

LATE MIOCENE TO QUATERNARY SEISMIC AND LITHOLOGIC
SEQUENCE STRATIGRAPHY OF THE CHARLOTTE HARBOR AREA:
SOUTHWEST FLORIDA

by

Mark W. Evans

A dissertation submitted in partial fulfillment of the
requirements for the degree of Doctor of Philosophy in the
Department of Marine Science in
the University of South Florida

August 1989

Major Professor: Albert C. Hine, Ph.D.

Graduate Council
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CERTIFICATION OF APPROVAL

Ph.D. Dissertation

This is to certify that the Ph.D. Dissertation of

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with a major in the Department of Marine Science has been
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ABSTRACT

A network of high resolution seismic profiles, 40 vibracores, 212 surface sediment samples and 22 lithologically and/or geophysically logged boreholes have been analyzed in order to evaluate the effects of sea level cyclicity, karstification and an intercalated, carbonate to siliciclastic transition on the stratigraphic development of late Miocene to Quaternary deposits of the Charlotte Harbor area in southwest Florida. Two regional unconformities bound six discrete depositional sequences within the late Neogene deposits with the lower regional unconformity (R-3) varying from about 20 m depth to more than 115 m. This surface is present as NW-SE and NE-SW trending troughs, sharp, irregular reflections, and isolated sinkholes indicative of subaerial exposure. The R-2 unconformity which overlies and truncates the late Neogene sequences occurs as a strong, flat-lying reflector seaward of the present coastline and a series of downcutting channel complexes in backbarrier areas. Six late Neogene depositional sequences prograde to the south/southeast with a proximal-distal transition from short, steeply dipping oblique clinoforms to long, low angle clinoforms in a downdip direction. Three end-member lithologies are present: quartz/phosphate sands, a calcilutite, and poorly to well indurated, sandy, molluscan calcarenite. The depositional

sequences can be interpreted as a series of unconformity bounded, prograding fluvial to deltaic deposits. Three Quaternary litho-sequences overlie the R-2 unconformity and indicate that Quaternary infilling of the estuarine-lagoon system consisted of six transgressive, quartz dominated paralic depositional sequences which are poorly preserved due to shoreface and fluvial-estuarine channelization. The karstified antecedent topography has controlled the distribution of the Neogene and Quaternary depositional sequences in two ways: first, fluvial systems are localized in the antecedent lows which restricts the distribution of fluvially transported sands. Second, the relatively impermeable calcilutite deposits which occur as thin beds outside the lows, thicken to 20-40 m within the karstic depressions which limits vertical drainage and initiates the development of fluvial systems on the carbonate terrain.

Abstract Approved:

Major Professor: Albert C. Hine, Ph.D.

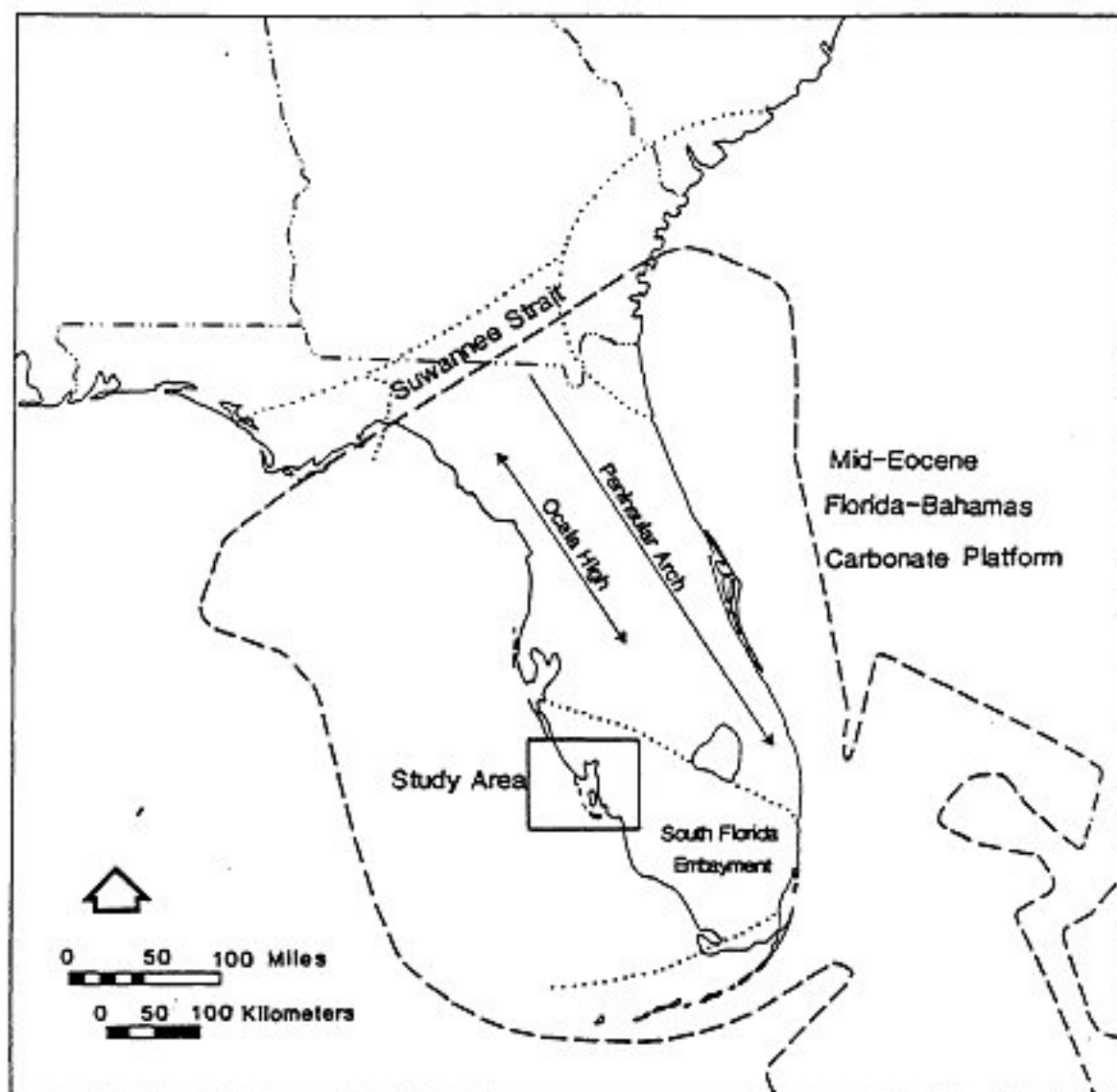
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May 23, 1989
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INTRODUCTION

The goal of this study is to identify and map the late Miocene to Quaternary depositional sequences of the Charlotte Harbor area from continuous high resolution seismic reflection profiles, and correlate and interpret the profiles using lithologic and geophysical data from boreholes, vibracores and modern depositional environments. The Charlotte Harbor region of southwest Florida lies along the northwest-southeast trending axis of the Florida Platform, which is a thick accumulation of Cretaceous and Tertiary age carbonates and evaporites of primarily shallow water origin (Figure 1; Chen, 1965; Mitchum, 1978).

The modern environments of southwest Florida are characterized by the deposition of quartz sand dominated paralic systems (Huang and Goodell, 1967; Enos and Perkins, 1977; Evans, et al., in press). The Neogene age Hawthorn Group and Tamiami Formation of south peninsular Florida consist of interbedded terrigenous clays, silts and sands, carbonates, phosphorites, and authigenic clay and silica (Scott and Upchurch, 1982) and collectively comprise a transition from the underlying Paleogene carbonates to the modern siliciclastic environments. The Neogene Period has been characterized as a time of glacio-eustatic sea-level cyclicity with high frequency (20-80 KY cycles), large magnitude (50-150 m) variations in



sea-level (Crowell, 1982; Snyder, et al., 1982; Matthews, 1984).

