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SOUTHEASTERN GEOLOGY

Table of Contents

Vol. 13, No. 1

1971

DIFFERENTIAL EROSION OF CARBONATE-ROCK TERRANES*

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ABSTRACT

Relief in carbonate-rock terranes may be local and small as expressed by many shallow sinkholes, or large as expressed by poljes and escarpments. In addition to relief within the karst terrane, there is commonly significant relief between a belt of carbonate rocks and an adjacent belt of noncarbonate rocks. Cuesta topography generally occurs where the adjacent rocks are slightly inclined; in some humid regions carbonate rocks lie under an insoluble caprock and crop out near the base of escarpments, but cases are numerous where carbonate rocks themselves cap the cuestas. Carbonate rock valleys between sandstone ridges are common in some humid regions, but ridge-forming carbonates between shale valleys are also common.

Differential erosion results from a combination of physical and chemical processes. There are prerequisites for both physical and solutional erosion. Rock must be decomposed or disintegrated before it can be physically removed; a permeable soil cover is necessary for effective solution of carbonate rocks.

Erosion in carbonate terranes is favorable under moderate rather than under extreme conditions of cover, purity of the carbonate rock, topographic relief, and precipitation. Denuded carbonate rocks are much more resistant to physical and chemical erosion than are carbonate rocks with a moderately thin soil and vegetal cover; where the soil and rock cover is very thick, physical erosion of the covered bed is impossible and chemical erosion may be retarded because of retarded water circulation. Further analysis indicates that some of these extremes are in turn related to extremes in cover. Pure carbonate rocks yield no insoluble residue to form a cover; also, intense precipitation tends to strip off a thin cover and to keep the rock denuded; also, if the

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carbonate rocks are relatively impermeable, water cannot easily penetrate the rock to encourage soil development. Thus, the degree of cover on a carbonate terrane is an important key to differential erosion and to much of the topographic relief.

INTRODUCTION

The principle of differential erosion is universally accepted. It focuses attention on more resistant rocks standing out higher in topographic relief than do less resistant rocks. The mental image one gets from this principle is so apparently clear that elaboration is seldom pursued; in fact, in most areal geologic reports one sentence commonly covers the point that a particular formation is more resistant than the adjacent formation. Carbonate rocks commonly are either less or more resistant to erosion than nearby rocks of other kinds. That they seldom respond to erosive forces in the same way as these other rocks invites a study of the factors that produce the observed differences. It is the purpose of this paper to evaluate the large topographic features of carbonate formations and to emphasize the factors that are involved in differential erosion where carbonate formations occur. An understanding of the development of the broader or larger topographic forms of carbonate rocks allows one to make better inferences of subsurface hydrologic conditions than would otherwise be the case.

Acknowledgments

The authors are indebted to our colleagues, especially to C. L. McGuinness and George H. Davis of the U. S. Geological Survey, for critical discussions and advice in the preparation of this paper. Watson Monroe, William A. White, and W. D. Thornbury kindly reviewed the manuscript and offered helpful suggestions. Miss Beulah Hilliard and John Teel, U. S. Geological Survey, were especially helpful in the final preparation of the manuscript.

EXAMPLES OF CARBONATE-ROCK TOPOGRAPHY

A wide range of topographic conditions occur on carbonate terranes. The term "karst topography" has been applied to carbonate-rock areas where sinkholes abound and streamless valleys are common. Even karst forms differ from place to place, some of these differences having led to the concept of a karst cycle (Grund, 1914, p. 621-640) in which a distinct sequence of topographic forms has been recognized. Youthful forms appear early and consist of sinks, which increase in number and are progressively enlarged. Drainage, which originally was surface runoff, is diverted underground. The youthful forms

gradually give way to mature forms, such as relatively large dry valleys. Solution valleys are the result of the coalescence of sinks, the sapping action of springs, or a combination of the two. Drainage is still largely underground. In the old-age stage the ridges have been reduced, and the general level of the land surface then approaches that of a plain.

It should not be implied that carbonate areas pass distinctively and in an orderly manner through the various stages of development. Rather than placing emphasis on stages of development of karst forms, it is sufficient here only to call attention to the variety of forms that are commonplace. In addition to topographic forms of carbonate rocks, consideration is given here to the contrasting topographic relief that almost everywhere has developed where carbonate rocks are adjacent to noncarbonate rocks. In many places carbonate rocks are topographically lower than adjacent noncarbonate rocks, but in many other places they are higher.

The following categories of topography that have developed on carbonate rocks are common and serve to show the great range in conditions. Figure 1 shows a view of each category.

1. Undulating plain with sinks - rolling topography consisting of sinks and intervening convex slopes - (examples, (a) Blue Grass Region of Kentucky, where local topographic relief is subdued and (b) the Cockpit Country of Jamaica, where local topographic relief is appreciable (not illustrated in Figure 1).)

2. Denuded homoclinal plain with sinks - dip of limestone and land surface coinciding to make a homoclinal coastward slope with scattered sinks, some of which have vertical walls - (examples, Nullarbor Plain of South Australia (Jennings, 1963) and north coast of Yucatan, Mexico).

3. Poljes - flat alluvial valleys bordered by relatively steep, bare limestone ridges, the valleys ranging from one-half mile to several miles in width and being somewhat elongated - (examples, valleys of southwestern Yugoslavia--some of which are many miles in length--and Nassau Valley, Jamaica).

4. Haystack or mogote plain - a nearly flat plain underlain by limestone and veneered with alluvium or residual soil on which conical limestone hills (mogotes) rise as much as 300 feet above the plain - (examples, parts of the north-coast limestone of Puerto Rico (Monroe, 1966) and parts of North Vietnam and south China).

5. Carbonate valley between noncarbonate hills - valley formed on carbonate rocks lying between ridges composed of more resistant sandstones and shales - (examples, Shenandoah Valley of Virginia, and the Sequatchie Valley of Tennessee and Alabama). These generally are breached anticlines or other structures where carbonate rocks are exposed after erosion of the overlying less soluble beds.

6. Noncarbonate lowland between carbonate ridges - valleys formed on shale or other noncarbonate rocks lying between denuded

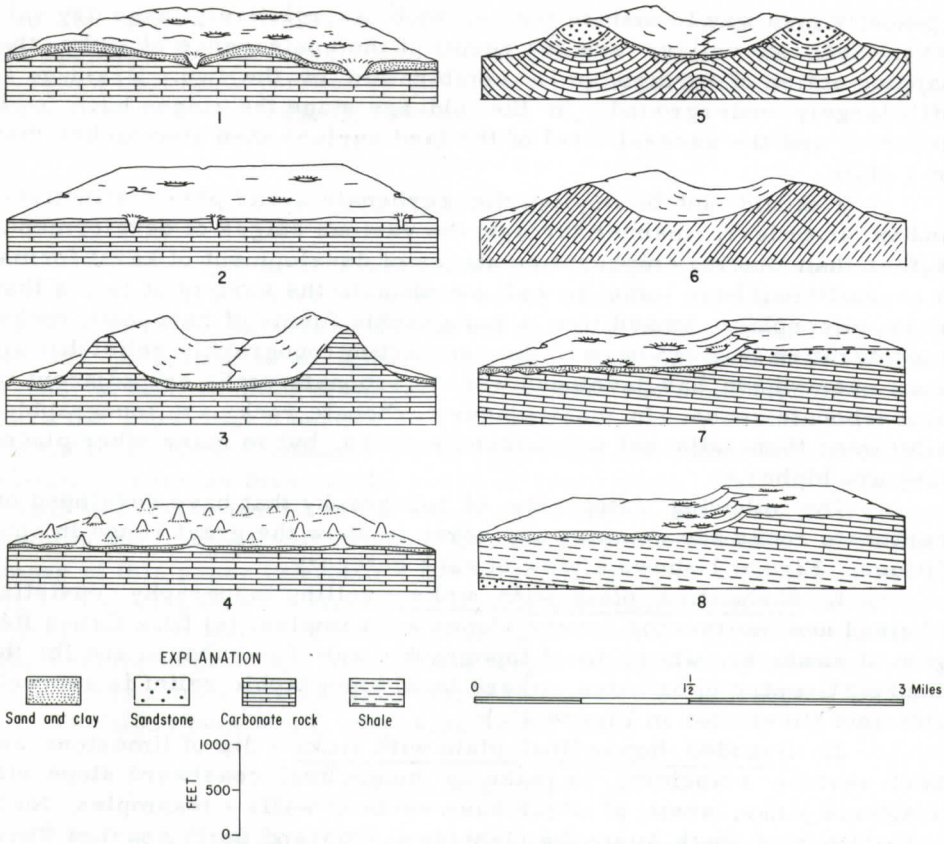

Figure 1. Types of landscapes in which differential erosion occurs on carbonate rocks (see text for explanation).

carbonate ridges - (examples, Sierra Madre near Monterey, Mexico (Lesser-Jones, 1967) and Silla Gibara area of east Cuba).

7. Escarpment and low-lying limestone plain - nearly flat plain underlain by carbonate rocks, mantled by thin soils, and bordered by a subdued escarpment and an upland plain capped by less soluble rocks - (examples, Nashville Basin-Highland Rim area of Tennessee, Dougherty Plain-Tifton Upland of south Georgia, Salem Plateau-Springfield Plateau of Missouri, and the Dripping Springs-Chester escarpment of Kentucky and Indiana).

8. Escarpment and higher plateau capped by carbonate rocks - cuesta on which carbonate rocks are more resistant than are the underlying rocks - (examples, Niagara limestone of Wisconsin, London Chalk of southeast England, and several cuestas in the Western Desert of Egypt).

PHYSICAL AND SOLUTIONAL EROSION

An exhaustive description of physical and solutional erosion of rocks is unnecessary for the purpose of this discussion, but it is necessary to state briefly some basic principles that have been accepted almost universally. The principles have real significance when placed in the context of the most favorable and least favorable conditions for erosion. In this context the differential erosion that characterizes carbonate terranes can be viewed to advantage.

Physical Erosion

Basic requirements for the full progression of physical erosion of interior regions are: the decomposition of rock into detritus, the movement of the detritus on slopes into the river bed, and the erosion of the river bed itself. General uplift of the land to provide interstream and stream slopes and ample precipitation to activate the drainage are requirements implied within this progression.

Elaboration of these basic requirements for physical erosion has led to different concepts of erosional development of landforms. Davis (1909, p. 350-380) pointed out that land features developed by stages, the ultimate development being maturely dissected peneplains unless the cycle were renewed. Hack's concept of dynamic equilibrium stresses the tendency for all elements of the topography to adjust to differential erosion so that they are downwasting at about the same rate (1960, p. 85). Both of these concepts have merit, but a discussion of them is avoided here so that only concise accepted principles may form the framework for the aspects of differential erosion to be discussed later.

Facts that are basic to an understanding of differential erosion include: (1) exposure of rocks above sea level to circulating water and air, and (2) different rates of erosion of a stream channel and adjacent interstream areas either now or at some early stage of erosion. If a perennial stream has a steep gradient and is incised in easily erodable rocks the stream is likely to be lowered at a faster rate than the adjoining hill slopes; this is especially true where the slopes are denuded and do not decompose to form detritus to be moved down slope. On the other hand, if a stream has a low gradient and is not aggressively lowering its channel the adjoining hill slopes may be lowered at a faster rate, especially if both weathering and rainwash are active on the interstream areas. Parts of many streams adjust their courses to coincide with easily erodible rocks or with other weaknesses such as joints and faults. This is a common form of differential erosion. On broad interstream areas differential erosion is also manifested by more resistant rock standing topographically higher than less resistant rocks. Rocks not easily fractured and not covered with soil and other detritus tend to be resistant.

Regardless of the prior causes for soil to be removed from resistant rocks that stand above less resistant rocks, the absence of soil per se commonly tends to perpetuate the resistance to erosion. As noted by many workers, this condition results partly from the moisture-shedding tendency of bare rock which in turn retards weathering and soil-forming processes. Wahrhaftig (1965, p. 1165) reports the more rapid weathering and erosion of granite in Southern Sierra Nevada, California, where covered with disintegrated weathered material than where the granite is bare. The relative purity of many carbonate rocks results in thin residual soils that are easily removed by physical erosion; yet, where the soils are removed new carbonate rock soils are not easily regenerated, and bare rock tends to be perpetuated.

Solutional Erosion

Favorable conditions for solutional erosion are: soluble rocks (such as carbonates), openings in the rocks that allow circulation of water, positioning of the rocks at or near land surface where water can enter the rocks and where it can discharge from them, and a permeable soil cover that yields water rich in carbon dioxide and undersaturated with respect to calcium carbonate. Unless all these conditions are favorable, solutional erosion may be slight. For example, carbonate rocks that have no soil and that are elevated above valleys erode slowly where fractured and very slowly where unfractured (Monroe, 1966, p. 7) (Stringfield and LeGrand, 1969, p. 374). Also, where carbonate rocks, even highly permeable, are buried deeply so that water cannot easily discharge from them the solutional erosion is negligible. Relatively pure carbonate rocks tend to produce very little residual soil, and where the soil cover is stripped away by physical erosion a reduction in solutional erosion occurs (Stringfield and LeGrand, 1969, p. 374). Thus, pure carbonate rocks may be no more favorable for solutional erosion than impure carbonates and under some conditions are less favorable.

A soil and vegetal cover is an important requirement for solutional erosion for the following reasons. A moderately permeable soil cover retards runoff and promotes slow and prolonged infiltration of water into the network of openings in the rock. The aerated soil zone results in increased carbon dioxide content of the soil water that is undersaturated with respect to calcium carbonate. In some organic-rich soils there is an input of carbonic acid into the soil water. The soil waters can aggressively attack and dissolve the underlying carbonate rocks to a much greater degree than can rainwater that has not had a chance to pick up carbon dioxide.

Solubility of different kinds of carbonate rocks is a complex physical and chemical problem that should be studied under natural conditions as well as in the laboratory. In separate investigations of the solubility of dolomites, Hem (1959) and Zogovic (1966, p. 39)

concluded that the ratio of solubilities of limestone and dolomite is very variable, depending on the composition and texture of the two kinds of rocks and the natural conditions for solution.

Some factors affecting the permeability and solubility of rocks are developed prior to exposure to subaerial erosion; these include texture, amount of soluble materials, and stratigraphic relations. Other factors coming into play after the carbonate rocks are exposed to erosion may have overriding effects that make the difference in solubility between limestones and dolomites insignificant in most cases; these factors center on the extent to which the soluble rocks lie in an active ground-water circulation system.

COMBINATIONS OF DIFFERENTIAL EROSION IN CARBONATE-ROCK TERRANES

Differential erosion is conventionally considered in terms of local differences in physical erosion, especially where relatively insoluble rocks are concerned. Local differences in solutional erosion are distinctive in carbonate terranes where karst topography occurs. In many carbonate terranes, solutional erosion is dominant; in many others, it is a complement to physical erosion. The complementary relation between physical and solutional erosion is not everywhere easy to observe or understand, but an appreciation of the importance of each in specific areas may result in a better understanding of hydrologic and other subsurface conditions. Conditions may be favorable or unfavorable for physical erosion, and favorable or unfavorable for solution. These four combinations of conditions lead to differences in the erosional response of a particular terrane. Within each of the four combinations of erosional conditions the ratio of physical to solutional erosion is relative.

Differential erosion implies that a certain topographic site stands higher than an adjacent site. The scale may range from micro to macro--from the bottom of a small sink to its rim and from a large valley in shale to the bordering limestone mountain. Some topographic conditions may now be analyzed as to differential erosion.

