposed to high-conductance water has been documented (Miller and others, 1990). Dilutions of seawater were mixed with parts of a well-mixed section of a sample core from a phosphate mine and two samples of phosphatic clayey wastes (slimes) from two phosphate mines. As shown in figure 25.2, the activity of dissolved Ra-226 increased with salinity to about 2 pCi of exchangeable Ra-226 per gram of air-dried solids at a salinity of about 35 ppt (parts per thousand). If one assumes a porosity of 0.4, a density of the solid phase of 2.7 g/cm^3, and a salinity of 35 ppt, then a Ra-226 radioactivity of about 8,000 pCi/L is calculated as an upper limit for pore water in an aquifer composed of these materials. The actual pore water activity, however, would be significantly less than 8,000 pCi/L because of the large water-to-solids ratio used in these experiments (5.1–11.5 g of air-dried solids in 900 mL of solution). These experiments demonstrate that, for the phosphate-mining area, (1) clays not previously exposed to high-conductance water contain a substantial amount of exchangeable Ra-226; (2) where clays are present, they control dissolved Ra-226 radioactivities; and (3) the release of Ra-226 is related to specific conductance and, therefore, to dissolved solids in the water.

Because a direct relation exists between dissolved solids and dissolved Ra-226, ground water from Sarasota County that contains more than 5 pCi/L of Ra-226 usually also exceeds the MCL's for hardness or dissolved solids. Consequently, ground water that is high in Ra-226 would probably not be used as a public water supply without receiving softening or reverse osmosis treatment, both of which also reduce Ra-226 below the MCL.

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26. Sedimentary Cycles and Sequence Stratigraphy in the Eastern Gulf of Mexico Coastal Plain

By Juergen Reinhardt[1,2]

INTRODUCTION

The sedimentary record of coastal transgressions and regressions along the passive southeastern margin of North America provides data for testing models of eustatic sea-level fluctuations. This paper presents a brief summary of the Upper Cretaceous stratigraphy mapped in the Chattahoochee River valley area in western Georgia and eastern Alabama (fig. 26.1) and discusses its relation to Cretaceous cycles depicted by Haq and others (1987). Each Upper Cretaceous formation in the study area contains facies deposited in middle shelf to alluvial-flood-plain environments that migrated across this area largely in response to shoreline changes associated with eustatic fluctuations in sea level. The architecture of the cycles resulting from these fluctuations is described below.

I mapped this area from 1976 to 1983 (1) because the formations are better exposed in the dissected topography than they are in most other parts of the eastern Gulf of Mexico Coastal Plain and (2) because the sedimentary record from at least the late Santonian through the middle Maastrichtian is fairly complete. U.S. Geological Survey biostratigraphers who collaborated with me in defining the ages of formations were Norman F. Sohl, who studied mollusks, Charles C. Smith (presently with Unocal), who studied the calcareous nannofossils, and Raymond A. Christopher (presently with Atlantic Richfield Company), who studied the pollen and spores.

STRATIGRAPHY, LITHOFACIES, AND SEDIMENTARY CYCLES

The Upper Cretaceous formations in the study area are, from oldest to youngest, the Tuscaloosa Formation, Eutaw Formation, Blufftown Formation, Cusseta Sand, Ripley Formation, and Providence Formation (fig. 26.1). The Tuscaloosa Formation unconformably overlies crystalline rocks of probable lower Paleozoic age and, in turn, is unconformably overlain by the Eutaw Formation. These contacts both represent substantial periods of subaerial exposure, weathering, and erosion; locally, paleosols, or fossil soils, are immediately below the contacts. The boundaries between the younger Cretaceous units are less obvious unconformities, but they are marked by sharp lithofacies changes and associated erosional surfaces.

During much of the Late Cretaceous Epoch, the shoreline trended east-west across the Chattahoochee River valley and then turned generally northwest-southeast west of the river (fig. 26.1). The shoreline position at any time was the controlling factor in the spatial distribution of continental and marine lithofacies. In downdip exposures along the Chattahoochee River valley, continental lithofacies are limited to the Tuscaloosa Formation, but in updip areas, continental lithofacies equivalent to the younger Cretaceous units, except the Ripley Formation, are found to the northeast in the drainage divide between the Flint and the Chattahoochee Rivers (fig. 26.1). The repetition of a limited number of lithofacies in the Chattahoochee River valley region was first noted by Eargle (1950). These sedimentary cycles have been described in considerable detail for both the Upper Cretaceous (Reinhardt, 1980, 1982) and lower Tertiary (Gibson, 1980; Gibson and others, 1982).