The exposure of the Florida platform during sea-level low-stands produces regional subaerial unconformities (Enos and Perkins, 1977; Vail, et al., 1977; Hunter, 1978), dissolution of carbonates (Esteban and Klappa, 1983; Trudgill, 1985; Back, et al., 1986) and fluvial reworking of selected areas (Kidwell, 1984; Evans and Hine, 1986). The alternating periods of exposure and submergence created by sea-level variations, in conjunction with the varied suite of lithologies produced by the long term carbonate-siliciclastic transition of the Florida platform have resulted in a complex stratigraphic record with extensive lateral and vertical variability.

This study describes the depositional controls and environments that existed during the late Neogene carbonate-siliciclastic transition, compares the resulting depositional sequences with the Quaternary sequences, and focuses on three related aspects of the depositional systems: 1) antecedent topography inherited from the underlying carbonate platform, 2) distribution and composition of lithologies and bounding unconformities comprising the transitional and siliciclastic depositional sequences, and 3) development of hydrogeologic aquifer-aquitard systems which control the preservation and alteration of the stratigraphic record by variable exposure to groundwater. The following sections examine the questions and significance of antecedent topography, sequence stratigraphy, the carbonate-siliciclastic transition, and the resulting

hydrogeologic systems in relation to the late Neogene to Quaternary evolution of southwest Florida.

Antecedent Topography

The importance of antecedent topography within siliciclastic, paralic environments has been illustrated by numerous studies (see Oertel and Leatherman, 1985 for review). The barrier/lagoon systems of Texas (Parker, 1959; Wilkinson, 1975), Florida (Otvos, 1982; Evans, et al., 1985), Georgia (Oertel, 1979), South Carolina (Moslow, 1980), North Carolina (Pierce and Colquhoun, 1970), Virginia (Morton and Donaldson, 1973) and Delaware (Kraft, 1971; Belknap and Kraft, 1985) are all localized by antecedent highs and/or increased sediment availability. These studies have identified three common facets of antecedent topographic control: 1) barrier island localization-stabilization, 2) control of lagoon location and morphology by regional slope, paleofluvial systems and headlands, and 3) variable preservation within antecedent lows.

The "highs" underlying the barrier islands of those earlier studies have been attributed to: 1) Pleistocene barriers (Pierce and Colquhoun, 1970; Oertel, 1979; Otvos, 1982), 2) interfluvial highs (Parker, 1959; Kraft, 1971; Morton and Donaldson, 1973), 3) an increased slope due to shoreface incision (Wilkinson, 1975; Evans, et al., 1985), or 4) Some combination of the above (Moslow, 1980). Whatever the origin of the antecedent high, it serves as a stable point for

localizing/stabilizing barrier development during a long term transgression which, in turn, creates the lagoonal basin.

The second important facet of antecedent topography is its role in the control of lagoon morphology and infilling sequences. Lagoonal deposits range from coarse sand and shell of tidal delta/overwash areas to marsh peats and laminated silt/clay of the deeper, quiescent basins (Lucke, 1934; Wanless, 1976; Oertel, 1985). The distribution of those environments is directly controlled by lagoon morphology (i.e., width, depth, tributary rivers, etc.) which is, at least partially, controlled by antecedent topography (Wanless, 1976; Demarest and Leatherman, 1985; Cleary and Hosier, 1986).

Halsey (1979) and Belknap and Kraft (1985) have also shown that the location of the Delaware lagoons (relative to barrier beaches) is controlled by the presence of paleofluvial systems that are linked to modern drainage systems and underlie the shoreface/nearshore area. The depositional sequences infilling the pre-existing topography consists of localized fluvial/estuarine sediments overlain by lagoonal and/or inlet deposits.

The third and final aspect of antecedent topographic control is the potential preservation of lagoonal infilling sequences and how antecedent relief affects that potential. Belknap and Kraft (1985) and Hine and Snyder (1985) document that the interaction of antecedent relief and shoreface incision result in negligible preservation of the Delaware and North Carolina estuarine/lagoon sequences, respectively. The

upper estuarine deposits that infilled the topographically lowest fluvial channels are selectively preserved (relative to lagoonal and barrier island deposits).

Collectively, the preceding factors indicate that antecedent topography significantly controls the location, morphology, infilling sequences and preservation of coastal depositional environments. The antecedent topography in all of these studies was attributed to shoreface incision on the inner shelf and/or dendritic, paleo-fluvial channelization with a more or less shore normal orientation underlying both inner shelf and backbarrier areas.

The siliciclastic, paralic environments of southwest Florida occur as a thin Quaternary veneer over the transitional, Neogene age, Hawthorn and Tamiami deposits. Extensive antecedent topographic relief within those transitional depositional environments is indicated by problems of stratigraphic correlation caused by large vertical offsets and discontinuities between closely spaced boreholes in the usually flat lying deposits of south Florida. Brooks (1972) showed a 20 m vertical offset of thick clays in wells 1-3 miles apart adjacent to Lake Okeechobee. Wedderburn, et al., (1982) and Missimer and Gardner (1976) refer to vertical discontinuities in wells and seismic records (respectively) in the Lee County area.

Several hypotheses have been suggested to explain these variations and the resulting problems of stratigraphic correlation. Tanner (1965) and Bond, et al., (1981) have

postulated a series of recent (< 7 MA) NW-SE, NE-SW trending faults with variably uplifted and downthrown blocks. Their hypothesis is based upon generalized patterns of basement structure and orientations of surficial geomorphic features, i.e., shoreline trends and stream channels.

Missimer and Gardner (1976), Missimer and Banks (1982) and Peck, et al., (1979) suggested that depth variation in "reflectors" (or stratigraphic horizons) is due to folding caused by differential subsidence and that rapid facies changes caused by sea-level fluctuations are responsible for the complex stratigraphy of south Florida. Klinzing (1982) has also indicated the presence of Pliocene patch reefs and suggested that the upward aggradation associated with these features may have some influence on stratigraphic correlations. Brooks (1972) summarized the geological literature of south Florida and concluded that both differential subsidence and Neogene structural faults have occurred and controlled stratigraphic relationships.

Wolansky, et al. (1983), Evans and Hine (1986), and Evans et al. (in press) used high resolution seismic profiles in the Charlotte Harbor area to suggest that karstic dissolution and collapse are responsible for the complex Neogene stratigraphic relationships. The shallow (100-150 m) geophysical and lithologic data presented in this report cannot address basement structure except where other investigators have extrapolated structural control into Neogene deposits.

Several studies have shown the importance of carbonate

dissolution processes (karstification) on creating and controlling depositional topography (Purdy, 1974; Enos and Perkins, 1977; Hutton, et al., 1984; Hine, et al., 1988; and Evans, et al., in press), however, there is no general acknowledgement of karstic processes in the depositional sequence models of carbonate systems (Wilson, 1975; Read, 1982). Karstic processes have been largely ignored in the carbonate platform literature because they are commonly perceived as random processes creating only isolated shafts and sinkholes and/or surficial dissolution (Longman, et al., 1987; Bond, et al., 1981), or operative only in the subsurface (Fontaine, et al., 1987).

Topographic relief on originally flat-lying carbonate platforms is created by surface and subsurface dissolution of the limestones (Choquette and James, 1988). Drainage of carbonate terrains is primarily downward with minimal runoff and limited fluvial development (Bögli, 1980). Dissolution is localized by fractures and joints and results in enclosed linear to circular depressions (depending on fracture/joint continuity; Bögli, 1980; Hutton, et al., 1984).