Where carbonate rocks with joints and other openings are mantled by a cover of soil through which water from frequent precipitation passes and where perennial streams are closely spaced, it is reasonable to expect both physical and solutional erosion to be active. Such conditions are likely to be quickly passing stages, displaying topography corresponding to early youthful forms. However, as water moves preferentially through the largest fractures and other openings toward a zone of discharge it enlarges the openings through solutional erosion. Thus, water can move downward more quickly, and the water table progressively lowers itself in the terrane. As sinks develop from removal

of soluble near-surface carbonates, the open drainage essential for physical erosion tends to disappear, and solutional erosion becomes predominant (1 and 2 of Figure 1).

Where soils are absent but where fractures in outcropping carbonate rocks allow water to infiltrate readily, solutional erosion may not be great but is still predominant over physical erosion. Examples are denuded carbonate plains, mogotes, and carbonate ridges. The solutional erosion in this case (2, 3, and 4, of Figure 1) may occur at a slower rate than in soil-covered carbonates. Solutional erosion is predominant over physical erosion but also quite slow where flat-lying and soil-covered and relatively impermeable carbonate rocks occur. In this case sinks may be absent and solutional erosion occurs slowly as water moves slowly along the contact of the soil and underlying bedrock. Such conditions occur in the chalk of the Selma Group of Alabama and Mississippi. This condition has been observed also on relatively soluble gabbros of North Carolina, South Carolina, and Virginia (LeGrand, 1952).

Solutional erosion is most pronounced in karst regions. Other factors being equal, the most favorable condition for solution erosion that initiates karstification requires some type of less soluble soil cover or geologic formation on the surface of the carbonate rocks. In general, there is evidence of two types of geologic cover in karst regions. One consists chiefly of unconsolidated deposits, such as the blanket sands of northern Puerto Rico (Briggs, 1966) and insoluble residue of the carbonate rocks. The other type of cover is a geologic formation that contains less soluble consolidated beds, which forms solution escarpments. An example of this type of geologic cover is the Hawthorn Formation of Miocene age that was deposited on a thick sequence of Tertiary carbonate rocks in the Southeastern States. Karstification of a large region of carbonate rocks in Florida and Georgia was initiated as the Hawthorn Formation was removed chiefly by retreat of escarpments. Another well-known example of this type of cover consists of Mississippian and Pennsylvanian formations in karst regions of Tennessee, Kentucky, Indiana, and the Ozark region of Missouri.

Sandstone formations as a cover on carbonate terranes form some of the most pronounced cuestas or escarpments of karst regions, although less pronounced escarpments are formed within the carbonate rocks after they are exposed to solutional erosion.

Although some semblance of escarpments may develop with the unconsolidated type of cover, the initiation of karstification may begin over the entire carbonate region as surface streams crossing the region cut through the blanket to the underlying carbonate rock. This starts the ground-water circulation, the water percolating downward into joints in the carbonate rocks in the interstream areas. The water forms a zone of saturation in which all joints and openings are filled with water to a certain level which represents the water table. Lateral

movement along vertical joints, bedding-plane joints, and other openings to the surface stream or other outlets begins the development of solution openings that result in sinks. Where the cover is removed locally from the carbonate rocks after karstification begins, the carbonate rocks commonly become casehardened and resistant to erosion. This may result in resistant carbonate ridges or ramparts along streams as described by Monroe (1967).

Also, as the sinkholes develop, the cover in the areas between sinks may be removed, permitting casehardening of the carbonate rocks at the surface. These casehardened or indurated surfaces where water can run off quickly are more resistant to erosion than the intervening sinkholes where solution was initiated under a soil cover. This differential erosion is the beginning of the hills in karst areas, known by different names such as mogotes, towers, cones, needles, and hums. The thickness of the soil cover probably also determines the shapes of the hills. Where the entire area of the surface of the hill is casehardened, the hill has the shape of a tower. Where the cover is thin and only the center of the original surface of the hill is casehardened, the hill is cone shaped or needle shaped.

Sinks in carbonate terranes have many shapes and sizes. The coalescence of sinks to form long dry valleys, or uvalas, continues to represent the dominance of solutional erosion over physical erosion. Larger valleys that are flat, soil-covered, and bordered by bare carbonate hills are called poljes (3 of Figure 1). It is not important to this discussion to debate whether poljes were initiated by graben faulting or by differential erosion. Examples of poljes initiated by grabens in Yugoslavia (Lobeck, 1939, p. 133) and Jamaica have been noted (Zans, 1951, p. 267), but the fact remains that differential solutional erosion is one of the predominant causes of their topographic form in full development (Birot, 1954, p. 176). Where bare carbonate mountains border a polje, the mountains are not prone to active physical erosion and only slightly more so to solutional erosion even where precipitation is great and even where infiltration through rock fractures is good; in contrast, solutional activity is likely to be great beneath the polje itself.

Having the same relation as regards erosive resistance to the denuded ridges bordering poljes are the conical denuded hills rising above a nearly flat floor in parts of Cuba, Puerto Rico, and North Vietnam (4 of Figure 1). These hills, referred to as mogotes, haystacks, and pepino hills, are high-rising remnants of limestone, now made resistant by their ability to shed water both externally and internally; there is very little carbon dioxide in the water to dissolve the limestone of the hills, and there is very little soil to remove. The plain at the base of the mogotes has developed by some combination of base-leveling conditions that have retarded erosion below the plain and have retarded the lowering of the water table beneath the plain.

Differential erosion is especially common where carbonate

rocks occur in broad belts adjacent to noncarbonate rocks. A classic example in which carbonate rocks are poorly resistant is the Ridge and Valley Province of the southeastern Appalachian region. In this region carbonate rocks are common in the valleys along the strike of the folds whereas the intervening ridges and mountains are commonly sandstones (5 of Figure 1). In this case the differences in topography result from a combination of differential physical and solutional erosion. The sandstones are not subject to much solutional erosion and they do not weather readily to form soil necessary for physical erosion to be strong. The carbonate rocks are entrenched by perennial streams that carry away the detritus from weathering of impure carbonates and of shales that are commonly present. The streams also represent discharge zones for circulating ground water in the carbonates, thereby aiding solutional erosion. In this setting both physical and solutional erosion are favorable in the valley carbonates. The importance of such a stream is demonstrated in the Sequatchie Valley of East Tennessee where the Sequatchie River, flowing southwestward into the Tennessee River, is entrenched a length of 75 miles into Paleozoic limestone and continues to remove detritus and dissolved mineral matter from the valley (Milici, 1967).

Valley and ridge topography in folded regions consists of ridge-forming carbonates and valley-forming shales in many cases (6 of Figure 1). In such cases the carbonates may be so nearly pure that soil does not form or earlier soil may have been stripped away as fast as it could form, thereby reducing further physical erosion. Once denuded of soil, solutional erosion is also lessened, chiefly because of limited contact with water high in carbon dioxide; the rocks thus become more resistant than the shales of the valleys. Resistant carbonate ridges are especially common in semiarid regions, but they are not uncommon in humid regions.

CARBONATE ROCKS IN CUESTAS

Cuestas are common where gently dipping resistant and nonresistant formations alternate. Because carbonate rocks are so readily accentuated by differential erosion, they are common in the rock sequences forming cuestas (7 and 8 of Figure 1). Excluding escarpments caused by faulting and considering only cuesta-type topography formed by erosion, one can cite three requirements for their development:

1. Sufficient regional uplift for topographic relief to develop
2. Gently tilted beds or, less commonly, certain broad structures with steep dips such as some breached anticlines
3. Alternating resistant and nonresistant formations

In contrast to the nonresistant carbonate rocks that underlie cap rocks of some escarpments are carbonate rocks of similar lithology that form resistant cap rocks of other escarpments. This contrast in differential

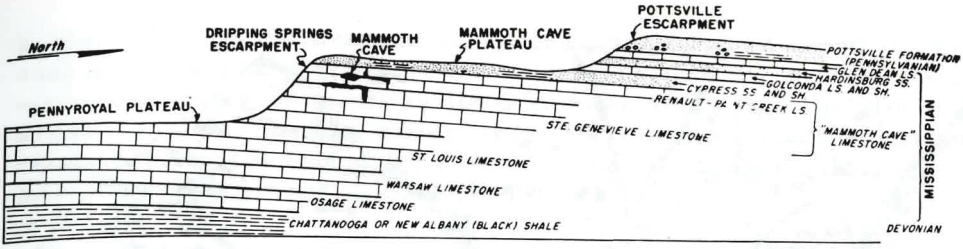

Figure 2. A north-south section through the geologic formations in the vicinity of Mammoth Cave would pass through the rocks shown above. In the soft, soluble limestone lowlands, caves and sinkholes are developed. The highlands above the Pennyroyal Plateau and separated from it by the Dripping Springs Escarpment are protected by sandstones from rapid erosion (after Livesay, 1953).

erosion of similar carbonate rocks invites an explanation. Where carbonate rocks lie in the cuesta sequence they may lie beneath a few feet to as much as several tens of feet of sandstone or other resistant cap rock, the carbonate rocks cropping out on the escarpment face.

Where carbonate rocks are nonresistant and do not cap the escarpment they may extend outward to make the surface of the lower lying plain, as is the case of the Nashville Basin in Tennessee and the Pennyroyal Plain of Kentucky (Figure 2). While the overlying insoluble formation has been subjected only to physical erosion, the subjacent carbonate rocks of the lowland plain have been subjected to both physical and solutional erosion. The favorable condition of a soil cover for solutional erosion, if not provided by the insoluble residue of the carbonate rocks may be provided by the down-slumping overlying insoluble rocks that once overlay the carbonates on the lowland plain. A typical example is the mixture of insoluble residue of Tertiary limestone and clays of the overlying Hawthorn Formation in the Dougherty Plain of southwest Georgia; the Dougherty Plain represents the low solution plain lying below a dissected escarpment and the higher-lying Tifton Upland where the Hawthorn Formation is intact (Figure 3). If soil is stripped from the lowland plain and if permeability of the low-lying carbonate rocks is poor, as in the Nashville Basin, physical erosion is slight and solutional erosion may be confined to a relatively shallow zone paralleling the rock surface. A soil-covered lowland plain, such as the Dougherty Plain along the Flint River in Georgia (Herrick and LeGrand, 1964), would encourage solutional erosion along the top of bedrock as long as ground water can discharge into a nearby stream. The tendency for surface streams to go underground in karst areas and to be widely spaced results in relatively slow movement of detritus except near the perennial streams. Thus physical erosion is relatively slight. Differential solutional erosion is predominant in the escarpment region. The retreat of escarpments is slow, especially if perennial

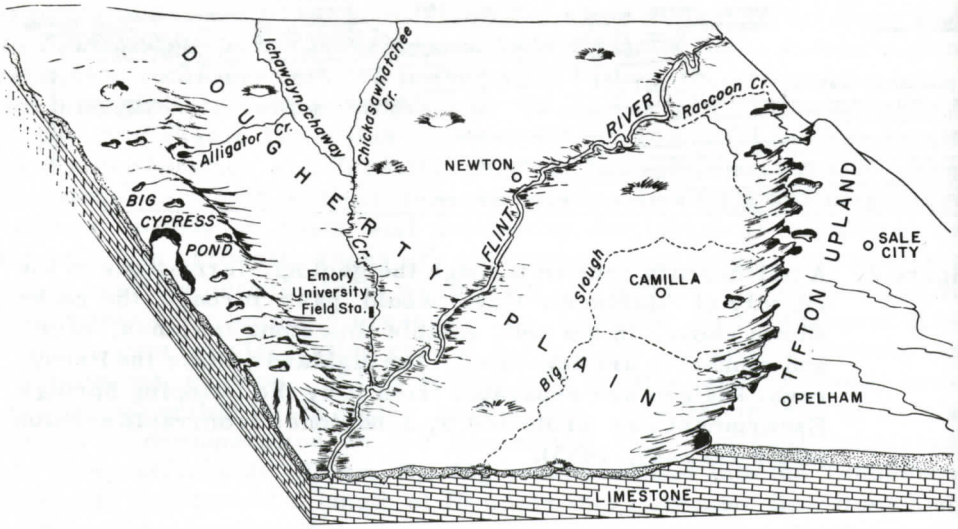

Figure 3. Diagrammatic sketch of cuesta topography in southwest Georgia, showing extent of solutional erosion (after Hendricks and Goodwin, 1952).

streams are sparse and if the overlying caprock is as much as a few hundred feet thick.

Examples of carbonate rocks capping cuestas are common where the rocks have been arched upward and have been eroded. Erosion on the structural dome of the Black Hills in South Dakota has resulted in a limestone formation being more resistant than underlying granites and metamorphic rocks; a prominent limestone cuesta forms an ellipse around a central mass of generally lower-lying Precambrian crystalline rocks (Darton, 1901). The Niagara limestone and dolomite of Silurian age forms a broad arcuate escarpment extending from New York into the Northern Peninsula of Michigan and downward into Wisconsin; throughout much of its outcrop area it stands appreciably higher at the escarpment top than do other rocks in the sedimentary sequence. Although glacial erosion has modified the topography of the Niagara cuesta, this topography was initiated by a combination of differential physical and solutional erosion. It is interesting to note that the degree of hardness of carbonate rocks is not necessarily a factor in resistance to erosion. For example, the London Chalk of southeast England is not ordinarily a hard rock, but it makes prominent escarpments. Trueman (1938, p. 61) described the geography of the region and emphasized that owing to its porosity and scarcity of surface streams, the Chalk "escapes most of the effects of river erosion and its surface is not lowered to anything like the same extent as the clays which lie along its borders: the Chalk thus stands out in ridges as surely as if it were a

harder rock." In the Western Desert of Egypt both Cretaceous and Eocene limestones have extensive limestone escarpments, the limestones exposed on the escarpments having been more resistant to erosion than adjacent shales. Several major reentrants into the escarpments are topographically low enough for oases to form. It is thought that escarpments in Egypt formed earlier when the climate was more humid than at present.

Two or more belts of cuestas may occur where the stratigraphic sequence is represented by two or more relatively insoluble formations separated by more carbonate formations and where the region has been uplifted (Stringfield and LeGrand, 1969, p. 358). One example is the Ozark Plateau Province of Missouri where Mississippian limestones of the Springfield Plateau lie below the Boston Mountain Escarpment capped by resistant Pennsylvanian sandstone; the Springfield Plateau lies above the Eureka Springs Escarpment, capped by resistant cherty limestone and the lower lying (stratigraphically) Salem Plateau, on which older Paleozoic limestones occur (Thornbury, 1965, p. 267-270). Another example is in central Tennessee where the Nashville Basin extends eastward and higher topographically and stratigraphically to the Eastern Highland Rim Plateau, east of which is the higher Cumberland Plateau; the escarpments between them are badly dissected, and limestone is widespread on the Nashville Basin and Highland Rim.

Earlier workers in the Ozark Plateau (Bretz, 1965) and in the central Tennessee area (Hayes, 1899) considered three distinctive periods of uplift and three cycles of erosion to account for the step-like plains in these regions. The delicate balances and imbalances of physical and solutional erosion are adequate to explain these conditions without the requirement of several erosional cycles (Stringfield and LeGrand, 1969). This latter view is in accord with studies by Strahler (1944) in the Kaibab Plateau of Arizona, by Meisler (1963) in the folded Paleozoic carbonate rock terrane of Pennsylvania, and by Hack (1966) in the Highland Rim region of Tennessee.

Two base levels of solutional erosion appear to be necessary for the development of two carbonate plains of different elevations separated by a less soluble formation at an escarpment. A perennial stream is or has been the base level for the lower lying carbonate plain, and the less soluble formation near the top of the escarpment is the base level of the upper carbonate plain. On the lower carbonate rocks, the resultant of physical and solutional erosion has been greater than that on the less soluble rocks of the escarpment; the sandstones or other rocks composing the escarpment are less prone to solution erosion, of course, and they are not likely to be physically eroded to a great extent because no perennial streams occur near the foot of the escarpment to take away rock materials. Erosion on the upper and lower carbonate rocks is concurrent but not necessarily even as to intensity or state of development. As the base level of the upper carbonate formation is the underlying less soluble formation and not the same base

level as the lower carbonate formations, two periods of uplift to account for the two plains is unnecessary.