The Chattahoochee cycles are similar in their internal organization to cycles in the lower Tertiary of Texas (Fisher, 1964) and the Upper Cretaceous of New Jersey (Owens and Sohl, 1969). They consist of a basal transgressive phase, an inundative or highstand phase, and a regressive phase at the cycle top. In general, the cycles are highly asymmetrical, as the basal transgressive phase makes up less than 10 percent of the cycle, and the other phases make up the rest of the cycle's thickness. Overall, the sedimentary cycles record a rapid transgression followed by marine sedimentation and a gradual shoreline regression.

The composition of each phase of a sedimentary cycle depends on the extent of the marine trangression and the paleogeographic position of any given geologic section. The transgressive phase generally consists of a basal sand,

[1]U.S. Geological Survey, Reston, VA 22092.
[2]Deceased.

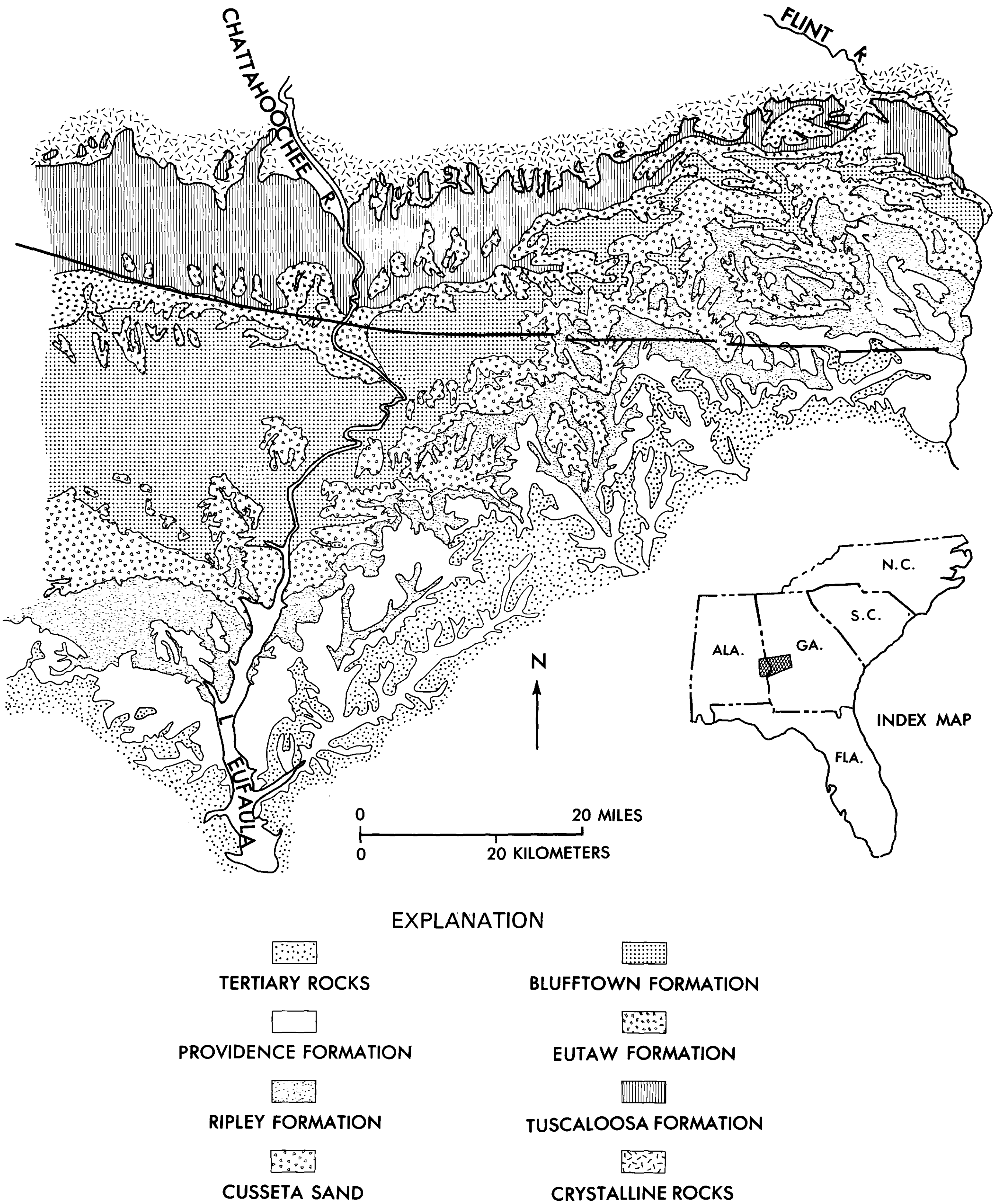

Figure 26.1. Generalized geologic map of the Cretaceous units in the Chattahoochee River valley region on the northeastern margin of the Gulf of Mexico Coastal Plain. Heavy line trending generally east-west is the approximate maximum inland shoreline position during the Eutaw transgression. Figure modified from Reinhardt (1980).