Regional fracture patterns, which usually reflect structural or tectonic stresses (Price, 1966; Trudgill, 1985; Culbreth, 1988), provide a means of imprinting localized topographic relief over a large area of carbonate terrain. In light of the shallow water origin of most carbonates (Blatt, et al., 1980) and subaerial exposure resulting from eustatic sea-level variation (Vail, et al., 1977; Haq, et al., 1987),

processes of carbonate dissolution may be an important, albeit, neglected aspect of carbonate platform development and evolution.

It is a hypothesis of this study that processes of sea-level cyclicity, and carbonate dissolution and collapse control the stratigraphic development of the transitional and overlying clastic environments rather than Neogene structural adjustments as suggested by earlier investigators. Whatever the cause, it is evident that: 1) significant antecedent relief underlies both the Neogene transitional units and the Quaternary siliciclastic units, 2) this relief has important controls on the location, morphology, distribution and preservation of the depositional sequences, and 3) if the relief is due to karstic processes, it will have a fundamentally different expression than the antecedent topography of terrigenous fluvial and paralic systems.

Sequence Stratigraphy

This study focuses on the formation and infilling of the Charlotte Harbor estuarine/lagoon system through the delineation and interpretation of unconformity-bounded depositional sequences. The acceptance of orbitally induced glacio-eustatic sea-level fluctuations (Milankovich cycles; Imbrie and Imbrie, 1980; Arthur and Garrison, 1986), the advent of seismic stratigraphy, and the interpretation of high

frequency punctuated aggradational cycles (PAC's; Goodwin and Anderson, 1985) have resulted in the recognition of high frequency depositional sequences within coastal and continental margin deposits.

Unconformity-bounded sequences have been a conceptual model for many years (e.g. Sloss, et al., 1949; Wheeler, 1958; Sloss, 1988). The craton-wide sequences as described by Sloss, et al., (1949; reviewed by Sloss, 1988) were attributed to basin wide tectonic cycles and transcended formational units (i.e., lateral extents of 800-1200 km and 1-10 million year durations). The unconformity-bounded PAC's as described by Goodwin and Anderson (1985) consist of 1-5 m thick, coarsening upward cycles in shallow water, Devonian limestones. These cycles, defined from outcrop, are correlated over tens of kilometers and interpreted to represent short term (30-50 KA), eustatically controlled, depositional sequences.

Seismic stratigraphers have also developed a nomenclature for defining and interpreting unconformity-bounded depositional sequences. Seismic depositional sequences are defined as:

A stratigraphic unit composed of a relatively conformable succession of genetically related strata and bounded at its top and base by unconformities, or their correlative conformities.

(Mitchum, 1977)

Seismic sequences have lateral and vertical extents of tens to hundred of kilometers, and centimeters to hundreds of meters, (respectively; Mitchum et al., 1977).

From the variety of lateral and vertical scales of different unconformity-bounded depositional units it is obvious that it is possible to define sequences on the scale of varves to Milankovitch-based climatic cycles to periods of tectonically-based sea-level fluctuations (Anderson, 1986). The International Subcommission on Stratigraphic Classification attempted to standardize the sequence nomenclature by creating a formal stratigraphic terminology based on synthems which are defined as:

A body of rock bounded above and below by specifically designated, significant, and demonstrable discontinuities in the stratigraphic succession (angular unconformities, disconformities, etc.), preferably of regional or interregional extent.
(Salvador, 1987)

Synthems are slightly different from seismic sequences in that the latter can be identified where the bounding surfaces become conformable. However, according to Sloss (1988) that difference "is the natural product of the continuum of observations made possible by seismic exploration." More significant is the caveat that synthems, as stratigraphic entities, should only be defined where they are necessary and serve a useful purpose (Salvador, 1987).

Unconformity-bounded units are wholly based on the distribution of unconformities (or correlative conformities) and are fundamentally different from lithology-based stratigraphic units. The description of formations is based

on:

- 1) Definite lithologic composition or a distinctive, interbedded or intergraded succession of lithologic types;
- 2) Observable lithologic separation from adjacent units above and below, and;
- 3) Traceability from exposure to exposure (or from well to well in subsurface).

(Krumbein and Sloss, 1963)

One of the principal objectives of this study is to define and correlate both seismic and lithologic sequences. If the correlations are meaningful and useful (as indicated by the mapping and interpretation of seismic and lithologic unconformities), the depositional sequences should be formally defined as synthems. In order to evaluate the significance and utility of the resulting depositional sequences, they must necessarily be compared with the existing lithostratigraphic terminology. Thus, an indirect result of this study will be an evaluation of the relative value of lithology and unconformity-based depositional units for regional correlations and paleo-environmental interpretations.

It has already been stated (as a hypothesis) that high frequency sea-level cyclicity is an important control on deposition of the late Neogene to Quaternary sediments in the Charlotte Harbor area. Cyclicity in the geological record is the rule rather than the exception (Fischer, 1984; Arthur and Garrison, 1986). These cycles can be driven by processes internal to the depositional environments (autocyclic) or

external (allocyclic; Grotzinger, 1986).

Autocyclic processes can occur in a variety of environments, (i.e., delta lobe switching, channel or inlet migration, barrier overwash, etc.) and may result in vertical stacking of unconformities and depositional units (Einsele, 1982). In order to prove that the Charlotte Harbor sequences are the result of allocyclic sea-level fluctuations it will be necessary to show some vertical sequence(s) or vertical juxtaposition of submarine deposits and subaerial deposits or unconformities (or vice versa) which cannot be affected without base level changes.

The primary significance of unconformity-bounded depositional sequences is that Walther's Law is not valid across unconformities (i.e., time lines do not cross unconformities; Goodwin and Anderson, 1985; Vail, 1987). This study will define and map discrete depositional sequences within which lateral facies transitions and corresponding vertical sequences can be correlated.

Carbonate-Siliciclastic Transition

The depositional environments and sequences produced by carbonate-dominated systems are lithologically and geometrically distinct from those of siliciclastic systems (Coogan, 1972; Wilson, 1975; Brown and Fisher, 1980; Fontaine, et al., 1987). Although carbonate and siliciclastic systems

each include a large variety of depositional processes and environments (reviewed by Scholle, et al., 1983, and Scholle and Spearing, 1982, respectively), generalized depositional models indicate the basic difference in sequence geometries as allochthonous, fluvial/deltaic progradation of siliciclastic margins versus autochthonous aggradation of carbonate platforms/ramps (Sangree and Widmier, 1977; Bubb and Hatelid, 1977; Brown and Fisher, 1980).

In a comparison of the hydrogeologic regimes of the carbonate Yucatan Peninsula and the siliciclastic-dominated Florida Peninsula, Back and Hanshaw (1970) showed that differences in permeability and susceptibility to diagenesis resulted in increased lateral runoff and the development of fluvial systems on the Florida peninsula, and vertical drainage with enhanced karstification on the Yucatan peninsula. These different drainage styles present important questions with regard to the transition of the Florida peninsula from a carbonate to siliciclastic dominated regime. Specifically, what are the depositional environments that comprise the transition and, secondly, what are the lateral and temporal extent of those deposits that result in decreased rates of vertical drainage and the subsequent development of fluvial systems on carbonate terrains?

Walker, et al., (1983) has described a model for transitional carbonate to clastic sequences based on middle Ordovician deposits of the southern Appalachians. This model, as well as the craton-wide sequences described by Fischer

(1961), Wanless (1972), Galloway and Brown (1973), and Wilson (1975) document stratigraphic sequences which include basinal turbidites, shelfal limestones and deltaic muds and channel sandstones. These cycles which have been shown to persist over the entire mid-continent area (1500 km from the Appalachian mountains to western Kansas) are clearly not applicable to the paralic transitional sequences of southwest Florida.

Enos and Perkins (1977) examined the Quaternary depositional sequences of south Florida and found that quartz to carbonate transitions occurred via redistribution of a quartz sand source. Although this transition occurred within paralic deposits, they did not identify a distinct transitional environment(s) or a depositional sequence that would result in long term permeability changes to the carbonate terrain.