Under the requirements of cuesta development mentioned earlier, any bed on which cover is removed and is not easily regenerated is likely to be the cap rock of an escarpment. Without the cover, the exposed part of the bed is resistant to both physical and solutional erosion in relation to adjacent rocks in the sedimentary sequence where some cover is present. The difference between poorly resistant and strongly resistant carbonate rocks can be understood if the significance of a cover is recognized. The same bed may be resistant to erosion at one place and nonresistant at another place, depending on cover and relative resistance of adjacent beds. Reconstruction of the erosional history of a scarp-forming bed of carbonate rock shows a breaching of the sequence of beds up dip and at a higher topographic position at an earlier time. The uppermost and earliest breaching of the scarp-forming bed was almost surely made where cover was present and the bed was locally weak. Structures on which earliest breaching occurred are commonly (1) an anticlinal arch or (2) a simple belted coastal plain. In both cases the earlier cover could be composed either of insoluble residue from the carbonate rocks or from overlying sedimentary material.

SUMMARY AND CONCLUSIONS

A contrast in topographic relief is characteristic of carbonate rocks. The relief may be local and small as expressed by many shallow sinkholes or large as expressed by poljes. In addition to relief within a karst terrane, significant relief commonly occurs where a belt of carbonate rocks is adjacent to a belt of noncarbonate rocks. Cuesta topography occurs where the adjacent rocks are only slightly inclined; carbonate rocks lie under an insoluble caprock and crop out near the base of escarpments in some humid regions, but carbonate rocks are the more resistant cappings of many cuestas. Folded rocks show long carbonate valleys between sandstone ridges in some humid regions, but ridge-forming carbonates between shale valleys are also common.

Differential erosion results from a combination of physical and chemical processes. Each component needs to be considered in relation to other components. Physical erosion may be considered in terms of decomposition of rocks into detritus, the movement of detritus on slopes and into the river bed, and the erosion of the river bed. The chemical or solutional erosion may be considered in terms of factors favorable for rock to dissolve and for the dissolved material to be carried away. "Preparation for erosion" is a prerequisite for both physical and solutional erosion. Decomposition or disintegration of rock is essential for physical erosion, and a permeable soil cover to allow slow infiltration of water containing carbon dioxide aids solutional erosion;

without a soil cover very little carbon dioxide (that is necessary for solution) is generated.

Consideration of soil cover leads to the following significant principle. Erosion in carbonate-rock terranes is favorable under moderate rather than under extreme conditions of cover, purity of the carbonate rock, topographic relief, and precipitation. For example, denuded carbonate rocks are much more resistant to physical and chemical erosion than are carbonate rocks with a moderately thin soil cover; where the soil cover is very thick, physical erosion of the covered bed is impossible and chemical erosion is retarded because of retarded water circulation. The extreme limits of the conditions of purity of carbonate rocks, relief, and precipitation, which tend to be unfavorable for erosion, are related to extremes in cover. For example, pure carbonates yield no insoluble residue to form a cover; also, intense precipitation tends to strip off a thin cover and to keep the rock denuded, especially on steep slopes. Thus, the degree of cover on a carbonate terrane is an important key to differential erosion and to much of the topographic relief. After physical erosion of most of a thick cover such as resistant and insoluble Pennsylvanian formations on the Nashville Dome and the Ozark uplift that have almost exposed underlying carbonat es, chemical erosion becomes important; in such a case, lowering of the surface of the carbonates is more rapid than that of the resistant beds forming the escarpment. There is a tendency for denuded upland rocks to stay denuded, and there is a tendency for small amounts of soil to move into adjacent low areas to aid solution there. This combination of tendencies causes topographic relief to increase until the low areas have reached a karst base level.

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LITHOLOGY OF ROCKS DREDGED FROM THE BLAKE PLATEAU

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ABSTRACT

Dredge samples and photographs from the Blake Plateau, off the southeastern coast of the United States, show that the bottom material consists of a sediment component and a rock component. Sediment is characteristically very well sorted foraminiferal sand except near the Bahama Banks where it is carbonate lutite. The rock fraction can conveniently be separated into limestone, coral, phosphate concretions and manganese concretions. Dredged limestone consists of reef rocks from along the edge of the continental shelf and indurated foraminiferal calcarenite from the Plateau that forms the bed rock in the area. It contains fossils ranging from Upper Cretaceous to Recent in age, but much of the dredged rock appears to be multicyclic. Ahermatypic coral grows in prominent topographic mounds or coral banks up to 20 meters in height. Coral is most abundant beneath the axis of the Gulf Stream and owes its growth to a long-term equilibrium condition between sedimentation and erosion maintained by the currents. The coral helps to stabilize the bottom by trapping fine sediment. Phosphate nodules occur extensively from the slope to the outer escarpment across the northern end of the Plateau, and in a small area off Miami, Florida. They are mostly conglomeratic nodules but near the outer escarpment appear to be an integral part of sedimentary formations. The marine phosphate has many petrographic and stratigraphic similarities to land deposits of the Carolinas and Florida: on the Blake Plateau it has accumulated as a lag gravel after erosion of the parent formations by the Gulf Stream. Manganese concretions are forming under present day conditions. They occur as concentric, roundish nodules up to 10 cm in diameter, as tabular plates and as a continuous pavement in combination with phosphate nodules. The manganese concretions grow by replacement of phosphate nodules as well as by simple accretion. Their overall chemical composition is similar to that of deep-sea deposits except that they reflect the high carbonate content of the Blake Plateau sediment. Various vertebrate fossils, granitic erratics and human artifacts were also picked from the samples. The Gulf Stream is the controlling environmental factor in the area and the distribution of all rocks and most fine sediment can be related to the current regime.

INTRODUCTION

The Blake Plateau is located off the southeast coast of the United States between the continental shelf and the Blake escarpment, which extends north-south from Cape Hatteras to the outermost Bahama Islands. The area is unique because the water depth of 400 to 1100 meters is intermediate between that of the continental shelf and the deep-ocean basin to the east (Pratt and Heezen, 1964). The major environmental factor in the area is the Gulf Stream which flows north through the Straits of Florida and across the Blake Plateau with a volume of 20 to 40 million cubic meters/second. The Gulf Stream has essentially shaped the Straits of Florida and controls the depth to the bottom on the Blake Plateau by scouring. It generally maintains a balanced condition between erosion and deposition in the area (Pratt, 1966; Ewing and others, 1966). It is the purpose of this paper to describe some of the rocks from the sea floor of the Blake Plateau and to interpret their origin in relation to the Gulf Stream environment.

Initial dredging and surveying on the Blake Plateau was begun in 1881 by the Coast Survey Steamer "Blake". The sediment samples were described by Bartlett (1883) and are still noted on the Coast and Geodetic Survey charts (#1001, 1002). Manganese and phosphate nodules dredged at the time were described by Sir John Murray (Murray, 1885) and were pictured by Alexander Agassiz (1888). No further sampling was done until 1947 and 1958 when cores and a few dredge hauls were taken by ships from Lamont Geological Observatory and Woods Hole Oceanographic Institution during the course of seismic refraction shooting in the area (Ericson and others, 1952; Hersey and others, 1959; Ewing and others, 1966; Sheridan and others, 1966).

The present compilation of data is from work that the writer was directly involved in between 1960 and 1969 (Figure 1). The sampling has been coordinated with a program of bottom photography, seismic profiling (Stetson, 1961 and 1969; Uchupi, 1967), suspended sediment sampling (Manheim and others, 1970) and current measurements and hydrography (Pratt, 1963 and 1966). The writer gained additional insight during a descent to the bottom of the Blake Plateau aboard the deep submersible research vehicle "Alvin" (Milliman and others, 1967). The various separate observations have been combined as data points to compile overall sediment and rock maps of the sea floor (Figures 2 and 3).

Acknowledgments

The present compilation of data and ideas is the culmination of nearly eight years of work involving seven sampling cruises to the Blake Plateau. The writer is indebted to the officers and crew of the R/V Gosnold and the R/V Atlantis of the Woods Hole Oceanographic Institution and the R/V Eastward of the Duke University Marine

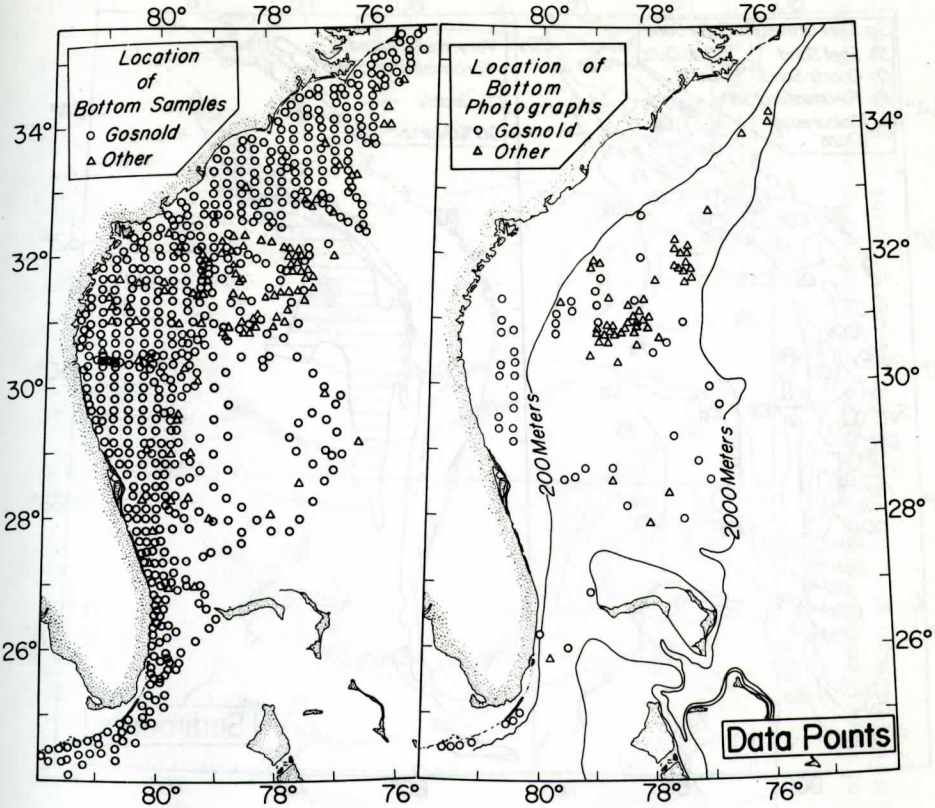

Figure 1. Location map showing the general location of the Blake Plateau and the position of bottom samples and bottom photographs used in the study.

Laboratory for the success of the dredging. All members of the various scientific parties participated in the observational phase of the work. Thomas Stetson was chief scientist on Cruise Atlantis-266 in June and July of 1961. The writer was chief scientist or principal investigator on all Gosnold and Eastward cruises referred to in the study.

Financial support for the Atlantis was through the Navy's Bureau of Ships Contract NOBSR-72521, and the Gosnold was supported by Office of Naval Research Contract ONR-2196(00) with Woods Hole. The cooperative program with Duke University Marine Laboratory, and use of the R/V Eastward on Cruises E-9-69 and E-36-69 is supported through the National Science Foundation Grant GB-8189. General support of laboratory work was through the joint Woods Hole Oceanographic Institution - U. S. Geological Survey cooperative continental shelf program headed by K. O. Emery at Woods Hole and supported through Contract 11508. The writer is particularly indebted to his colleagues

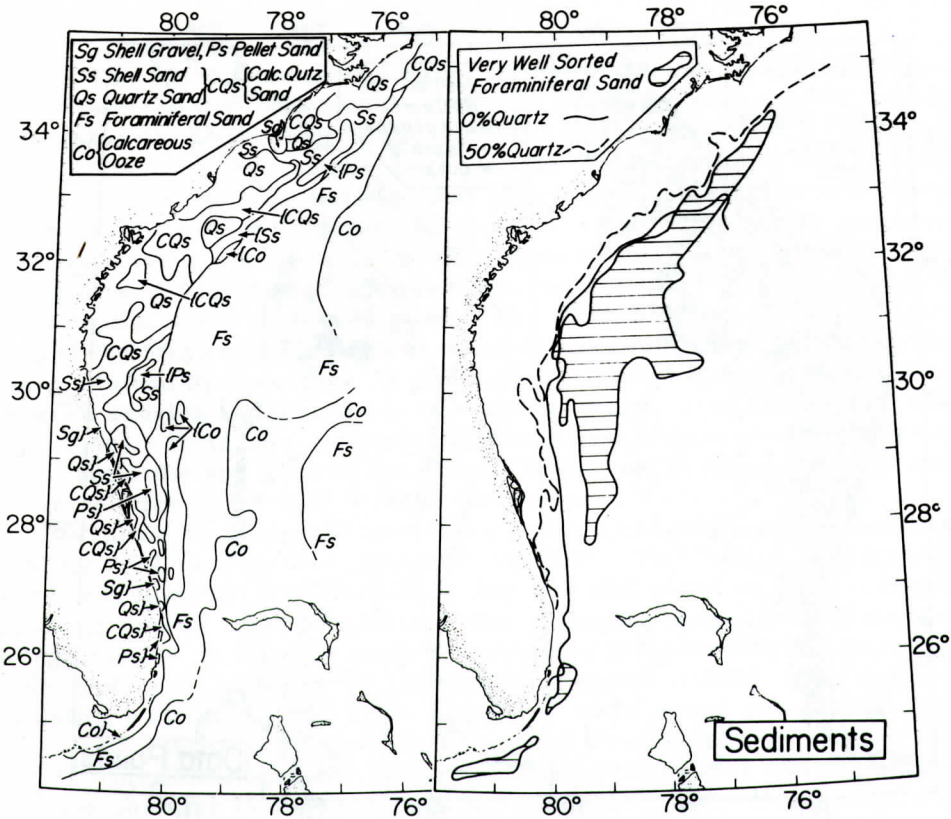

Figure 2. Sediment distribution on the Blake Plateau and adjacent continental shelf. The quartz boundary roughly follows the Florida-Hatteras slope and clearly separates the terrigenous shelf sediments from the biogenic carbonate sediment on the plateau.

at Woods Hole, John Hathaway, Peter McFarlin, Frank Manheim, Elazer Uchupi, John Milliman, Thomas Stetson, K. O. Emery and Edward Zarudzki for thoughts and discussions which made the compilation possible. Peter McFarlin ran the x-ray diffraction patterns. Frank Manheim is working on the geochemistry of the samples (Manheim and Pratt, in preparation; Manheim, Pratt and McFarlin, 1967). This paper is contribution number 2497 from the Woods Hole Oceanographic Institution.