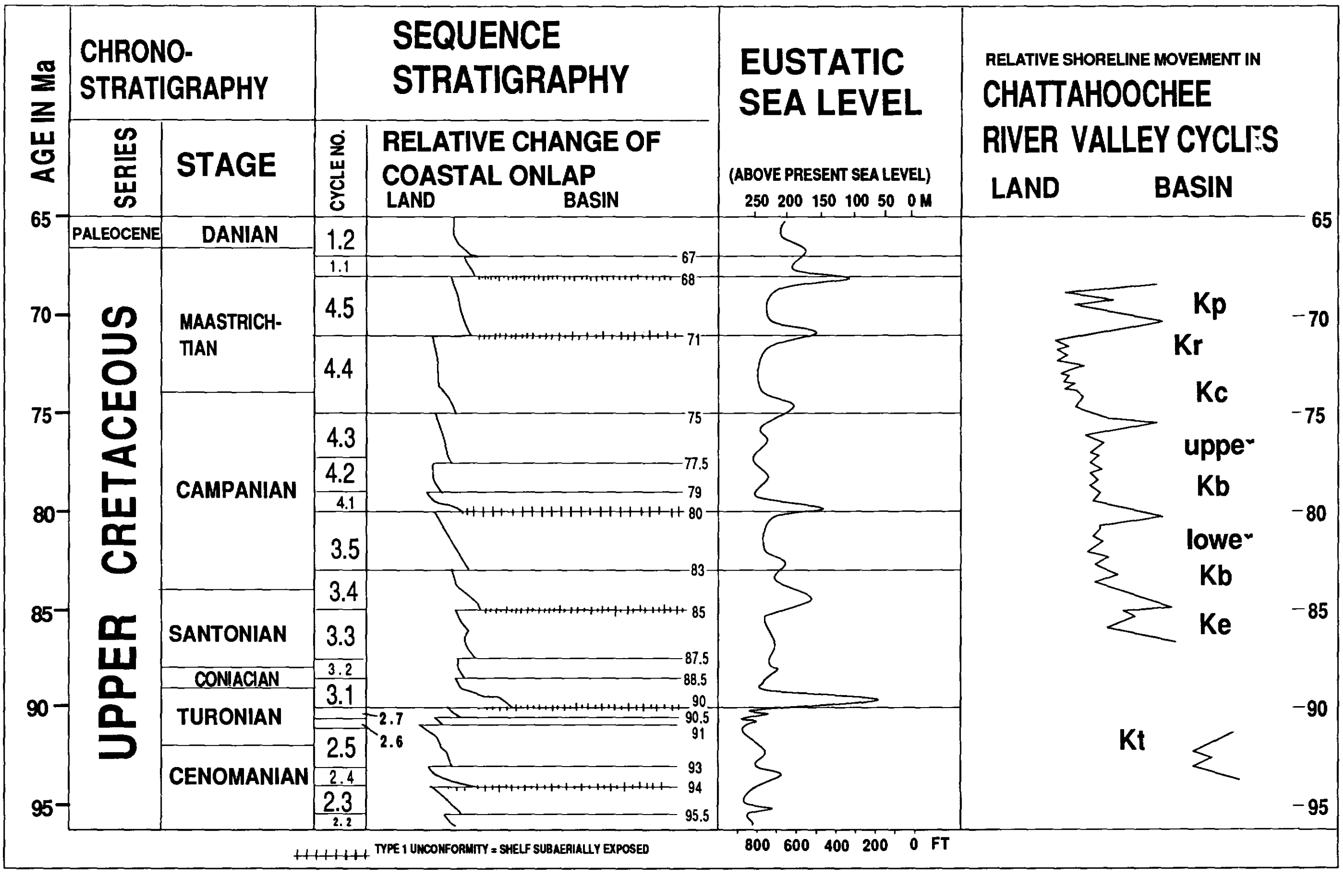

Figure 26.2. The relation of Upper Cretaceous depositional cycles in the Chattahoochee River valley to the model of coastal onlap and eustatic sea-level curves of Haq and others (1987). Kt, Tuscaloosa Formation; Ke, Eutaw Formation; Kb, Blufftown Formation; Kc, Cusseta Sand; Kr, Ripley Formation; and Kp, Providence Formation.

commonly crossbedded, and an overlying massive transgressive lag deposit or a ravinement. In downdip areas, a transgressive lag bed (containing phosphate and quartz pebbles, reworked shell, and bone) immediately overlies the top of the preceding cycle, as little sediment accumulated during the marine transgression.

The inundative phase contains the deepest water deposits; most of these are massive to fissile marine clays containing a diverse but diminutive fauna, which typically occurs near the base of the phase. Fine glauconitic sands and shelly marls make up most of the downdip sediments deposited during the inundative phase, whereas crossbedded sands representing tidal inlet and strandplain deposits are preserved in updip areas. Shoaling in the sedimentary basin, as sedimentation exceeded subsidence, delimited the top of the sedimentary cycle.

At this stage, the barrier shoreline typically "jumped" seaward as offshore sandbars emerged, and back-barrier sediments accumulated in the lagoon behind them. The regressive phase varies in thickness and generally consists of laminated carbonaceous silt and clay or massive, mottled silty clay. In updip areas, alluvial-flood-plain clays and lensoid, crossbedded, alluvial-channel sands accumulated during this aggradational phase.