Missimer and Banks (1982) and Peck et al., (1979b) have also documented carbonate-clastic depositional sequences in the mid-Neogene deposits of southwest Florida. These sequences, based on subsurface well log data which show rapid lithologic transitions in vertical sequence, are interpreted as regressive cycles beginning with shelfal limestones which are overlain by prograding clastic deltas. Riggs, et al., (1982) and Riggs (1984) have recognized transgressive, carbonate-clastic sequences in the Neogene deposits of the North Carolina inner continental shelf as a series of phosphatic, terrigenous sands overlain by clayey, phosphorite sands and capped by weathered or indurated shelfal carbonates.

Collectively, these studies indicate that there is a

question as to whether the transitional, late Neogene, paralic deposits of southwest Florida accumulated during transgressive or regressive conditions, and consequently what depositional processes are responsible for introducing quartz sands into the carbonate environment. This study will address that question by defining and mapping depositional sequences in 3-dimensions from a closely-spaced seismic profile grid and adjacent borehole data.

Hydrogeology

The upper Hawthorn Group and Tamiami Formation contain several important aquifer systems and collectively function as an aquitard for the underlying Floridan aquifer (Wedderburn, et al., 1982; Wolansky, 1983). Because lateral and vertical facies changes control the distribution of permeable and impermeable sediment types (i.e., sandy and clayey units, respectively), facies changes also control the distribution of hydrogeologic units.

The permeability of carbonate rocks is not only dependent upon sediment type (i.e. intergranular porosity), but also on jointing and bedding (Freeze and Cherry, 1979; Bögli, 1980). Aquifer distribution is therefore dependent on the distribution of both basic depositional facies, the regional fracture (jointing) pattern and the postulated enlargement of those fractures by dissolution.

The permeable aquifers and impermeable aquitards that result from the distribution of lithologic types and limestone fractures are not passive reflections of the depositional record. Processes of carbonate dissolution and diagenesis which alter the stratigraphic record are ongoing (Hanshaw, et al., 1971; Fanning, et al., 1981) and are directly controlled by the chemical and physical nature of the groundwater systems (Bathurst, 1975; Bögli, 1980; Trudgill, 1985).

Numerous local and regional hydrogeologic investigations have occurred in the Charlotte Harbor area; e.g., Sutcliffe, (1975), Wilson, (1977), Boggess, (1974), and Wedderburn et al., (1982). Although there is a general consensus on the regional hydrogeologic framework, correlation of specific upper Hawthorn/Tamiami aquifers has not been successful (Wolansky, 1983). This study examines the genetic relationship between lithologic sequences and the underlying pattern of karstification, and provide the basis for understanding the distribution and relationships of the hydrogeologic units.

It is an assumption of this study that much of the current confusion of south Florida lithostratigraphic, hydrostratigraphic and biostratigraphic nomenclature is due to the attempted regional correlations of lithologically determined formations. A hypothesis of this study is that these Neogene and Quaternary formations represent multiple, eustatically controlled, depositional cycles with progressive, updip-downdip facies translations and that those sequences can be seismically and lithologically identified and correlated.

Thus, within the overall goal of documenting the late Miocene to Quaternary depositional sequences, the specific objectives of this study are:

- 1) To define the pattern and role of karstification in creating small intra-platform basins;
- 2) Evaluate the pattern of karstification with respect to basement structure and regional fracture patterns;
- 3) Define and correlate the depositional environments and distribution of the seismic and lithologic sequences comprising the late Miocene strata of southwest Florida;
- 4) Examine the resulting depositional model with respect to the hydrogeologic system; and,
- 5) Document the role of karstified antecedent topography upon the development of the Quaternary siliciclastic-dominated fluvial/estuarine environments of south Florida.

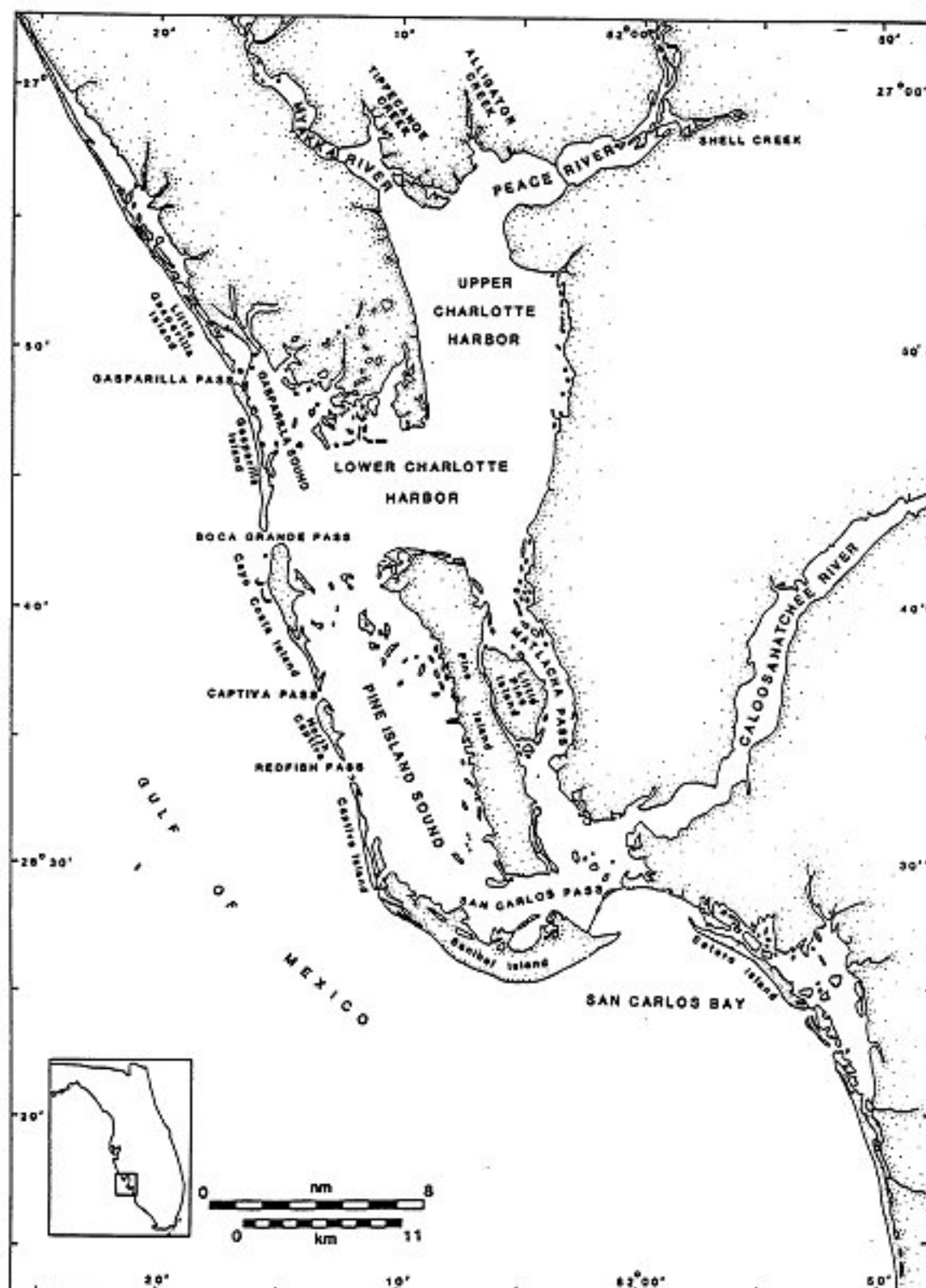
STUDY AREA

Climate

The climate of southwestern Florida is humid and subtropical with long, wet summers (75% of 127 cm annual average, June to October) and relatively drier, cooler winters. Maximum temperatures average 34.4° C with a range of 30.5° C (January) and 38.0° C (July-August) at Punta Gorda (Estevez, 1986). Prevailing winds are from the northeast during the winter and late spring (November-March) and from the south to southeast during the remainder of the year. Hurricanes and tropical storms are common from June to November with a 50% probability in any year (Ho and Tracey, 1975). Five direct and 24 indirect hurricanes have affected the area since 1900 (Estevez, 1986).