SEDIMENTS

Sediment on the Blake Plateau is from two sources: (1) detrital

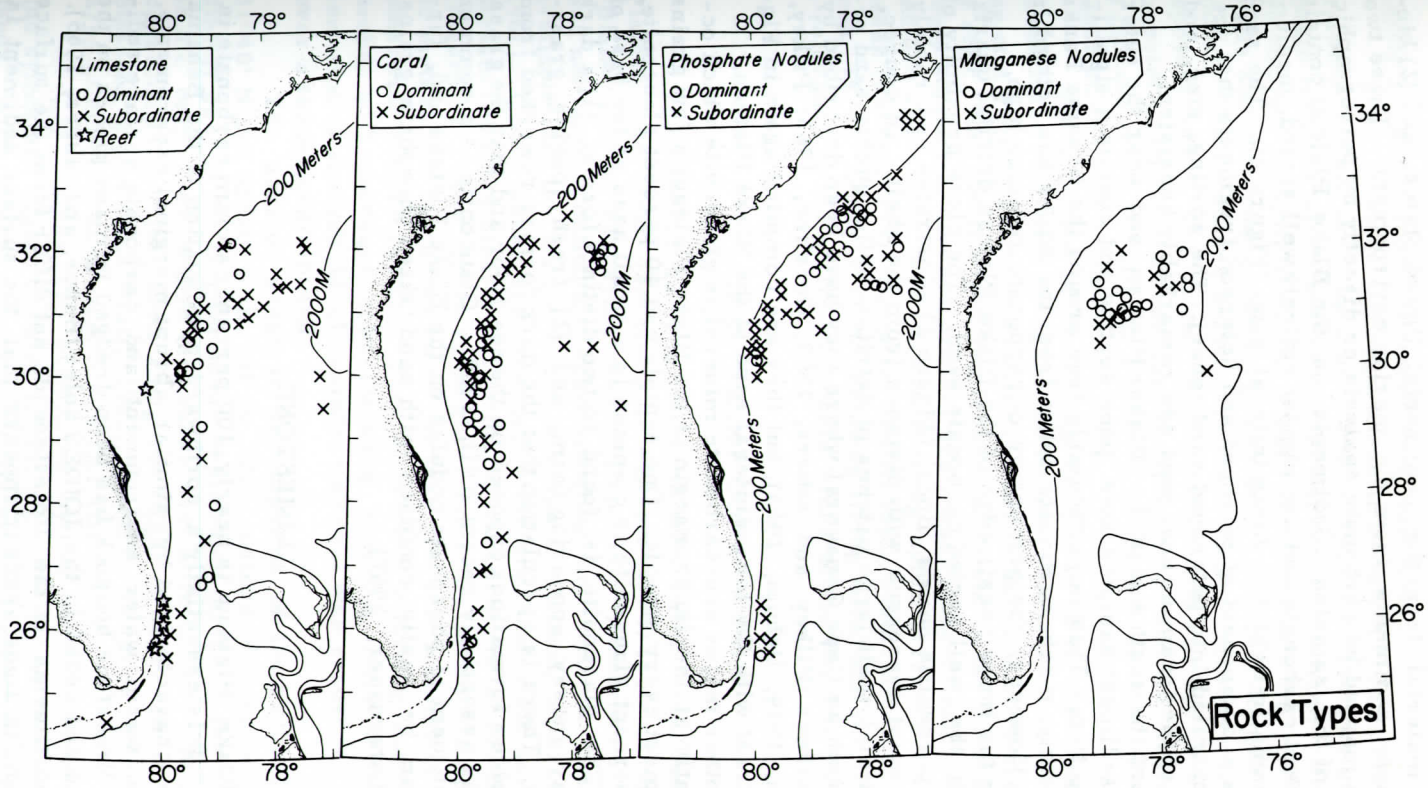

Figure 3. Distribution of the four basic types of rock dredged from the Blake Plateau. Dominant and subordinate are used in a very subjective sense to give an indication of the amount of material in the dredge samples.

terrigenous material from the continental United States, and (2) biogenic carbonate sediments from the marine environment. These two types can be mapped by carbonate analysis or directly by petrographic examination of the samples. Sediments on the Blake Plateau contain 98 to 99 percent carbonate and are almost entirely well sorted, medium grained, cream- to light-tan foraminiferal sand (Figure 2). The sand also contains a component of petropod and heterpod fragments that are locally concentrated into peteropod sand (ooze). The sorting, size, and composition of the foraminiferal sand are remarkably consistent.

Toward the south end of the Blake Plateau and near the outer escarpment sediment samples have poor sorting and contain a significant clay fraction. This is particularly true around the Bahama Banks where apparently much carbonate lutite from the Banks has settled in deep water. However, the percentage of carbonate remains high.

Along the inner (west) edge of the Blake Plateau detrital quartz is mixed with the well sorted carbonate sand. The close proximity of the 50 and 5 percentile quartz lines (Figure 2) illustrates how rapidly the quartz content decreases with distance from the shelf. Of course, on the continental shelf large patches of detrital, relict quartz sand occur as far south as Cape Canaveral where they have been described by numerous workers (Pilkey and others, 1969; Gorsline, 1963; Pilkey, 1964; Stetson, 1939, Uchupi, 1963); but they are compiled again in Figure 2 to contrast with the foraminiferal sand of the Blake Plateau.

The other major non-carbonate mineral is glauconite which occurs abundantly at the inner margin of the Blake Plateau; a few grains appear in almost every sample. One sample had 40 percent glauconite, and 5 to 10 percent glauconite is common in a small area at the base of the slope. The glauconite is found in two distinct forms: (1) a dark greenish-gray mossy appearing form, and (2) fresh looking, grass-green grains. There is speculation that the dark form is reworked from Miocene or older formations exposed on the sea floor and that the grass-green form is essentially recent (Dill, 1968). Glauconite is a common detrital constituent of phosphate nodules on the Blake Plateau and, furthermore, can be easily confused with sand sized phosphate grains (Pilkey and Luternauer, 1967).

LIMESTONE

The Blake Plateau is nearly 100 percent calcium carbonate in composition. It is essentially a northward continuation of the Bahama Banks and is characterized by similar abrupt marginal escarpments, deep canyons, warm water environment and carbonate surface sediments. The limestone bedrock has been dredged in many places on the Blake Plateau and cored in the JOIDES holes (Bunce, and others, 1965). The basic constituents of the limestone do not differ from the surface sediments, and in some areas it appears that the surface sediment is

mostly reworked from the underlying limestone.

The limestone is cream to tan to brown in color, fine to medium grained, fair to well indurated and composed entirely of organic carbonate remains with intergranular carbonate cement. Many samples are entirely foraminiferal but many include coral stems, small mollusc shells and pteropods. One large chunk is composed dominantly of pteropods. The limestone becomes stained with manganese oxide with long exposure on the sea floor. Recrystallization was not observed in the limestone samples but submarine cementation and lithification obviously has occurred (Milliman, 1966). Calcarenite is the most applicable descriptive name term for the rock.

Another type of limestone occurs in 80 to 100 meters of water in scattered localities along the shelf edge. This is the "reef" rock consisting of a variety of shallow-water carbonate facies and is probably mostly relic from the lowered sea level conditions of the Pleistocene. The reefs support a diverse present day fauna (Menzies and others, 1966), and are associated with distinct topographic terraces along the shelf break (Uchupi, 1967; Zarudzki and Uchupi, 1968; MacIntyre and Milliman, 1970). South of Cape Kennedy living coral and molluscs indicate a relevance to the active carbonate banks of the Florida Keys.

DEEP WATER CORAL BANKS

The widespread occurrence of deep-water coral on the Blake Plateau was first described by Agassiz (1888) from dredgings of the "Blake" and the Fisheries research steamer "Albatross". However, it was not until continuous echo-sounding and seismic profiles became available that the banks could be mapped (Stetson, Squires and Pratt, 1962). Coral banks are confined to the area under the axis of the Gulf Stream from the Straits of Florida to the north end of the Blake Plateau (Figure 3). They range in height from a few meters to approximately 30 meters and occur as more or less discrete mounds of bands on an otherwise smooth bottom. Several series of dredge hauls over single banks, with relevant echo-sounding records, have demonstrated the isolation of the individual mounds; no coral was recovered until we were over a topographic mound.

The area of greatest development of the banks lies between 31°40' N and 32°10' N in 800 to 1000 meters of water and lies immediately east of the largest scour depression on the Blake Plateau. This is the only area where the banks are more or less continuous. Low coral banks are localized along the edges of depressions (Pratt, 1966; Stetson, Squires and Pratt, 1962), possibly because they must have an area of exposed hard rock for initial attachment.

Coral banks are dominated by the two species of ahermatypic (no symbiotic algae) corals, Lophilia prolifera and Dendrophyllia profunda, with minor occurrences of several other species (Stetson,

Squires and Pratt, 1962). The general factors for their optimum growth are probably similar in many ways to shallow water coral reefs. The fact that the coral needs a firm rock substrate on which to initiate growth suggests that the coral outlines the area where the Gulf Stream has scoured to bedrock. On the other hand, the area with indication of maximum current action (31-32° N and 78° W) is covered with manganese pavement and supports no coral.

Coral growth seems to have played a role in trapping and stabilizing sediment on the bottom and, conversely, the trapped sediment adds mass to the banks enabling the coral to continue its upward growth. Dredging confirms that living coral is mostly confined to the crest of the banks. There is a suggestion from the appearance of the banks on echo-sounding profiles from bottom photography and direct observations from deep submersibles (Milliman and others, 1967) that the banks steepen toward their apex. The recovery of large masses of bank material reveals the role that carbonate lutite plays in packing and stabilizing the coral framework into a tough, coherent mass of sediment; on the other hand, the sudden input of large amounts of sediment could bury and kill the coral. Bottom currents are an important factor in bank growth because they bring in food and enough sediment to stabilize the banks without allowing excess sedimentation.

PHOSPHATE NODULES

Phosphate nodules and phosphatic sediments occur extensively on the north end of the Blake Plateau and extend to the outer escarpment (Figure 3). They have also been reported from off the Florida Keys (Portalis Terrace) (Gorsline and Milligan, 1963), the Straits of Florida and elsewhere along the coast in near-shore sediments. Phosphatic sediments were described from the JOIDES drill holes (Bunce and others, 1965) and phosphatic sands have been described from the continental shelf and slope adjacent to the Blake Plateau (Peaver and Pilkey, 1966; Pilkey and Luternauer, 1967). There seems to be little doubt that phosphate from the Blake Plateau surface is genetically similar and stratigraphically about the same age as the well-known deposits from Florida, South Carolina and North Carolina. Evidence for the stratigraphic continuity and depositional similarity of the off shore and on shore deposits are: (1) their aerial continuity, (2) similar fossils and age, (3) similar lithology and mineralogy, (4) evidence of erosion of much of the Tertiary section from the Blake Plateau and (5) direct photographic and visual evidence of nodules weathering out of limestone on the bottom. Several dredge hauls indicate a concentration of nodules in scour depressions, and generally their occurrence is best described as a lag gravel resulting from erosion by the Gulf Stream.

The phosphate nodules appear as dark brown to tan, irregularly shaped lumps, characteristically smooth on the bottom and rough and

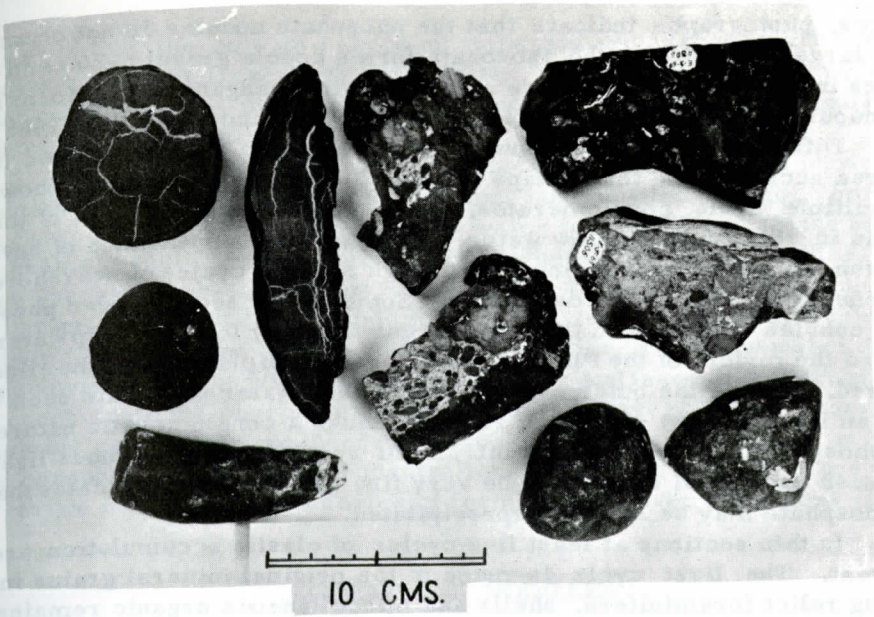

Figure 4. Various types of nodules and rocks from the Blake Plateau: upper left round concretionary manganese nodules; center, phosphate nodules partly replaced by manganese; upper right, phosphate nodules; right center, polished conglomeratic phosphate, lower right, two granitic erratic cobbles; lower left, end of a fossil rib bone, possibly from a dugong.

nodular on the upper surface (Figure 4 and 6). They are typically conglomeratic and unweathered. It is evident from their smooth bottom, and truncation of structures and fossils on upper surfaces, and their well-healed fractures that local accretion and recementation are common where they are not exposed to circulating sea water; but their exposed surfaces are slowly dissolving. Their structure emphasizes the well-known mobility of phosphate, and adds to the problems of determining the primary form of the phosphatic sediment. Cretaceous through Pliocene fossils, have been found both in dredged samples and on land. Sharks' teeth often protrude from the nodules because of differential solution. The nodules from the Blake Plateau are very similar to those described by Dietz, Emery and Shepard (1942) off southern California, where there is also some confusion about the age.

The largest nodule recovered weights 57.1 kilograms and measures 48 by 40 by 19 centimeters (Figure 6). It, and others nearly as large, come from an area next to the inner margin of the Blake off Wilmington, North Carolina in about 300 meters of water. Even though the largest nodules are about the maximum size recoverable with our

dredges, photographs indicate that the phosphate nodules do not occur much larger. Typically the phosphate forms cobble gravel 6-20 centimeters in size, however, where cemented with manganese they form a continuous manganese-phosphate pavement (Pratt and McFarlin, 1966).

Different kinds of phosphate nodules can be roughly mapped in the area according to their shape and size. The largest nodules showing multiple cycle conglomerates, come from the northern end of the plateau in relatively shallow water. They have the appearance of having been reworked and recemented through several cycles of nearshore, high energy erosion and deposition. Some of the second order phosphate cobbles cemented in them are larger than any found in deep water. Toward the middle of the Plateau nodules are often platy and rather fine textured. Toward the outer escarpment in deep water phosphate seems to be an integral part of the bed rock and lacks a conglomeratic nature. The phosphate nodules dredged off Miami were often homogenous light yellowish brown and seemed to be very fine grained which indicates that the phosphate may be largely reprecipitated.

In thin sections at least five cycles of clastic accumulation are apparent. The first cycle is made of the original mineral grains including relict foraminifera, shells and miscellaneous organic remains. Relict clastic grains of quartz and feldspar are common in nodules near the slope; glauconite is widespread, often very noticeable in hand specimens and follows closely the distribution in the modern surface sediments (see section on sediments). The progressively coarser phases are composed of cemented nodules and fossils from the previous cycle. Nodules often show layers of phosphate accretion and healed cracks. In several sections large cracks appear to be filled with recent carbonate lutitle containing small foraminifera. It is apparent that this cyclic multiplicity of phosphate could lead to much confusion about the age of the nodules. The degree of phosphatization also varies greatly, even with individual nodules.

The petrology of the phosphate concretions on the Blake Plateau can be simplified into a limestone-phosphate-manganese sequence. The basic rock type is indurated carbonate ooze or limestone. The matrix of the limestone is calcilutite, much of which is too fine to be resolved by optical means. Abundant relict carbonate structures in the nodules show that the limestone was phosphatized in place by direct replacement phenomina with no intermediate pellets or primary phosphate mineral structures. Several thin sections show parallel zones of phosphatization transecting the carbonate structures. Foraminiferal tests often remain pure calcite although the foramin and other intersticies are filled with a microcrystalline aggregate of apatite. Complete phosphatization was observed in a few thin sections but in many sections it is so nearly complete that only a few specks of relict carbonate (probably coccoliths) remain. A very few thin sections show euhedral crystals of apatite. Specific gravity measurements, indicies of refraction and X-ray diffraction analyses support the other optical

evidence that the only two abundant minerals are carbonoapatite (carbonate-apatite or francolite) and calcite. The poorly-crystalline carbonoapatite making up the phosphate rock of industry is commonly referred to as collophane.