SHORELINE MIGRATION AND SEQUENCE STRATIGRAPHY

A sea-level-fluctuation curve inferred from shoreline movements was first published for the Cretaceous deposits of the Chattahoochee River valley by Reinhardt (1980). Numerous stratigraphic refinements have been suggested since that time (for example, by Donovan, 1985; Skotnicki and King, 1986, 1989; Frazier, 1987), and several models have been proposed to explain the relative timing and distribution of major transgressive and regressive events recorded in the eastern Gulf of Mexico Coastal Plain. The chronology of deposition is based largely on the biostratigraphy of the deposits as summarized by Sohl and Smith (1980). Publication of sea-level-fluctuation curves for the Cretaceous by Haq and others (1987) permits their comparison with cycles in the Chattahoochee River valley area (fig. 26.2), and the remaining text will be such a comparison.

Deposition of the Tuscaloosa Formation appears to have been temporally coincident with the highest inferred worldwide sea levels for the Mesozoic and Cenozoic Eras. Prior to the deposition of the Tuscaloosa, the eastern Gulf of Mexico Coastal Plain was emergent and tectonically stable, as indicated by the preservation of mature paleosols on the bedrock surface underlying the Tuscaloosa (Sigleo and Reinhardt, 1988). The pollen record indicates that the Tuscaloosa was deposited in the Chattahoochee River valley during the late Cenomanian and possibly the early Turonian (Christopher, 1982). The unlikely association of a major clastic wedge with an inferred maximum sea-level highstand is most readily explained by assuming that substantial uplift was taking place in the Appalachians at this time.