Physiography and Bathymetry

The Charlotte Harbor estuarine system occupies an area of 725 km² on the southwest coast of peninsular Florida (Figure 2). The Harbor is a complex estuary composed of two lagoons (Pine Island and Gasparilla Sounds), the estuary proper (Charlotte Harbor), an intra-estuarine lagoon (Matlacha Pass) and the tidal portions of the three tributary rivers (Myakka,

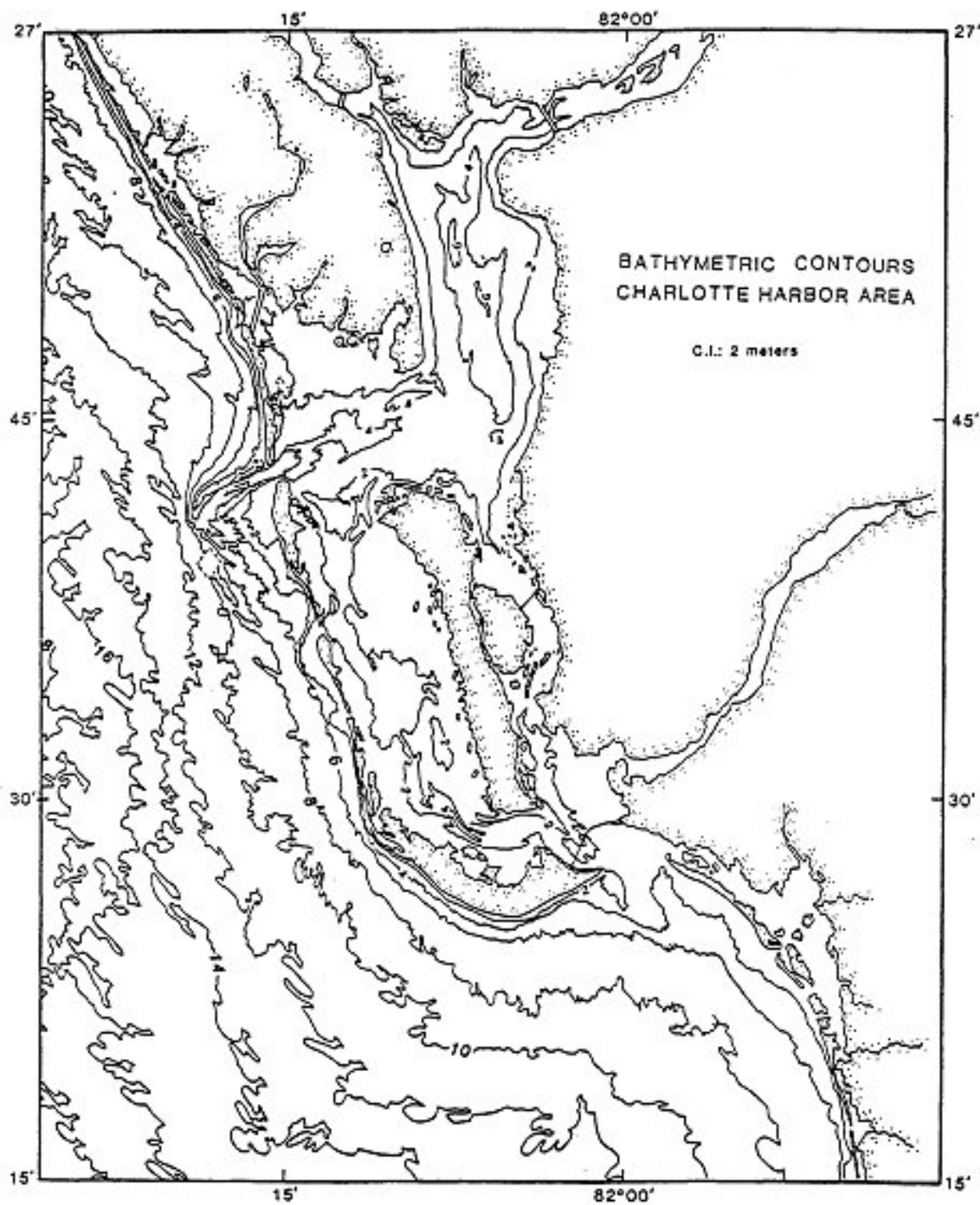


Peace and Caloosahatchee; Figure 2). The average depth of the system is approximately 2.3 m; however, the depth of each component varies significantly (Figure 3). Huang and Goodell (1967) have divided the system into four bathymetric zones: 1) the broad shallow Harbor bounded by deep narrow channels, 2) shallow lagoons with extensive sand/seagrass flats from 0-2 m, 3) slopes adjacent to the primary tidal channels with depths between 2-4 m, and 4) tidal channels with depths greater than 4 m (to a maximum of 17 m in Boca Grande Pass; Figure 3).

The combined drainage basins of the Myakka/Peace Rivers and numerous coastal creeks occupy an area of about 8500 km² (Figure 4). The basin of the Caloosahatchee River (excluding Lake Okeechobee and its tributaries) covers approximately 3300 km² (Figure 4). Topographic elevations within the collective basins are low, averaging 5.5 m in Sarasota County, 0.9 m in Charlotte County, 2.1 m in Lee County, and 17.1 m in DeSoto County (Estevez, 1986).

Hydrology

Discharges of the tributary rivers correlate closely with rainfall, with the highest flows during the rainy season in late summer and early fall. High flow conditions produce vertical stratification in the estuaries with saline bottom waters reaching 8-15 km upstream (Taylor, 1974; Stoker, 1986). Average discharge of the Myakka River is 7.2 m³/sec and includes several periods of no-flow during each year.



Two dams on the river affect flow during annual drought conditions and probably contribute to the number of no-flow occurrences (Stoker, 1986). The maximum recorded discharge is 246 m³/sec with flows greater than 85 m³/sec accounting for 1% of occurrences (Estevez, 1986).

Average discharge of the Peace River is 327 m³/sec with recorded ranges of 1.1 to 1030 m³/sec. A dam on Shell Creek, a large tributary of the Peace River may affect discharge of the lowermost portions of the River (Stoker, 1986). The Caloosahatchee River is controlled by a series of locks and gates designed to minimize flooding of Lake Okeechobee. Average discharge of the Caloosahatchee is 40.8 m³/sec and ranges from 0.04 to 606 m³/sec (Estevez, 1986).

The estuarine system is separated from the Gulf of Mexico by a series of barrier islands (Gasparilla, Cayo Costa, North Captiva, Captiva and Sanibel; north to south, respectively, Figure 2). Tidal exchange occurs at the intervening inlets (Gasparilla, Boca Grande, Captiva, Redfish and San Carlos Passes, north to south respectively; Figure 2). Primary tidal exchange occurs at Boca Grande and San Carlos Passes with the other inlets having only localized hydraulic exchange (Huang, 1966).

The tides are mixed diurnal/semi-diurnal with a variable range which averages 56 cm at Boca Grande to 82 cm at San Carlos Pass (NOAA, 1985). Maximum flood tide at San Carlos precedes that at Boca Grande by about 15 minutes which, combined with freshwater discharge and wind stress, creates a

pattern of residual currents throughout the estuarine system.