Evidence for shallow water replacement-origin of the phosphate nodules includes, (1) shallow water and land fossils (dugong, horse, Inoceramus and pebbles with shallow water foraminifera), (2) the rounded pebble conglomerates that must have been formed in a moderately high energy, near-shore environment, (3) numerous unphosphatized relict grains of glauconite and quartz, and (4) at least four cycles of pebble formation and recementation. In addition, a shallow water estuarine environment provides a reasonable sink and favorable hydrography to concentrate the phosphate along the coast (Peaver, 1966). Possibly, anerobic conditions are needed to break the nutrient cycle and release the phosphate in sufficient concentrations to replace the calcareous mud. A review of the various theories of formation is given by Degens (1965).

MANGANESE NODULES

Manganese deposits occur in three principal modes on the north end of the Blake Plateau, (1) round nodules (2) flattened nodules, and (3) continuous pavement. Round nodules are found around the southern edge of the area (Figure 3). They average six to 10 cm in diameter, are more or less round and have concentric internal structure. I surmise that the 10 cm maximum size of round nodules may be the limit at which they can be "rolled" by the bottom currents. No nuclei and few inclusions of extraneous material were observed. Nodules larger than 10 cm are flattened and some are plate shaped. Very large irregular masses several meters across have been observed from deep submersibles (Hawkins, 1969; Milliman and others, 1967). Extreme development of the manganese concretions is the continuous manganese pavement occupying about 500 km^2 in the center of the nodule area (Pratt and McFarlin, 1966). Samples from the pavement shows that it is a mixture of accretionary manganese oxide covering and cementing phosphate pebbles and limestone, and is forming under present environmental conditions. X-ray diffraction and chemical analyses indicate that even the purest looking manganese concretion have a high phosphate content (Manheim and others, 1967).

The manganese concretions are mostly later than the phosphate nodules and consist of two phases of accumulation: (1) an accretionary and (2) replacement phase. Simplest is the accretionary manganese that forms the typical round nodules and some of the tabular masses. In hand specimens it is soft and soils the fingers. Accretionary textures are distinguished by the nearly complete lack of included detrital mineral matter or relic organic forms, and by a fine network of calcium carbonate exudation veinlets. The small veinlets characteristically are

parallel to the concentric banding of the nodule and show double terminations within the nodule. In thin sections the carbonate crystals appear clear and well-defined and some show equal growth from both walls and abutting terminations down the middle of the veinlet. The veinlets seem to be exudation structures formed by the explusion of excess calcium carbonate from the nodule and indicate a shrinkage and rearrangement of the constituents during the growth. They appear to be smaller toward the periphery. Occasional veinlets have voids that contain accicular crystals of aragonite (McFarlin, 1966); displaced calcium carbonate occurs in typical x-ray diffraction diagrams.

The opaque layers of the accretionary nodules are composed of iron-manganese oxides. X-ray diffraction and chemical analysis reveal that the only two metallic minerals involved are goethite and todorokite. However, the subdued nature of the x-ray peaks and the lack of any crystal definition by optic means suggest that much of the iron and manganese may be unordered and amorphous (limonite and wad) which is similar to pavement described from San Pablo Sea mount (Aumento, Lawrence and Plant, 1968). Under reflected light, occasional fine bands appear intense black, whereas most of the opaque material appears brownish-red. Microprobe analyses indicate that black areas have the highest manganese content and the rest of the brownish opaque material is proportionally high in iron. Aureols of orange limonite stain, especially at the surface, are an indication of some secondary weathering.

The second means of formation of manganese-iron oxide concretions is by replacement of pre-existing phosphate nodules. Replacement criteria include (1) thin apophyses and dendrites of opaque manganese oxide in the centers of phosphate nodules, (2) relict detrital minerals such as quartz and glauconite, (3) differential replacement of relict organic structures, and (4) the lack of carbonate exudation veinlets and concentric structures. In complete replacement, thin sections appear almost opaque and structureless under transmitted light except for relict grains of quartz and glauconite. Under reflected light, some of the opaque material high in manganese appears intense black. Reddish-brown limonitic staining is an indication of total lack of manganese.

The chemical and mineralogical composition of the manganese concretions reflects the carbonate environment of the Blake Plateau rather than a silicious, red clay province and the nodules are noticeably lacking in silica. Nearly all parts of the concretions are high in calcium except the detrital quartz and glauconite. The phosphate has replaced the calcite of the limestone and the iron and manganese oxides have in turn replaced the phosphate nodules. Average manganese values are greater in the accretionary manganese nodules than in replacement nodules and pavement but are still considerably less than values of deep-sea averages (Mero, 1965; Manheim and others, 1967). Trace elements, including Cu, Ni and Co, are a little less than deep ocean

Figure 5. Various vertebrate remains from the Blake Plateau: Top four, mesorostral bones from beaked whales; lower left, sharks teeth; lower center, vertebrate rib bones; lower right, fish spines; bottom center, ear bones of whales and porpoises.

averages (Manheim and others, 1967).

FOSSILS

Included in the dredge hauls from the Blake Plateau are numerous organic remains (Figure 5), most of the occurrences are definitely fossil and an integral part of phosphate nodules. However, a few bone fragments are unphosphatized and appear to be modern. Sharks' teeth in particular are often loose and are difficult to place stratigraphically.

The oldest fossils are a Late Campanian-Early Mastrachian (Late Cretaceous) fauna from the northern Blake Plateau off Charleston, South Carolina. At least five dredge hauls recovered macrofossils which include 29 probable species. Additional localities were probably overlooked because of the difficulty of working with fossils in the highly cemented phosphate. Diagnostic genera are the cephlopods Baculites, Neophylloceras and Eutrephoceras and three species of Inoceramus. Coral, sharks' teeth, additional pelecypods and gastropods all support the late Cretaceous age. The preservation of the forms and the type of original shell patch preservation (including aragonitic shell material) suggest that the fauna is stratigraphically in place and confirms the

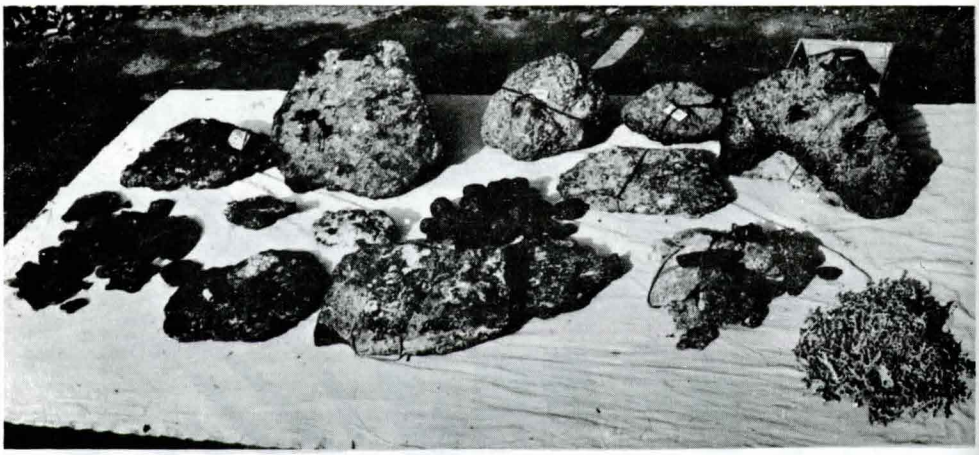

Figure 6. Assorted rocks from the Blake Plateau including coral (far right), large phosphate nodules (rear), rounded manganese nodules (center), and combination phosphate-manganese concretions.

presence of Upper Cretaceous strata on the sea floor. The collection is similar to Gulf and East Coast Coastal Plain assemblages in general character. All material is being retained by the U. S. National Museum for further description and reference (N. F. Sohl, E. G. Kauffman and R. M. Pratt, manuscript in preparation). The presence of Cretaceous limestone in the vicinity was confirmed independently on the basis of foraminifera from a piece of limestone recovered by Alvin Dive #201 (Milliman and others, 1967). The dive also clearly revealed phosphate nodules being eroded out of light colored limestone slabs and lying as a lag deposit on the bottom. The time of phosphatization is unknown but is probably different from the stratigraphic ages of the formations. The Cretaceous strata are flat-lying beds, more or less exposed over most of the northern end of the Blake Plateau, as revealed by seismic profiler and refraction studies (Hersey and others, 1959; Uchupi and Emery, 1967). Structurally they are warped by the seaward extension of the Cape Fear Arch.

Sharks' teeth were recovered from most of the phosphate localities. Mostly the teeth are imbedded in the conglomeratic nodules and are impossible to free from the matrix. The largest free tooth measured six cm in length. Teeth range in age from Cretaceous to modern, but are mostly nondiagnostic. There is a good possibility that loose teeth are multicyclic and are not indicative of the age of the bottom (Dunkle, Report on referred fossils, U. S. National Museum, 7-8-1966).

One dredge sample from off Miami (Gosnold 2461) contained a number of silicified rib bones that are believed to be from a fossil dugong (sea cow). Dugons appeared in Florida in Miocene time and are common as fossils in the phosphate deposits of central Florida (Frank

Whitmore, Jr., Report on referred fossils, U. S. National Museum 6-22-1966). Fragments of ribs were noted from several other dredge hauls. Another item of interest was the recovery of four mesorostra bones from the elongate snouts of beaked whales (family Ziphiidae, Figure 5). According to Frank Whitemore (Report on referred fossils, U. S. National Museum 10-16-1969) the bones could well belong to the modern genus (Ziphius, which has been tentatively identified from the lower Pliocene strata of Europe. Apparently, old living beaked whales develop mesorostral bones with a stony appearance so that the preservation and recovery of such bones does not necessarily mean petrifaction. Of additional note is the recovery of two tympanic bulla (ear bones) from beaked whales, a tympanic bulla from a porpoise (Delphinidae) and a tooth place from a porcupine fish (Diodon) (Frank Whitemore, Report on referred fossils, U. S. National Museum, 11-20-1969).

Human artifacts from deep water are always interesting because they illustrate a possible means of dispersing erratic geological material. Coal and cinders obviously dumped from passing ships were found in about half the samples. A ham tin was dredged from 800 meters off Miami and a German beer bottle from 1000 meters in one of the scour depressions at the west edge of the Blake Plateau.

Four dredge hauls recovered specimens of modern bone. One was sawed and obviously a steak bone dropped from a ship, the second was identified only as modern mammal and one dredge haul was half full of whale bones. In this case we must have dredge right over an old carcass on the bottom; but it is curious that the only rocks in the dredge haul were rather large granite and limestone erratics.

ERRATICS

The Blake Plateau is a northward continuation of the Bahama Banks carbonate reef complex. It was, therefore, surprising when granitic boulders and cobbles were found in 12 of the dredge hauls (Figure 4). One haul (E-36-69, Sta. 13037) had 47 erratics, another (Sta. 13045) had nine, but eight hauls had only one. The largest boulder measures 15 x 25 x 20 cm in size and they range down to 2 cm pebbles barely retained by the dredge. The shapes are angular to subrounded and are often flattened or elongate reflecting schistose structure of the rock. One small limestone pebble showed good striations but these may well have been caused by abrasion during mass wasting. The texture of the erratics as a whole does not have a glacial "look."

The petrography of the erratics was examined by slicing, for binocular microscopic examination, and 18 petrographic thin sections were made. Out of a total of 78 separate erratics 51 percent proved to be granite and granitic gneiss, 28 percent dense limestone and dolomite, 16.4 percent dark dike-and metamorphic-rock, 2.6 percent volcanic rocks, and 2.0 percent clastic sediment. Thin sections show that

most of the granitic rocks are true granite with as much as 75 percent orthoclase.

There is no question that the granitic rocks are erratic, that is, they have been transported into the area from an outside source. The question of their origin resolves into (1) ice rafting from the north during the Pleistocene, or (2) tree root rafting in the Gulf Stream and (3) ballasts dropped from ships. Ice rafting is unlikely because (1) the Gulf Stream flow was not appreciably altered during the Pleistocene as shown by the distribution of coral and sediments and general hydrography, (2) there is no evidence of cold water (glacial) faunas south of Cape Hatteras on the continental shelf, and (3) there is no evidence of erratics on the shelf. In fact, both the high proportion of fresh granite and the dominance of the Gulf Stream current off shore are strong arguments agains the southeastern states ever being a major contributor of the erratics. The varied size of the erratics with a dominance in the pebble range, their location away from reefs and harbors, and their very number make it improbable that they were all ballasts dropped from ships, although undeniably an occasional ballast rock might be included in the collection. The most reasonable source for the granitic rocks is the Guiana Shield of northeastern South America where more than half a million square kilometers of Precambrian and younger granite is exposed. Not only are the great South American rivers the best source of tree flotsm, but lithologic comparison of the erratics with the description of granitic rocks from the area is very convincing (Pratt, 1970).

Supporting evidence for the northward transport of erratic material by the Gulf Stream is the description of large mats of shallow water turtle grass floating north in the current (Menzies, Zaneveld and Pratt, 1966). The grass is derived from the Bahaman Banks and is particularly abundant after hurricanes. Two dredge samples from deep water contained worn fragments of shallow water (hermatypic) reef coral typical of Caribbean shores. The explanation for this is that it is thrown up on the beach, dried out and then washed back into the sea during storms where it floats north in the Gulf Stream until it becomes water logged and sinks (Kornicker and Squires, 1962). No large trees have been observed floating north in the Gulf Stream by the writer but coconuts, reeds and debris are common. Two apparent tree root masses were picked from bottom photographs in the area. A simple calculation shows that it would only take a few large tree loads of rock a year to substantially cover the area in Pleistocene time and the possibility exists that most of the debris enters the ocean during brief times of exceptional flooding.

SUMMARY

The dredged rocks from the Blake Plateau reflect the transporting and erosional power of the Gulf Stream. Direct observations with

bottom photographs, current meters, deep-submersibles and hydrographic instruments have demonstrated the northerly flow of the current on the bottom (Milliman and others, 1967; Pratt, 1963, 1966). The Gulf Stream acts as a barrier to the seaward distribution of quartzose sediment from the southeast United States and is a factor in controlling the quartz-carbonate boundary along the inner edge of the Blake Plateau. The bottom sediment on the Blake Plateau is very well sorted carbonate sand of pelagic origin that is being winnowed across the bottom. Even the indurated calcarenite bedrock in the area gives evidence of being multicyclic. Erosional depressions and direct observations show that the bedrock is readily eroded if the currents become too strong, but the general cover of loose sediment and partially filled depressions are indicative of stability and even lithification under the present day current regimen in the area.

Deep water coral banks are particularly good evidence supporting the concept of balance between erosion and deposition. It is apparent that excess sedimentation would quickly bury the coral banks but their absence from areas of bare rock surface suggests that a certain amount of trapped sediment is necessary to compact and stabilize the banks. Also, currents supply nutrient and oxygen and thereby control the growth and orientation of coral and other sessil fauna, just as in shallow water coral reefs.

Manganese and phosphate concretions form a rock pavement over much of the northern end of the Blake Plateau. Again, Gulf Stream currents are the decisive factor in the accumulation of the nodules, but for two entirely different reasons. The phosphate nodules are in the form of a lag gravel that often forms a continuous cover on the bottom. They are a mechanically formed lag deposit that is relict from former Tertiary strata, and are now undergoing solution where exposed to currents. The manganese is later than the phosphate and is actively replacing phosphate or accreting on it under present conditions. The rounding of the nodules is probably the result of movement and turning by bottom currents over long periods of time. The round nodules grade into large flat slabs and continuous pavement. The slow rate of manganese accumulation (.01 mm/ thousand years; Manheim and others, 1967) suggests a long period of stable environment, probably going back to the mid-Tertiary. In addition the same non-depositional condition allows the accumulation of granitic rock erratics and various odd fossils. The erratics are probably the result of tree root rafting from a source in northeastern South America and again show the dominance of the Gulf Stream.