About 4–5 m.y. elapsed between deposition of the top of the Tuscaloosa and the marine transgression marking the base of the Eutaw Formation. Very high worldwide sea levels are inferred from the depositional record in places like the western interior of the United States during this period for which the Chattahoochee River region has no record. Therefore, subsidence of the region must not have taken place prior to the Santonian, when the area was first inundated by the sea. Beginning with the base of the Eutaw, deposition for the next 19–20 m.y. was relatively continuous, and interruptions (nondeposition or erosion) were less than a single biostratigraphic zone (probably less than 0.5 m.y.) at cycle boundaries. This relatively continuous deposition suggests that the rate of subsidence or downwarping along the inner edge of the Coastal Plain was relatively constant and probably approximated the rate of accumulation, roughly 20 m/m.y.

The relation of Eutaw lithofacies to shoreline position and migration has been carefully considered by Frazier (1987). Frazier suggested that the lower part of the Eutaw was deposited during the transgressive phase in cycle 3.3 of Haq and others (1987) and concluded that this deposition took place during about 1.5 m.y. Because much of the Eutaw cycle consists of back-barrier facies (regressive phase of the cycle), rather than of highstand deposits, and the transgressive lag is at or near the base of the unit, I would argue that little record of the initial Eutaw coastal onlap has been preserved. Rather, the basal transgressive lag was probably deposited close in time and space to the updip limit of the maximum marine transgression.

Mollusks and poorly preserved calcareous nannofossils in the basal Eutaw suggest that the unit may be as old as early Santonian or that the base may be as young as late middle Santonian (Sohl and Smith, 1980). If a minimum age for the basal Eutaw is assumed, then the topmost Eutaw sediments were deposited at about 85 Ma, the age of the downlap surface of Haq and others (1987). The regressive phase of the Eutaw began to be deposited as soon as the basin shoaled enough for a new barrier system to form seaward of the outcrop belt. The landward migration of that barrier system across the restricted facies of the Eutaw marks the base of the Blufftown Formation.

The 150-m-thick Blufftown presents an outstanding, yet relatively untested, opportunity for comparing Chattahoochee cycles with the sequence stratigraphic model of Haq and others (1987). Reinhardt (1982) described two cycles, separated by a transgressive lag, for the Blufftown on the basis of exposures along the Chattahoochee River. The base of the lower cycle is close to the Santonian-Campanian stage boundary, and most of the cycle is of early Campanian age; see Sohl and Smith (1980) for a discussion of the basal age assignment of latest Santonian versus early Campanian based on nannofossils and mollusks, respectively. The upper Blufftown is thought to be middle and late Campanian.

The boundary between the two cycles within the Blufftown has not been accurately placed biostratigraphically, but it may correspond to the major cycle boundary between cycles 3.5 and 4.1 of Haq and others (1987). Using different criteria for shallowing and deepening of the basin, Skotnicki and King (1986) postulated three sedimentary cycles for the Blufftown. In either interpretation, most Blufftown deposits represent inner to middle shelf deposits (highstand deposits) consisting of massive, glauconitic fine sands and marls.

Within the time interval of lower Blufftown deposition, two cycles are shown by Haq and others (1987). Although the sequence boundary shown within the lower Blufftown interval is not considered a major boundary, we should look for some indication of the postulated shallowing, especially in updip areas. The eustatic curves and the coastal onlap pattern suggested for the upper part of the Blufftown are even more complex, as three episodes of highstand deposition were separated by times of downlap. The cycles of Haq and others (1987) imply that considerable detail in terms of sea-level fluctuation could be more completely resolved or tested for in parts of the Blufftown Formation in the Chattahoochee River valley.

Mollusks have been used to assign the Cusseta Sand to the late Campanian and earliest Maastrichtian (Sohl and Smith, 1980). The contact that separates the upper Blufftown cycle from the basal Cusseta appears to be coincident with a major sequence boundary in the sequence stratigraphy of Haq and others (1987). In outcrop, scattered pods of carbonaceous clay at the top of the Blufftown are evidence for abrupt shallowing. The coarse sands at the base of the Cusseta are widespread, thicken updip, and appear to correspond well to the transgressive phase of cycle 4.4 of Haq and others (1987).