Tidal currents at the inlets are variable with flood-dominated velocities at Boca Grande (120/98 cm/sec; flood/ebb) and San Carlos (54/48 cm/sec; flood/ebb; Coast and Geodetic Survey, Chart #856-C). Captiva Pass, which is located between Cayo Costa and North Captiva Islands, has a flood velocity of 98 cm/sec and an ebb velocity of 102 cm/sec (Coast and Geodetic Survey Chart #856-C). It is not known if the ebb-dominated asymmetry of Captiva Pass can be extrapolated to Gasparilla and Redfish Passes. Ongoing work by the U.S. Geological Survey (Tampa Sub-district) on the hydrography of the estuarine system should provide answers to these questions.

Surficial Geology

The stratigraphic nomenclature used to describe the geologic units of Florida is in a state of flux (T. Scott, personal communication). Consequently, the names and ages of most of the Neogene and Quaternary units are also changing. However, for the purpose of this discussion, it is the sedimentologic composition of the surficial deposits in the source area that are of primary significance. The surficial deposits of the combined drainage basins are a mixed assemblage of limestones, dolomites, quartz sands and clays that are all of post-Eocene age (Brooks, 1981).

The quartz sands which constitute most of the

sedimentologic substrate of the basin are Appalachian in origin and reflect multiple periods of deposition and transport (Huang, 1966; Brooks, 1981). The numerous reworkings of the quartz sands have resulted in a generally homogeneous texture and composition. Unstable heavy minerals have been chemically or physically destroyed and the quartz is present as fine to very fine sand (Huang, 1966).

The limestones are biogenic and generally represent in situ deposition with complex diagenetic histories and frequent alteration to dolomite. Erosion and transport of the carbonates is limited due to lithification. Inorganic clay minerals (primarily montmorillonite, kaolinite and palygorskite) are present throughout the basin (Brooks, 1981). The clays are present as areally limited, cohesive beds and disseminated, accessory components.

Primary phosphorite (as carbonate-fluorapatite) is disseminated throughout the Neogene units and constitutes up to 15 % of the total sediment in some deposits. Fluvially reworked phosphate deposits occur as silt to pebble-sized clasts which are present in most Charlotte Harbor samples in 1-2 % concentrations and up to 9 % in some samples (Huang, 1966). The phosphorite is mined and processed in Polk, Hillsborough, Hardee, and Manatee Counties directly in or adjacent to the Charlotte Harbor tributaries.

Figure 4 is a portion of the geologic map of Florida (redrawn from Brooks, 1981). The combined drainage basins contain the following surficial units (from oldest to

youngest):

1) Hawthorn Group (Miocene age)

Mh5- Interlachen facies; quartz sand and quartzite gravel with kaolinite clay beds.

Mh4- Groveland Park facies; deeply weathered clay sand and granular sand with kaolinitic clay, white to pale orange with thick paleo-soil (orange/red).

Mh3- Statenville type; sand, silty sand to clay, oyster bars common, mixed montmorillonite/palygorskite clays.

Mh2- Brooks Sink type; impure dolomite, clay and sand.

M1- Labelle formation (previously called lower Tamiami fm.); clastics and impure limestone, variable phosphorite concentrations, gray to green to tan matrix.

2) Bone Valley Formation (Pbv; Pliocene age)- sand, clayey fine sand with montmorillonite clays and phosphorite clasts in a greenish matrix.

3) Tamiami Formation (Pt1; Pliocene age)- impure clayey to sandy to marly limestone with phosphorite grains, soft to medium hard, tan to gray matrix.

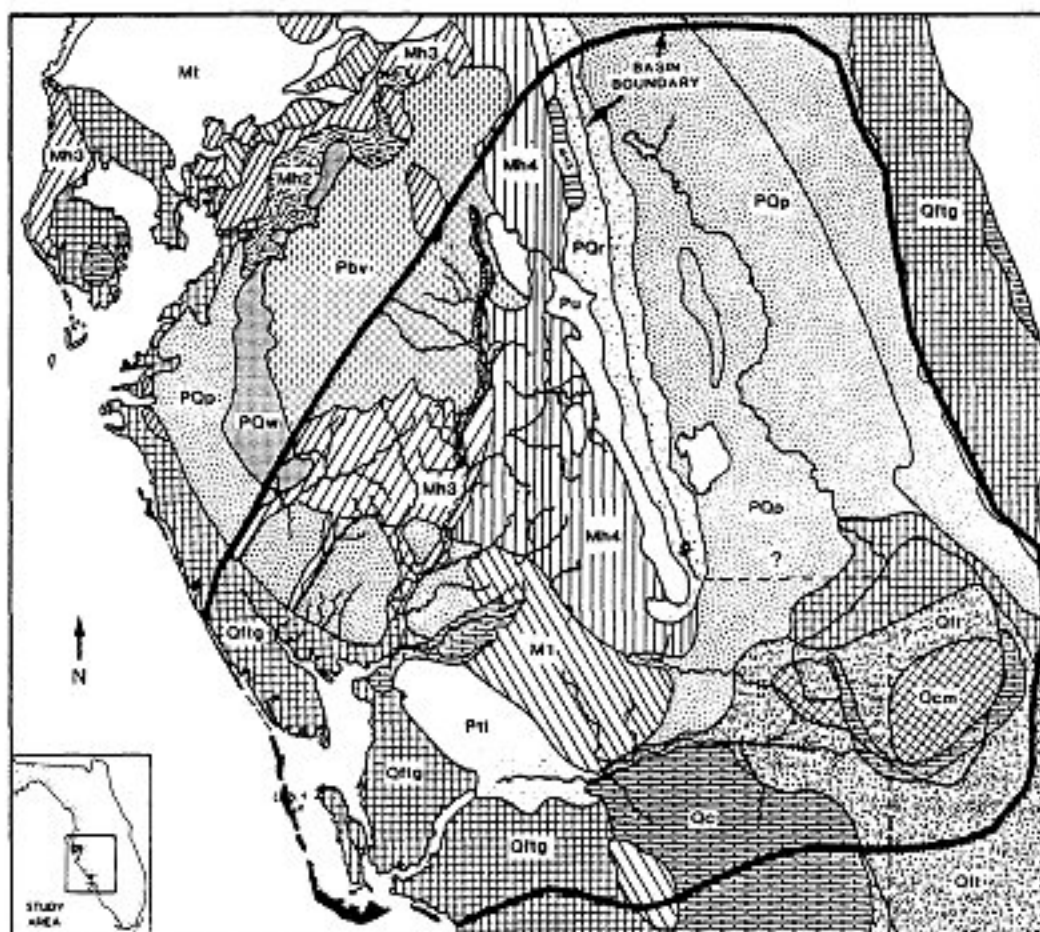
4) Unnamed (Pu; Pliocene age)-undifferentiated quartz sand (fine to very fine) with humate zones, heavy mineral zones and gravel lenses.

5) Unnamed (PQr/PQp; Plio-Pleistocene age)- deeply weathered coarse to fine sand with some clay lenses of beach/dune origin (PQr) and shelly, silty gray to greenish gray sand of lagoonal origin (PQp).

6) Caloosahatchee Formation (Qc; early Pleistocene age)-calcareous shelly sand, unconsolidated to indurated.

7) Fort Thompson Formation (Qft/Qftg; mid to late Pleistocene) shelly Chione sand with multiple hard sandy limestone caps and caliche crusts.

8) Unnamed (Qh; Holocene age)-undifferentiated quartz sand, shell, peat, and clay.



0 50
KILOMETERS

EXPLANATION

	Qh HOLOCENE UNNAMED		Pu PLIOCENE UNNAMED
	Qhg MID/LATE PLEISTOCENE FT. THOMPSON GROUP		Pti PLIOCENE TAMIAHI FM.
	Qh FT. THOMPSON MEMBER		Pbv PLIOCENE BONE VALLEY FM.
	Qcm COFFEE MILL HAMMOCK MEMBER		Mh5 } INTERLACHEN M.
	Qc EARLY PLEISTOCENE CALOOSAHATCHEE FM.		Mh4 } MIDCENE GROVELAND PARK M.
	PQp		Mh3 } HAWTHORNE GROUP STATENVILLE M.
	PQw		Mh2 } BROOKS SINK M.
	PQr		M1 } MIDCENE LABELLE FM.