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THE OFFSET COURSE OF THE ST. JOHNS RIVER, FLORIDA

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ABSTRACT

The northward-flowing St. Johns River is the dominant river in eastern peninsular Florida. At a point east of Sanford, Florida, the river turns toward the west to reach an older valley that it follows northward almost to Palatka before turning to the east to flow again over a younger terrain. The course of the river over the older, more westerly surface is the St. Johns River Offset. The offset valley probably marks part of the course of an ancestral river. Factors believed to be important in the development of the valley include faulting and fracturing, solution of carbonate sediments, and favorable structures for artesian discharge. The preservation of the offset valley as a major stream course is due primarily to the discharge of large quantities of artesian water. Much of this discharge is made possible in the northern part of the offset area by the relatively permeable nature of the aquiclude and in the central and southern areas of the offset by the exposure of the aquifer. The numerous springs from Lake George southward mark sites where the aquifer is exposed or breached.

INTRODUCTION

The major streams along the east coast of Florida flow parallel with the coast following former lagoons or swales between beach ridges. These are the coast-parallel streams discussed by White (1958, p. 11). The normal flow of these coast-parallel streams is to the north or to the south with occasional eastern turns or jogs as the streams cut from one consequent course to another consequent course closer to the coast. The easterly turns may be through ancient tidal inlets. In the case of the St. Johns River this normal trend of river flow is reversed at Lake Harney, east of Sanford, Florida (Figures 1 and 2). At Lake Harney the St. Johns River turns to the west and flows for 20 miles to reach a valley cut into an older, higher terrain. The river flows northward along this valley for approximately 75 miles to a point south of the town of Palatka. Here the river jogs back to the east to traverse an area of younger, lower surface over which it follows a generally northward

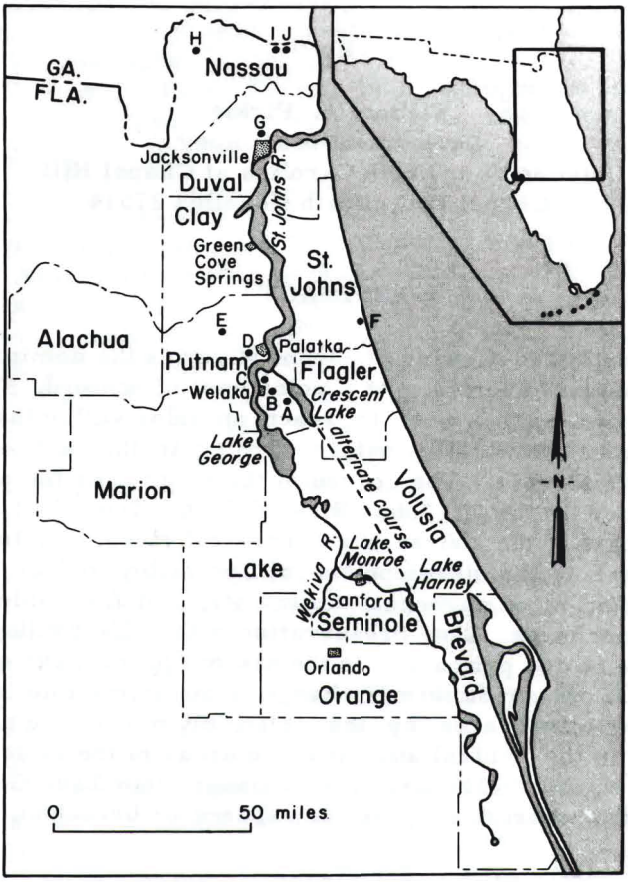

Figure 1. Locations of counties, major cities, lakes and rivers in the study area. Capital letters indicate sites of samples reported in Table 2.

direction with occasional eastern jogs until emptying into the Atlantic Ocean at Mayport in eastern Duval County. The part of the St. Johns course where the river flows on the older, more westerly surface has been named the St. Johns River Offset by White (in Puri and Vernon, 1964, p. 13 and Fig. 6). It is the origin of this offset course that constitutes the first objective of this study.

Since inception of the offset course there has always been the possibility that the river would readjust to a shorter alternate course along a younger, more easterly swale or former lagoon. In fact a course northward from Lake Harney through Crescent Lake would be a shorter route for the river to follow than its present route along the St.

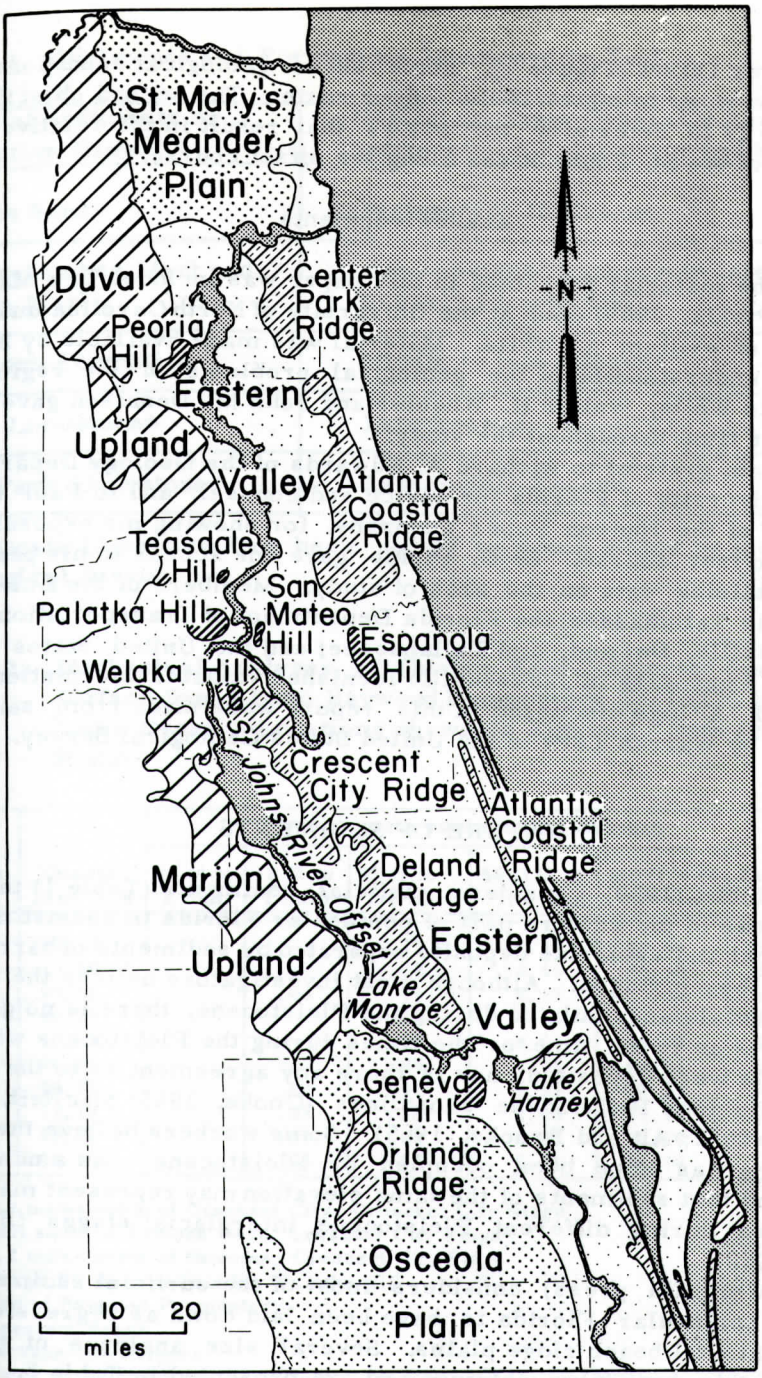

Figure 2. Landforms in the region of the St. Johns River. From White (in Puri and Vernon, 1964).

Johns River Offset (Figure 1). Some set of geological conditions has caused the river to maintain the longer route. The second objective of this study is to determine the factors that result in the relative permanence of the St. Johns River along the offset course.

Acknowledgments

The material presented in this paper was included in a Masters Thesis (Pirkle, 1969) done at the University of North Carolina under the guidance of William A. White. Material was made available by him to aid in an understanding of the geological problems of the region. In addition to White, Walter H. Wheeler and John M. Dennison gave freely of their time throughout the study.

I am grateful to Richard A. Edwards of the Geology Department of the University of Florida for advice and counsel, and to Paul White, geologist of the Hudson Paper Company, for showing me exposures along the offset course. E. C. Pirkle gave me access to his personal files containing data on the area of study. Members of the Bureau of Geology in Tallahassee; the Florida Department of Transportation staffs in Gainesville, DeLand, and Tallahassee; and the United States Army Corps of Engineers in Jacksonville furnished valuable information. The "LP" and "A" logs discussed in this report were made from samples obtained by Warren Leve of the United States Geological Survey.

NATURE OF SEDIMENTS

Cooke (1939) considers surficial sediments (Table 1) through large areas of the eastern part of peninsular Florida to consist mainly of lagoonal and estuarine deposits separated by sediments of barrier islands or beach ridges. Although most investigators believe the surficial materials accumulated during the Pleistocene, there is no general agreement as to the time or the times during the Pleistocene when the sediments were deposited, nor is there any agreement as to the elevations of various Pleistocene shorelines (Cooke, 1945; MacNeil, 1949; Vernon, 1951; Alt and Brooks, 1965). Some workers believe the same lagoonal areas have been occupied by Pleistocene seas a number of times and that sediments of the same elevation may represent materials laid down during different Pleistocene interglacial stages (Brooks, 1966, 1968).

White (in press) considers most of the surficial sediments in eastern peninsular Florida to have been laid down as regressional or progradational beach ridge plains. Several size analyses of surface sands of this area were obtained and are presented in Table 2. These analyses are similar to the analyses of a great number of sand samples obtained from the region. The consistent size distribution of the sand in these samples is compatible with a regressional or progradational

Table 1. Sediments and Formations Mentioned in Text.

Types of Sediments or Formation Name	Geological Time Generally Assigned to Sediments or Formations in Study Area
Surface Sands	Pleistocene to Recent
Caloosahatchee Marl and Nashua Marl	Ages assigned range from late Miocene to early Pleistocene
Hawthorn Formation	Middle to late Miocene
Ocala Limestone*	Late Eocene
Avon Park Limestone	Late middle Eocene

*The Ocala Limestone has been raised to group status by the Florida Geological Survey.

Table 2. Mechanical Analyses of Quartz Sand. Selected Samples of Surface Sands from Baywood Promontory Eastward to the Atlantic Ocean. Locations shown on Figure 1. Data from E. C. Pirkle.

Sample No.	Quartz Sand in %	Percent quartz sand retained on mesh					
		10 (4 to 2 mm)	18 (2 to 1 mm)	35 (1 to 1/2 mm)	60 (1/2 to 1/4 mm)	120 (1/4 to 1/8 mm)	230 (1/8 to 1/16 mm)
A	99.59			.11	7.24	85.59	6.66
B	99.60			.28	10.66	85.40	3.26
C	97.62		.04	.11	4.17	83.34	8.96
D	98.72			.07	4.83	83.11	10.70
E	97.83			.19	8.54	83.13	5.96
F	99.93			.03	2.93	92.84	4.14
G	98.77		.02	.17	5.55	89.17	3.88
H	97.79		.02	.28	2.90	85.10	9.49
I	99.01		.01	.08	2.05	82.70	14.17
J	98.15		.06	.17	5.01	86.76	6.15

A 4.2 miles north of Crescent City, Crescent City Ridge
B Just south of Pomona Park, Crescent City Ridge
C 1.2 miles north of Satsuma, Crescent City Ridge
D Poage Pit west of Palatka, Palatka Hill
E Top of Baywood Promontory
F Crescent Beach
G Jacksonville Airport
H Thick sands along Seaboard Coast Line Railroad north of Hilliard
I Reids Bluff along St. Marys River
J Roses Bluff along St. Marys River

beach ridge origin for the sediments.

The river channel along the offset course of the St. Johns River contains much sediment fill. The river must have cut deeper when sea level was lower. With rise of sea level the deepened valley was filled with fine sand, silt and clay. Included in this sedimentary fill are occasional scattered shell fragments and, locally, lenses with a high shell content. The shell lenses are the materials to which the name "Nashua Marl" was originally applied (Matson and Clapp, 1909).

DIASTROPHIC MOVEMENTS, FRACTURE ZONES, AND SOLUTION

Faulting and other movements associated with the Ocala uplift have played a more significant role in determining the course of the St. Johns River than previously has been realized. Fracturing and faulting allow a freer circulation of ground water with a corresponding increase in solution of carbonate rock. This solution would be especially significant during times when ground-water movement was downward along fractures.

Evidence from drill holes has caused investigators to report the presence of faults along nearly the entire course of the St. Johns River (Figure 3). Vernon (1951, Pl. 2) shows a fault parallel to the St. Johns River from its headwaters area to the southern border of Orange County. Lichtler et al. (1968, Fig. 6) in their report on Orange County present two east-west cross sections, one near the northern border of the county and the other near the southern border. In both sections a fault with its up-thrown block on the east is shown lifting the Ocala and Avon Park limestones upward in the area of the St. Johns River. In this uplift area the relatively impervious Hawthorn sediments that overlie the Eocene limestones are very thin, being no more than a few feet in thickness. In contrast, the thickness of Hawthorn sediments overlying the Eocene limestones immediately west of the fault varies from 100 to more than 300 feet.

Barraclough (1962) and Wyrick (1960) report an east-west graben along the northern part of Seminole County and the southern part of Volusia County. It is through this graben that the St. Johns River flows westward from the Eastern Valley to the offset valley.

Wyrick (1960, Fig. 4) shows a fault running parallel with the St. Johns River from Lake Monroe at the southern end of the offset valley to a series of lakes approximately 10 miles south of Lake George. This fault continues northward passing to the east of Lake George. Bermes et al. (1963, Fig. 9) report a fault west of Lake George. The St. Johns River follows the fault to Palatka where the river turns to the east to re-enter the Eastern Valley. Leve (1966) reports a fault running from northeastern Clay County to northern Duval County. The St. Johns River follows this fault until making its final eastward jog toward the ocean.

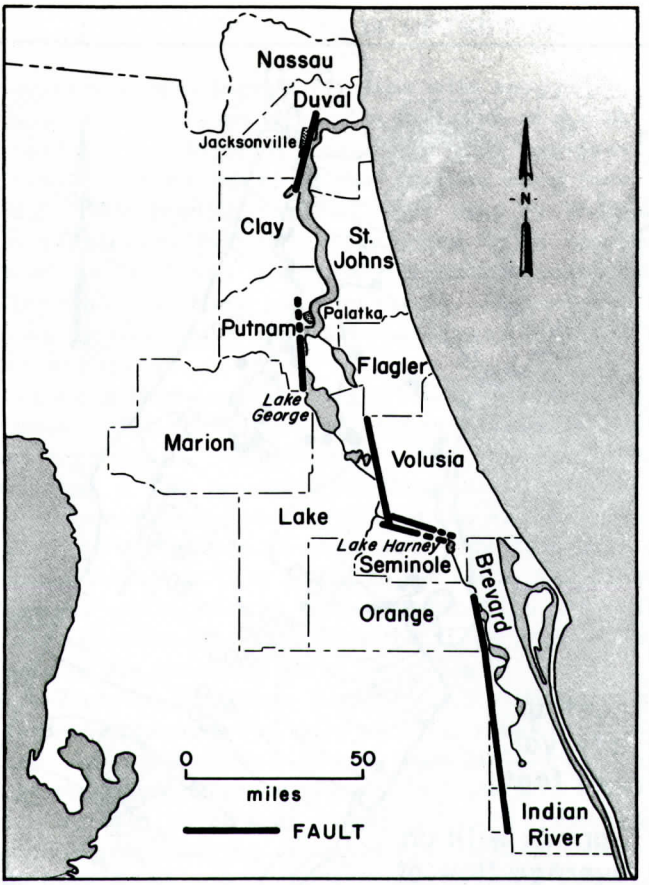

Figure 3. Locations of faults reported along the St. Johns River.