The transition to highstand deposits represented by the Ripley Formation was associated with a widespread marine transgression in western and into central Georgia. Such a widespread transgression is unexpected from the model of Haq and others (1987) because the relative coastal onlap is shown as less than onlap during other parts of the

Cretaceous. We might, therefore, speculate that subsidence of the basin during the early Maastrichtian might have been greater than subsidence at other times during the Late Cretaceous.

Skotnicki and King (1989) have postulated two progradational cycles in the Ripley on the basis of well logs in central and eastern Alabama. They have equated those depositional cycles with eustatic cycles 4.4 and 4.5 on the chart by Haq and others (1987). This interpretation creates substantial problems in reconciling the placement of the overlying Upper Cretaceous units.

Following deposition of the Ripley, the area was subaerially exposed. Donovan (1985) documented the incised topography that separates the Ripley and Providence Formations and the valley fill deposited during a marine lowstand in the middle to late middle Maastrichtian. Donovan (1985) informally defined the Alexander's Landing beds for this stratigraphic interval. Although the unit has not been mapped in detail, its surface and subsurface distribution resolves anomalous biostratigraphic relations between units previously mapped as Providence Sand and Ripley Formation in the Chattahoochee River valley (Sohl and Koch, 1986). The Alexander's Landing beds are a lowstand wedge that may have been deposited at the beginning of cycle 4.5 of Haq and others (1987).

Donovan (1985, 1986) divided the Providence Formation into an upper and lower member. These units were defined as shallowing-upward sequences separated by a transgressive disconformity. Donovan's level of stratigraphic resolution, as well as his detailed analysis of the shoreline configuration and changes in the Providence, greatly exceeds that in the analysis of the underlying Cretaceous units. The two depositional cycles identified by Donovan in the Providence have no counterpart in the cycles of Haq and others (1987).

The downdip equivalent of the Providence is the Prairie Bluff Chalk. The substantial hiatus between these units and the overlying Clayton Formation (Danian) has been the focus of considerable study in central Alabama because of its implication for Cretaceous-Tertiary boundary questions (Reinhardt and others, 1986). In downdip areas, the hiatus may have been as short as 1 m.y., and deposition may have begun again prior to the beginning of the Tertiary, but in the Chattahoochee River valley section, the hiatus appears to have been 2–3 m.y., and it included the boundary interval. For a review of the boundary relations, see Jones and others (1987), Donovan and others (1988), and Mancini and others (1989).

CONCLUSIONS

The correlation between Upper Cretaceous sedimentary cycles related to shoreline migration across the eastern Gulf of Mexico Coastal Plain and the cycles of Haq and others (1987) is generally good from the middle Santonian to the middle Maastrichtian. Preservation of specific transgressive events was controlled in large part by the subsidence history of the northeastern margin of the Gulf of Mexico basin. Although several worldwide sea-level highstands have been reported in the late Early and early Late Cretaceous, there is no record of marine deposits associated with these transgressions in the surface geology of the eastern Gulf of Mexicio Coastal Plain.

Although each of the unconformity-bounded units in the Chattahoochee River valley section has unique sedimentary and faunal characteristics, similar lithofacies patterns are preserved in each sedimentary cycle. Most contacts between sedimentary units represent relatively short depositional hiatuses and can be related with greater or lesser confidence to the major and some minor sequence boundaries in the sequence stratigraphy of Haq and others (1987). The Eutaw Formation cycle correlates well with the upper part of cycle 3.3. The lower Blufftown cycle corresponds to cycles 3.4 or 3.5 or both. Similarly, the upper Blufftown cycle is within cycles 4.1, 4.2, and (or) 4.3. The two Blufftown cycles appear to be separated by a major sequence boundary.

The Cusseta corresponds to the transgressive phase of cycle 4.4, and the Ripley Formation consists of highstand deposits in that same cycle. Finally, the Providence Formation consists of two depositional sequences, which probably accumulated during cycle 4.5. The major lowstand that followed the deposition of the Providence in the early to middle late Maastrichtian led to emergence of the entire Gulf of Mexico Coastal Plain until shortly before the end of the Cretaceous. The marine transgression associated with the oldest Tertiary deposits in the eastern Gulf of Mexico Coastal Plain did not deposit sediments in the Chattahoochee River valley prior to the middle Danian.

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