REDRAWN FROM BROOKS, 1981

PREVIOUS STUDIES

Evolution of the Florida Carbonate Platform

Peninsular Florida is the subaerial portion of the much larger Florida carbonate platform (Figure 1). The subaerial portion of the platform consists of a relatively thin veneer of quartz sand (with minor percentages of clays and carbonates as described in the previous section) overlying thick sequences of carbonate and evaporitic rocks (Chen, 1965; Brooks, 1972). This section outlines the current understanding of those carbonate depositional systems and their relationship with the underlying basement structures.

A structure contour map of the pre-Cretaceous surface/postrift unconformity by Klitgord, et al. (1984) shows that the depth to basement varies from < 1 km in north central Florida to > 5.5 km in south Florida. The basement is composed of crystalline and/or Paleozoic sedimentary rocks in north central Florida with a southward transition to rhyolite/diorite/granite and/or Jurassic igneous rocks (Klitgord, et al., 1984; Barnett, 1975; Applin, 1951). The composite basement formed by these rocks represents the pre-Cretaceous postrift-unconformity surface upon which sediments gradually overlapped landward (Klitgord, et al., 1984).

The gravity anomaly map of Klitgord, et al., 1984 also reflects the basement composition and structure as a series of

NW-SE, NE-SW trending gravity lows. The distribution of crystalline basement and Jurassic igneous and volcanic rocks has been interpreted as a small block of African continental crust sutured to the North American Plate along the Brunswick magnetic anomaly, with a southward transition to attenuated continental crust (Nelson, et al., 1985; Opdyke, et al., 1987).

The carbonate deposits comprising the Florida carbonate platform were preceded by the deposition of pre-marine salt in topographic depressions caused by crustal extension (Klitgord et al., 1984; Salvador, 1987; and Winker and Buffler, 1988). Continued drifting of the African, Yucatan and North American plates resulted in the formation of the Gulf of Mexico and Atlantic Ocean during Middle to Late Jurassic time.

Oceanic crust under the central Gulf of Mexico has been documented many times (Ball and Harrison, 1969; Pindell and Dewey, 1982; Klitgord, et al., 1984; Salvador, 1987; Winker and Buffler, 1988) and related to Atlantic spreading centers via transform faults or fracture zones (FZ) under the Florida peninsula i.e. Jacksonville, Bahamas, Sunniland and Cuba FZ (Klitgord, et al., 1984; Pindell, 1985; Salvador, 1987). These fractures zones are apparently continuous under the Florida peninsula and Charlotte Harbor is underlain by the Sunniland FZ and an intersecting hinge zone (Klitgord, et al., 1984).

The opening of the Atlantic and Gulf of Mexico basins allowed the onset of marine conditions and initial carbonate deposition during the late Jurassic (Sheridan, et al., 1981). Back-reef platform limestones, dolomites and evaporites

accumulated on a postulated megabank consisting of the Florida platform, the Blake Plateau and the Bahama platform (Figure 1; Chen, 1965; Sheridan, et al., 1981; Winker and Buffler, 1988). Shallow-water, carbonate/evaporite deposition persisted and reached its zenith on the megabank during Aptian to Albian time.

The shallow-water deposition flourished behind aggradational shelf edge reefs at the Florida and Blake-Bahama escarpments (Freeman-Lynde, 1983; Ball, et al., 1988). Sheridan, et al. (1981) indicate that shallow water limestones and dolomites are continuous across the Northwest Providence Channel and Tongue of the Ocean. Reconstructions by Chen (1965), Mullins and Lynts (1977), and Winker and Buffler (1988) show these channels as deep barriers to autochthonous shallow-water deposition from at least middle Cretaceous time. These different interpretations of the distribution of shallow water Cretaceous carbonates highlight an ongoing discussion of the origin of the Bahamian Platform carbonates from discrete topographic highs separated by oceanic channels (Mullins and Lynts, 1977) or an initial megabank with subsequent differential subsidence or collapse forming the intra-bank channels (i.e., Tongue of the Ocean and Northwest Providence Channel; Schlager and Ginsburg, 1981). This discussion leads, at least rhetorically, to the question of whether infilled channels once dissected the Florida Platform as have been described by Eberli and Ginsburg (1987) on the Bahama platform.

The infilled Suwannee Strait (or Channel) separating

peninsular Florida from Georgia is well documented by Hull (1962); Chen (1965); Pinet and Popenoe (1985); and Opdyke, et al. (1987). This channel was between 100 and 300 m deeper than adjacent shelves (Chen, 1965; Schmidt, et al., 1982) and existed from at least Late Cretaceous through the late Eocene time (Chen, 1965; Pinet and Popenoe, 1985). Strong easterly flow in the Strait prevented deposition and separated Appalachian derived clastic sedimentation from carbonate deposition on the Florida Platform (Chen, 1965; Pinet and Popenoe, 1985).

Infilling of the Suwannee Strait probably began during the Late Cretaceous, with the formation of small reefs at the western (Antoine and Harding, 1965) and eastern ends (Dillion, et al., 1979). Gradual northward extension of carbonate depositional facies from Late Cretaceous to late Eocene time indicates that the channel was displaced northward (Chen, 1965). Final infilling and closure of the Strait occurred during the late Eocene (Pinet and Popenoe, 1985).

There was a major sea-level lowstand during late Eocene/early Oligocene time (Haq, et al., 1987) and world-wide changes in the paleoceanographic current system (Berggren, 1982). Resulting changes in the flow through the Suwannee Strait may have contributed to an increased rate of sediment deposition. Similarly, increased deposition rates in the Strait would result from increased fluvial sediment transport to continental margins due to the lowstand (Vail, et al., 1977), and Wilhelm and Ewing (1972) have suggested that

continent-wide sediment availability was increased due to Laramide-age orogenic reactivation of the Appalachians.

The Paleogene units of peninsular Florida have been mapped and summarized by Chen (1965), with limited paleo-environmental interpretations by Yon and Hendry (1972), Randazzo (1972) and Randazzo and Saroop (1976). Those units comprised primarily of dolomite and limestone with minor percentage of evaporites generally thicken towards the south, but show a distinct NW-SE trending trough underlying Tampa Bay and Lake Okeechobee (Chen, 1965; his Figure 38).

Chen (1965) concludes that the distribution of the Eocene Series illustrates depositional control by the pseudo-structural Peninsular Arch (Figure 1; Applin, 1951) and a slight northeastward shift of the South Florida Basin. The "Ocala Uplift" is a younger, parallel structure that originated during the late Oligocene/early Miocene as a broad, regional flexure (Vernon, 1951). Winston (1976) states that the "Uplift" represents a depositional thickening of Middle Eocene units and that interpretation is at least partially supported by isopach maps from Chen (1965).

Petrographic studies of late Eocene and Oligocene units by Randazzo (1976) and Randazzo and Saroop (1976) indicate that the depositional environments of northwest peninsular Florida were shallow marine and intertidal carbonate producing systems.

These environments have common Holocene analogues such as tidal flats and seagrass-dominated shallow shelves and the deposits possessed numerous sedimentologic and diagenetic

features indicative of repeated cycles of sea-level transgression and sub-aerial exposure.

Neogene Stratigraphy

The Hawthorn Group

Phosphatic beds of Alachua and Columbia Counties were originally described by Johnson (1888) who assigned them to the Waldo Formation. Dall and Harris (1892) abandoned the Waldo Formation and called them the "Hawthorne beds". Although Dall and Harris (1892) did not designate the Hawthorne as a formation, Dall is credited by later workers for its designation and description of a type locality at the Devils Millhopper Sink in Alachua County (Pirkle, 1956; from Scott and MacGill, 1981).