Through faulting and associated Ocala uplift movements, soluble sediments have been lifted close to the land surface at some sites resulting in an acceleration of solution and excavation of valleys. Furthermore, some areas may have subsided slightly through faulting (Wyrick, 1960; Barraclough, 1962). Such low regions would capture surface runoff and help establish surface drainage routes. Captured surface runoff would result in a greater quantity of water percolating downward in these low areas, resulting in an increase in solution.

In addition to influencing solution and stream genesis, warping and faulting have played major roles in giving relative permanence to the St. Johns River through some areas by increasing artesian discharge (Figure 4). For example, the aquiclude over areas where the artesian aquifer has been uplifted usually is thin and relatively permeable, conditions favoring discharge. Also, faults may extend from the

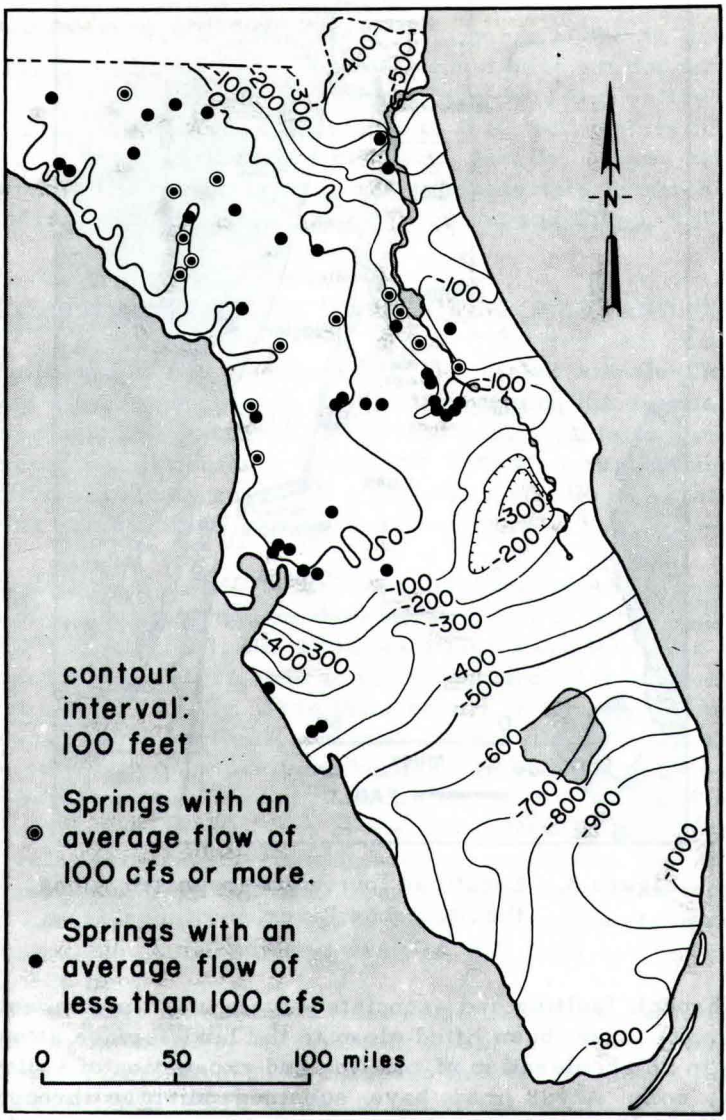

Figure 4. Contour map of upper surface of principal artesian aquifer. Spring data from Stringfield, 1966; contours from Vernon, 1955. In northern and central peninsular Florida, the elevation of the upper surface of the aquifer is largely the result of movements associated with the Ocala uplift.

artesian aquifer through the aquiclude to the land surface and serve as avenues for artesian discharge.

RIDGES AND HILLS IN OFFSET AREA

A comparison of topographic maps with several geological maps of Florida shows the St. Johns River, in the area of the St. Johns River Offset, to thread its way through ridges and hills composed of sediments called the Caloosahatchee Marl or the Nashua Marl (see Cooke's map of 1945 and the 1959 modification of this map by Puri and Vernon). These "Caloosahatchee-Nashua" ridges and hills are limited to the general area of the St. Johns River Offset. Such a correlation indicates the possibility that the flood plain of the St. Johns River Offset could have originated by solution of the carbonate fraction of the "Caloosahatchee-Nashua" sediments.

In evaluating this possible origin for the offset valley, the nature of the sediments within the various "Caloosahatchee-Nashua" ridges and hills becomes significant. Consequently logs made of cuttings from holes drilled along a line crossing the Crescent City Ridge (Figure 2), one of the most conspicuous and largest of the ridges, were examined. The line of borings extends in an east-west direction from the St. Johns River across the Crescent City Ridge to the town of Crescent City (Figure 5). Columnar sections are presented in Figure 6. The presence and amount of carbonate (shells) within the ridge sediments become the critical factors in the examination of these logs.

Hole A-7 on the St. Johns River near Fruitland was drilled to a depth of 48 feet and encountered only an occasional oyster shell. Holes A-8, A-9, and A-10 penetrated more than 50 feet of quartz sand and clayey sand beneath the ridge surface. In Hole A-8 a few shells were encountered at a depth of 68 feet. Hole A-9 was terminated at a depth of 61 feet without encountering any shells. No shells were found in the cuttings from hole A-10, all of the sediments being sand and highly organic sands and clayey sands. In holes A-11 and A-12 approximately 40 feet of loose sand was penetrated beneath which is a mixture of quartz sand and shells. This sand and shell zone extends from 40 to 63 feet in A-11 and from 38 to 48 feet in A-12. Beneath this zone in both holes are fossiliferous sands and clayey sands characterized by abundant small, transparent foraminifera. Hole A-13 at the edge of Crescent City penetrated 58 feet of loose sand and clayey sand. At 58 feet sediments containing a few shells were encountered. Hole A-14, by Crescent Lake, entered a shell bed at a depth of 28 feet. These shells are overlain by black, highly organic, quartz sand and plastic, gray, sandy clay or clayey sand.

Cuttings from other holes (Figure 5) in the area of the Crescent City Ridge give additional information. In well LP 196 at Welaka, limestone was encountered at a depth of only 60 feet. In that well the upper 25 feet of sediments consist of quartz sand and clayey sand with a few shell fragments. Phosphorite is present in the sands and clayey sands from 25 feet to 60 feet. In hole A-1, drilled on state road 309-A about one mile east of state road 309 at Welaka, the upper 33 feet are

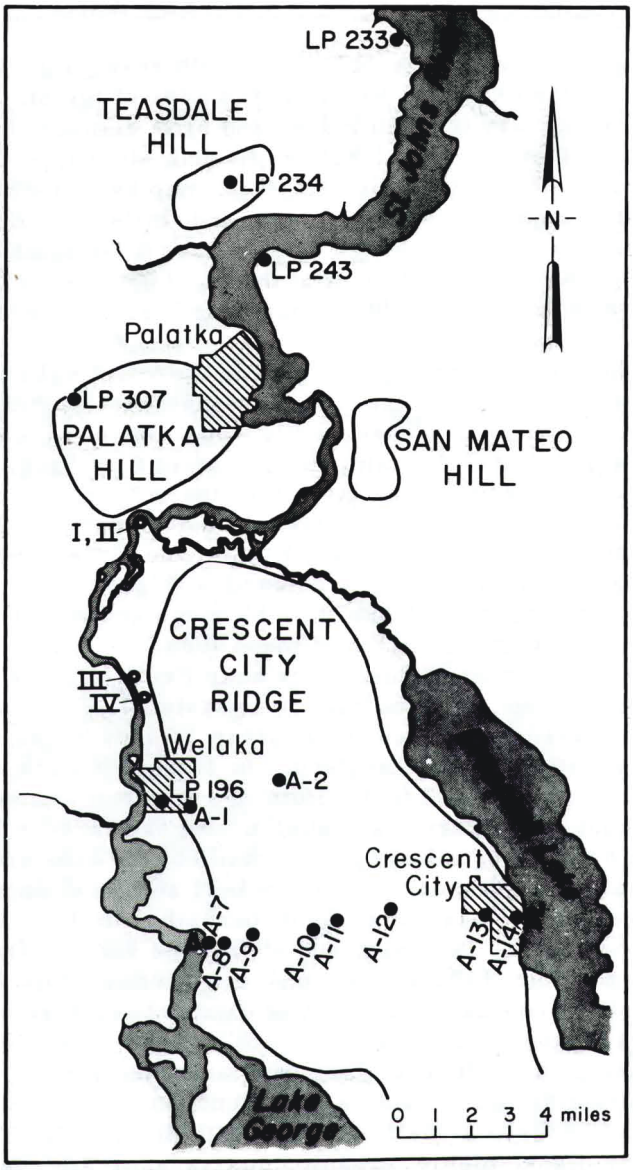

Figure 5. Index map to Putnam County wells discussed in text. Locations of samples collected along the St. Johns River are indicated by Roman numerals. Analyses of the samples are given in Table 3.

loose sand. The interval from 33 to 63 feet consists of clayey sand with a few shells. Hole A-2 was drilled near the southwestern city

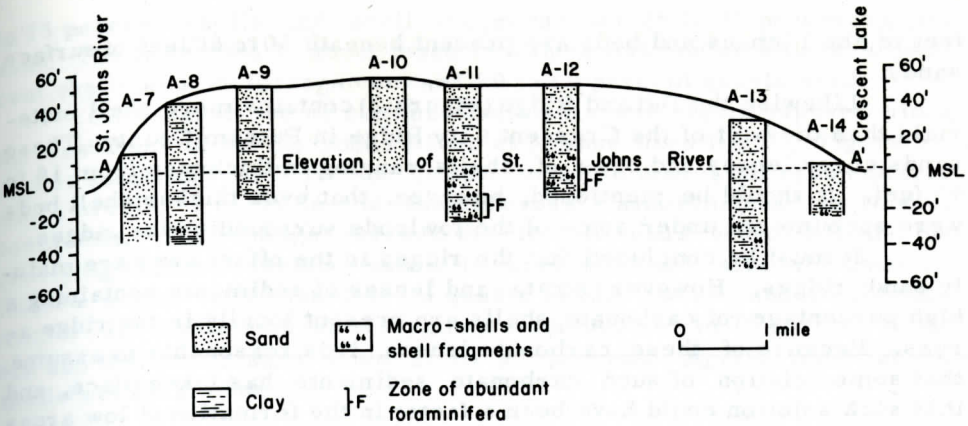

Figure 6. Columnar sections of wells drilled across the Crescent City Ridge from the St. Johns River to Crescent Lake. See Figure 5 for locations of the holes.

limits of Pomona Park about midway between the St. Johns River and Crescent Lake. The first 55 feet are loose quartz sand and clayey sand. There are rare fragments of fossils. Phosphatic sediments were encountered at a depth of 55 feet.

In summary, loose quartz sand and clayey sand make up the upper 30 to 50 feet of much of the Crescent City Ridge in Putnam County, and these sands may be even thicker in some places (Figure 6). It appears that the ridge is not held up by, nor is it composed primarily of, fossil shells of the Caloosahatchee or Nashua marls.

This same conclusion can be drawn for the hills areas just west of the St. Johns River in Putnam County (Figure 5). Well LP 307, drilled through Palatka Hill, penetrated 145 feet of sediments consisting mostly of quartz sand before encountering 20 feet of materials containing fragmented shells. Well LP 234, drilled on Teasdale Hill, penetrated 84 feet of quartz sand beneath the hill surface. These surface sands are underlain by 18 feet of sediments with a high content of shells.

In contrast, wells drilled in low areas near the St. Johns River in the same general latitudes encountered considerably less surface sand. For example, in well LP 243, about 3 miles north of Palatka, 23 feet of quartz sand, clayey sand and clay were penetrated. There are 72 feet of shell material beneath these surface sediments. In well LP 233, located on the river approximately 10 miles north of Palatka, 52 feet of quartz sand, clayey sand and clay were penetrated beneath the land surface. Under these surface materials are 63 feet of shells. It is apparent that the surface sands are thicker in the hill areas.

Logs of wells drilled on the southern part of the Crescent City Ridge in Volusia County show an increase in shell content. For example, at some sites east and southeast of Lake George from 23 to 75

feet of shell lenses and beds are present beneath 30 to 60 feet of surface sands.

Likewise the DeLand Ridge (Figure 2) contains more shell materials than the part of the Crescent City Ridge in Putnam County. These sands cover lenses and beds of shells ranging in thickness from 10 to 45 feet. It should be mentioned, however, that even thicker shell beds were encountered under some of the lowlands surrounding the ridges.

It must be concluded that the ridges in the offset area are mainly sand ridges. However strata and lenses of sediments containing a high percentage of carbonate shells are present locally in the ridge areas. Because of these carbonate lenses, it is reasonable to assume that some solution of such carbonate sediments has taken place, and that such solution could have been a factor in the formation of low areas and in the establishment of surface drainage patterns within the region of the St. Johns River Offset.

SEDIMENTS IN OFFSET AREA

A comparison of sediments composing the Nashua and Caloosahatchee marls with the quartz sands underlying the ridges and hills within the offset valley is important in further considerations of solution of the marls as a factor in the origin of valleys and residual sand ridges.

Sediments Exposed in the Northern St. Johns River Offset

Most of the sediments in the exposed banks of the St. Johns River from the northern extremity of the Crescent City Ridge to the central latitude of Lake George are nonfossiliferous, homogeneous fine quartz sands. A sample of this sand was collected from a bluff on the east bank of the river about 2 miles north of Welaka (location III, Figure 5).

Caloosahatchee or Nashua sediments are exposed at a few locations along the valley walls. In the extreme northern part of the St. Johns River Offset (locations I and II, Figure 5) 3 to 4 feet of gray to white nonfossiliferous surface sands overlie 8 to 9 feet of Caloosahatchee or Nashua sediments. Here the Caloosahatchee or Nashua consists of beds or lenses of quartz sand, shells and shell fragments alternating with thinner beds or lenses of nonfossiliferous, white micaceous quartz sand. Although some shells are unbroken, most of the shells are fragmented. The contact between the overlying barren surface sands and the Nashua sediments is very sharp. Samples were collected from the top two of several shell beds present in the exposure. These are the main shell beds exposed. They are separated from each other by a lens of white, micaceous quartz sand about one foot thick. The upper shell bed is approximately two feet thick and consists of 65

to 75 percent shells and shell fragments and 25 to 35 percent quartz sand. The lower bed is about three feet thick and contains 40 to 50 percent shells and shell fragments and 50 to 60 percent quartz sand.

Caloosahatchee or Nashua sediments are again exposed along the east bank of the St. Johns River at Nashua (location IV, Figure 5). At this location 4 to 5 feet of yellow to white nonfossiliferous quartz sands overlie 9 to 10 feet of fossiliferous Caloosahatchee or Nashua sediments. The change between the fossiliferous sediments and the overlying barren surface sands is abrupt. The Caloosahatchee or Nashua at this site consists of quartz sands, shells, and shell fragments. The shells and shell fragments compose 65 to 75 percent of the bed and the quartz sand from 25 to 35 percent. Again, most of the shells are fragmented.

Comparison of Sand in Samples from the St. Johns River Bank to Sand in the Ridge Area

Size analyses of the surface sands from sites within the Crescent City Ridge and Palatka Hill, samples A-D in Table 2, should be compared with analyses of samples collected along the St. Johns River, given in Table 3. In samples from the St. Johns River bank, the percentage of fine quartz sand (1/8 to 1/4 mm) varies from 62 percent to 68 percent. In contrast, for samples collected from the Crescent City Ridge and the Palatka Hill the percentage of fine sand varies from 83 percent to 86 percent. Furthermore, from 23 to 36 percent of the quartz sand extracted from the shell beds along the St. Johns River falls within the size range for medium sand (1/4 to 1/2 mm). In contrast, only 4 to 11 percent of the sand from the ridge and hill areas is of this size. The anomalously low percentage of medium sand and the correspondingly high percentage of very fine sand (1/16 to 1/8 mm) in the nonfossiliferous, homogeneous sand from the river bank (sample III, Table 3) is thought to be due to deposition of very fine sand by the St. Johns River during floods.

Based on these samples, the sands from the Crescent City Ridge and the Palatka Hill are finer than the quartz sand extracted from the Caloosahatchee or Nashua marls exposed along the valley walls. The ridges, therefore, could not be simple insoluble residues resulting from the solution of the carbonate fraction of the Caloosahatchee or Nashua sediments. Moreover it would be probable that the Crescent City Ridge and similar areas never, as ridge entities, contained sufficient quantities of fossiliferous Caloosahatchee or Nashua sediments for solution of these shells to form a depression as large as the St. Johns River Offset. Such conclusions, however, do not preclude the possibility that solution of Caloosahatchee or Nashua sediments deposited before the ridges and hills were formed, or deposited in low areas after the ridges were formed, played a role in the formation of the present St. Johns River Offset.

Table 3. Mechanical Analyses of Quartz Sand Extracted From Samples Collected Along St. Johns River.

Sample No.	Percent quartz sand retained on mesh					
	10 (4 to 2 mm)	18 (2 to 1 mm)	35 (1 to 1/2 mm)	60 (1/2 to 1/4 mm)	120 (1/4 to 1/8 mm)	230 (1/8 to 1/16 mm)
I			.89	30.73	67.45	.99
II			.61	35.83	63.10	.46
III			.13	2.16	68.09	29.62
IV		.09	3.15	23.30	62.71	10.75

I Upper shell bed of Caloosahatchee or Nashua sediments at extreme northern part of St. Johns River Offset (Sec. 27, T. 10 S., R. 26 E).

II Lower shell bed at extreme northern part of St. Johns River Offset (Sec. 27, T. 10 S., R. 26 E).

III Sand from bluff on east bank of St. Johns River about 1/2 mile south of Saratoga (Sec. 21, T. 11 S., R. 26 E).

IV Caloosahatchee or Nashua sediments at Nashua, Putnam County, (Sec. 28, T. 11 S., R. 26 E).

SHELL ACCUMULATIONS ALONG THE ST. JOHNS RIVER

There appears to be a thickening of shell accumulations along some parts of the St. Johns River. The average thickness of shell accumulations of 13 wells located on the tops of ridge and hill areas in the St. Johns River Offset is about 26 feet. The average thickness of shell beds in 5 wells located in low areas near the St. Johns River is about 64 feet. For example, well LP 234, located on Teasdale Hill (Figure 5), a few miles to the west of the St. Johns River, has only 18 feet of shell beds. Well LP 307 on Palatka Hill penetrated no more than 20 feet of shells. In contrast wells LP 243 and LP 233, drilled only a few feet from the St. Johns River have thicknesses of 72 feet and 63 feet of shells respectively, a threefold increase in thickness of shells along the river.

In addition to being thicker, the shells along the river apparently extend downward to greater depths (to lower elevations). The average elevation of the bottom of shell beds in seven wells located on the top of ridge and hill areas is about -26 feet. The average elevation of the bottom of shell beds in five wells located in low areas near the St. Johns River is approximately -94 feet.

If these wells are representative, it can be concluded that shell accumulations thicken along the St. Johns River. In addition, it can

also be concluded that the shells extend to greater depths along the river. Such relationships are suggested by several cross sections published by previous workers. For example, sections A-A' and B-B' made by Bermes et al. (1963, Fig. 6) show late Miocene or Pliocene deposits thickening and extending to greater depths beneath and in the immediate vicinity of the St. Johns River. One of these sections shows the deposits to be thicker and deeper along the river in west central St. Johns County. Another of the cross sections shows thicker and deeper deposits along the St. Johns River in Putnam County near Palatka.

If the shells both thicken and extend to greater depths along the river, then this thickening would be significant because it would indicate that a depression existed before the shells accumulated and probably means that the river course at these sites was established before the accumulation of the shells. The age of the shells, however, is controversial. The shells were named the Nashua Marl by Matson and Clapp (1909) and were considered to be Pliocene in age. The Nashua Marl has been correlated with part of the Pliocene Caloosahatchee Marl (Cooke and Mossom, 1929).

Other workers consider the shells as late Miocene or as upper Miocene or Pliocene deposits (Bermes et al., 1963). Clark et al. (1962, p. 69) assigned a Pliocene age to Caloosahatchee sediments present along the St. Johns River near Green Cove Springs in Clay County but later considered a Pleistocene age for these sediments as more probable (1964, p. 26, 27).

Brooks (1966, p. 42, Fig. 8), in a diagram giving his interpretation of the paleogeography of peninsular Florida during the late Pliocene, pictures the Nashua Marl as being present along the St. Johns River from about Palatka southward to the Wekiva River headwaters, a site approximately 15 miles southwest of Lake Monroe along the Seminole-Orange County line. The same diagram depicts an ancestral stream entering the southern extremity of the offset valley from the southwest. Brooks now considers the Nashua Marl to be of early Pleistocene age (1968, personal communication).

If the Nashua or Caloosahatchee shells, which thicken and extend to greater depths beneath the St. Johns River, are the same shells as those present beneath adjacent high areas such as the Crescent City Ridge, the conclusion must be drawn that the river course was in existence prior to the deposition of the sediments that compose the ridge. The offset valley of the St. Johns River, therefore, could not have originated by solution of such Nashua or Caloosahatchee shells.

It must be brought out, however, that not all investigators hold the view that the shell beds in the valleys are the same shell beds as those beneath the ridges. Brooks (1966, p. 41; 1968, p. 8) believes the valleys of all great rivers of Florida are eroded into upper Miocene or older strata. The valleys are now back filled with deposits of late Pliocene or younger beds. For example, the Bone Valley Formation occurs in eroded valleys of the Suwannee, New River, Santa Fe River, and the

Peace and other rivers of the Bone Valley area of central Florida. The Nashua Marl is present in the eroded straths of the St. Johns River. The "Caloosahatchee deposits" are present in the eroded valley of the Kissimmee River. Brooks states that it is even true that the type Caloosahatchee occurs in a valley-like depression between the Immokalee high to the south and highlands to the north.

If these ideas are correct and the Nashua Marl is Pliocene or early Pleistocene in age as Brooks believes, the valley of the St. Johns River Offset would have to be older than the marl and date back at least to early Pleistocene. The offset course of the St. Johns River would be relatively old; it would be a part of the valley of an ancestral lower St. Johns River and would predate other large segments of the present St. Johns River system. That is, the ancestral lower St. Johns River, perhaps a spring-fed river originating near the headwaters of the Wekiva River, would have followed a course now referred to as the offset valley of the St. Johns River.

PERMANENCE OF OFFSET COURSE

There are a number of factors important in giving relative permanence to the St. Johns River along the offset course. The unique and most significant feature of the offset area is the fact that it is a pronounced region of artesian discharge from the Floridan aquifer. Ground water enters the river valley by slow leakage from below. Also, numerous, well-defined springs rise in the discharge area to furnish large volumes of water to the river. At some sites discharge apparently is along faults (Bermes et al., 1963, p. 61). The discharge in the form of both slow leakage and more rapid flow is possible in some parts of the area because of the permeable nature of the aquiclude and at other sites because of the actual exposure of part of the aquifer.

Region of Discharge

Figure 7 shows the piezometric surface of the Floridan aquifer. The aquifer, composed mostly of Eocene limestones, is a part of one of the great artesian systems of North America. At any specific site ground water within the aquifer moves down gradient and at right angles to the isopiezic lines. It can be seen from Figure 7 that along the offset valley of the St. Johns River, ground water within the aquifer moves toward the river from all directions. The area of the offset valley receives ground-water discharge from a well-defined ground-water drainage basin of more than 3,500 square miles (U. S. Congress, 1938, House Document 194, p. 427). This ground-water drainage area of the offset valley is given in Figure 8. Large springs which rise within the discharge basin and flow into the St. Johns River include Alexander Springs, Juniper Springs, Ponce de Leon Springs, Silver Glen Springs, Salt Springs and Welaka Springs.

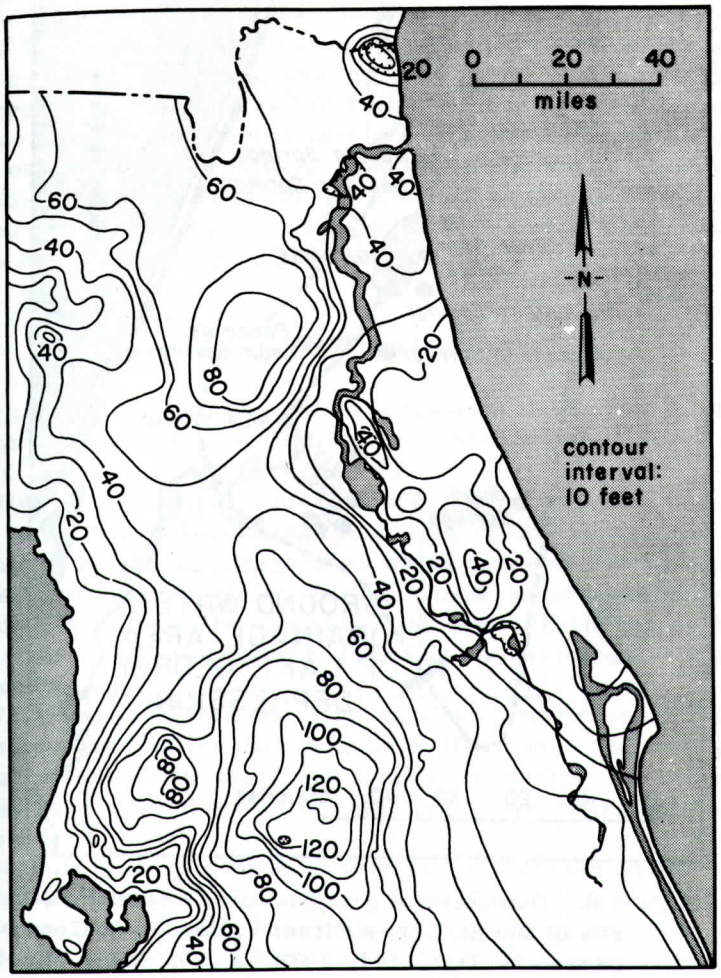

Figure 7. Piezometric surface of the Floridan Aquifer, in peninsular Florida, July 6-17, 1961. Prepared by U. S. G. S. and Florida Geological Survey.

Aquiclude

Goodell and Yon (1960, Pl. 1), in presenting a lithofacies map of post-Eocene strata of Florida, show only one area along the entire eastern part of the Florida peninsula where the ratio of sand to clay in the post-Eocene sediments is as high as, or higher than, 8 to 1. That

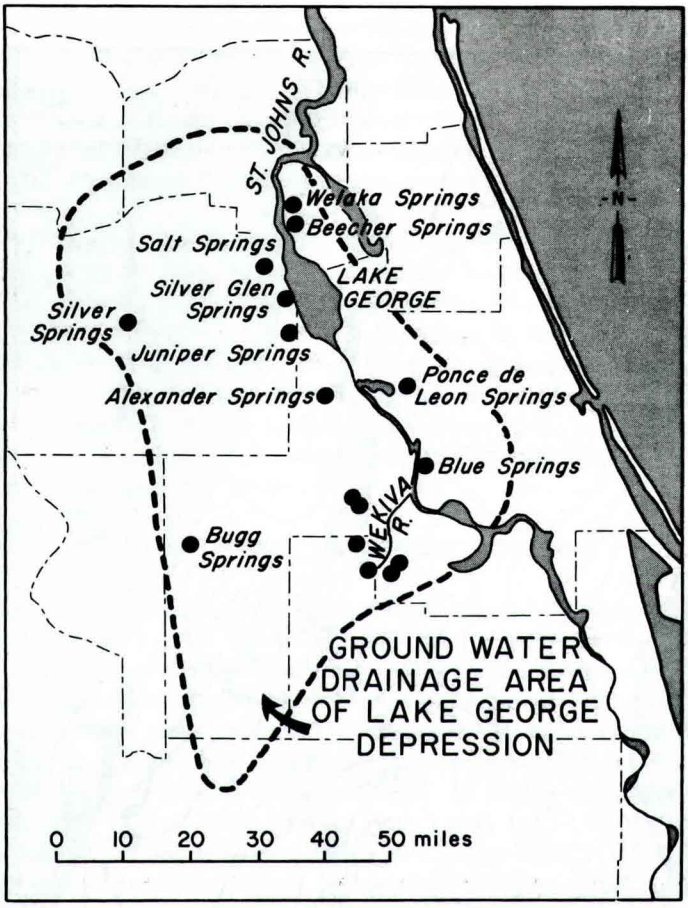

Figure 8. Boundary of ground-water drainage area of the St. Johns Offset Valley (U. S. Congress, H. Doc. 194, 1938, p. 19, Sheet 1). Sites of all major springs in the ground-water drainage area are given. Locations of springs which flow into the Wekiva River are indicated by unnamed black dots. Note the alignment of major springs along the offset course of St. Johns River.

zone in eastern peninsular Florida is relatively small and includes the region of the offset valley from the southern part of Lake George northward almost to Palatka. This area in which the sediments overlying the Floridan aquifer have relatively high permeability (low in clay) can be correlated with the area of high ground-water discharge along the northern part of the St. Johns River Offset. The clay content in the post-Eocene sediments along the St. Johns River Offset increases

somewhat southward from Lake George. Locally, however, Hawthorn sediments in this southern part have a relatively high carbonate content, are in hydraulic continuity with, and in fact, are a part of the Floridan aquifer. The numerous springs from the northern part of Lake George southward mark sites where these sediments are actually exposed or are under only a thin cover. The sediments can be seen cropping out at various springs, such as Salt Springs. In short, in the central and southern part of this discharge basin the Floridan aquifer is exposed locally and furnishes large volumes of artesian water to the St. Johns River through springs.

PRESENT INTERPRETATION

It is believed that the offset course of the St. Johns River is a part of the valley of an earlier river. This ancestral river may have originated as a spring-fed river near the headwaters of the present-day Wekiva River. Diastrophic movements associated with the Ocala uplift were important in localizing the course of this ancient river. These movements produced fracture zones which favor valley development by limestone solution, especially during times when surface drainage is downward. Therefore the course of the ancestral river probably was determined largely by fracture zones and by limestone solution along the fracture zones.

When sea level was lower the ancestral river cut its valley deeper. This deepened valley has been back-filled with sediments consisting of quartz sand, silt, clay, and lenses of shells. Solution of these shells may have played a role in establishing the course of the present St. Johns River.

Because of favorable structure in the Floridan aquifer, relatively permeable cover over the aquifer, and local breaching of the aquiclude, the region of the St. Johns River Offset today receives much discharge from the underlying Floridan aquifer. This discharge, received both directly from the aquifer and indirectly through springs and rivers which feed into the St. Johns River, gives relative permanence to the river along the offset course. Because of the relative permanence of flow along the offset course, the St. Johns River Offset is maintained as the major river course to the exclusion of any river course east of the offset area.

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