The lithologic section from the Millhopper location is presented in Figure 5 (from Scott, 1982). This core was taken adjacent to the sinkhole and contains a series of phosphate rich sands, clays, limestones and dolomites. Four lithologic units have been recognized in this core by Scott (1982; from the bottom): 4) a dolomitic, usually indurated section overlying the Eocene limestone, 3) dolomitic, sandy or clayey section, 2) dolomitic or calcareous, highly indurated section, and 1) a weathered, reworked, mostly clastic sandy/silty section. This core, along with three other nearby cores have been suggested by Scott (1982) as co-type sections for the

Hawthorn Formation which Riggs (1979) and Scott (1988) have proposed raising to Group status.

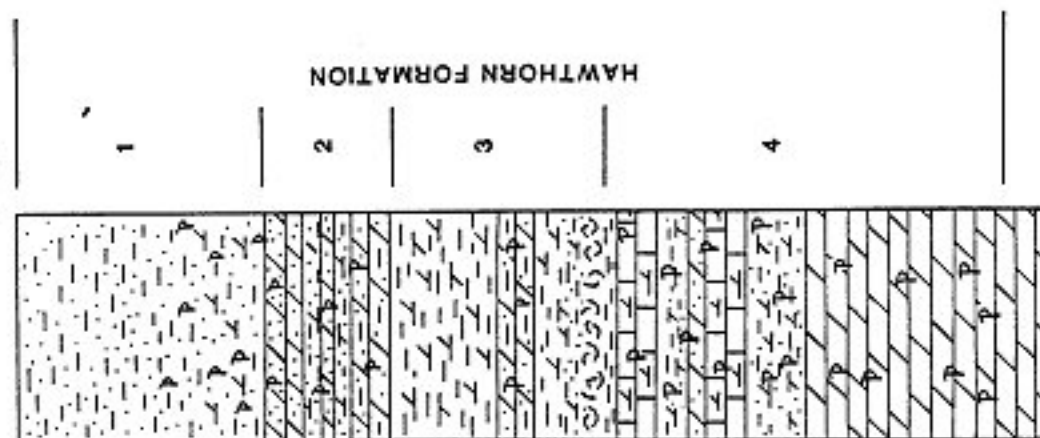
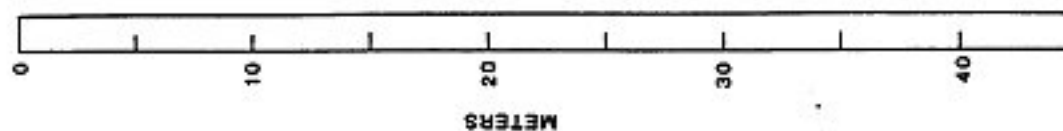
Formal description of the Hawthorne Formation was by Matson and Clapp (1909), who considered the it, at least partially, contemporaneous with the peninsular Tampa Formation and the Chattahoochee Formation of the panhandle. Sellards (1915) placed the Hawthorne beds in an Oligocene Alum Bluff formation which was contemporaneous and gradational with the Chattahoochee limestone.

Cooke and Mossom (1929) reinstated and redefined the Hawthorn (less the "e") Formation to include the Sopchoppy Limestone and Alum Bluff Formation. Cooke (1945) correlated the Hawthorn with additional panhandle units (Chipola and portions of the Shoal River Formations) and added the Brooks' Sink (Bradford County) as a co-type location for the unit. Bergendahl (1956) examined Hawthorn and post-Hawthorn deposits in Hardee and DeSoto Counties. In areas of DeSoto County that contact could only be discerned using diagnostic fossils in contrast to the type locations where the contact consists of pebbly conglomerate and weathering unconformities.

Ketner and McGreevy (1959) again redefined the Hawthorn Formation in southwestern Florida to include... "Middle Miocene rocks in the hard-rock phosphate belt including those generally assigned to the Alachua Formation and the Middle Miocene rocks in the land-pebble phosphate district."

Carr and Alverson (1959) recognized four lithologic units in the Hawthorn Formation: 1) a lower limestone unit, 2) a

HAWTHORNE #1
CORE W-11486 (FGS)



Lithologic Key

Clay	Microfossils Foraminifera
Silt	Pelecypod Fossils
Sand	Chert
Pebbles	Phosphate Phosphatic Pebbles
Limestone	Calcite/ Aragonite Cement
Dolomite	Dolomite Cement

phosphorite unit, 3) a sand unit and 4) an upper limestone. Although the lateral relationships of those units were not well understood, they regarded the sand and phosphorite units as weathering products of the limestones and mapped them collectively. They also suggested an unconformity between the Hawthorn and Tampa units based upon structure and isopach contours.

Reynolds (1962) used clay mineralogy to identify lithosomes and suggested that the Tampa Limestone and Hawthorn Formation interfinger in peninsular Florida. The eastern clastic Hawthorn lithosome possesses a montmorillonite-illite clay suite while the western carbonate Tampa lithosome contains an attapulgite-montmorillonite-sepiolite clay suite.

The Tampa and Hawthorn Formations were also combined by Wilson (1977) in a groundwater study of DeSoto County. He described a combined Hawthorn limestone unit that was underlain by the sand and clay unit of the Tampa Limestone with the top of the formation as the first limestone or dolomite or impermeable marl encountered in a well.

Riggs (1979), Scott and MacGill (1981) and Scott (1982, 1985 and 1988) have extensively reviewed the history, lithology, and distribution of the Hawthorn Formation and raised it to group status. This Group, which contains deposits that had been included in the Tampa, Hawthorn and Bone Valley Formations, is now composed of the Arcadia and Peace River Formations with a Tampa Member (in the Arcadia Formation; Scott, 1988). The lithologies of the Arcadia and Peace River

Formations are the same as those described for the Hawthorn Formation and Tampa Limestone by Reynolds (1962) and Wedderburn, et al. (1982).

In Lee County, the Hawthorn Formation has been recognized by several authors (Sproul, et al., 1972; Boggess and Missimer, 1975; Missimer, 1978; and Peck, et al., 1979a/b). Wedderburn, et al. (1982) have redefined the Hawthorn Formation in Lee County to include a green to gray, phosphatic, sandy, slightly clayey dolosilt, which had been designated by Parker (1951) and later authors as part of the overlying Tamiami formation.

The top of the redefined Hawthorn formation (of Wedderburn, et al., 1982) ranges from 0 to -40 m (NGUD) and the thickness from 130 to 250 m). The expanded Hawthorn formation is composed of a lower, predominantly carbonate member and an upper predominantly clastic member. Each of the two members consists of three lithic zones with variable percentages of fine-grained carbonate (marl-?), limestone, quartz sand, phosphorite, and clay (Wedderburn, et al., 1982).

The assigned age of Hawthorn deposits has varied from Oligocene to Pliocene (Figure 6). Dall and Harris (1892) originally designated the Alum Bluff Formation to the "older" Miocene. Extension and redefinition of the Hawthorn Formation by Matson and Clapp (1909) and the Alum Bluff (Sellards, 1915) recognized three Miocene stages (early, middle and late) and placed the Hawthorn into the middle Miocene stage.

Recent biostratigraphic analyses of the Hawthorn Formation by Huddleston (1982) and Huddleston, et al. (1982)

[illegible]

using planktonic forams and diatoms, identified an early Miocene to Pliocene age for Hawthorn Group deposits. The wide range of depositional ages is supported by correlation with interpreted ages of land mammals at sites throughout peninsular Florida (McFadden and Webb, 1982). An early to late Miocene age for Hawthorn Group deposits is also recognized by Scott (1982, 1988).

Biostratigraphic analysis of Lee County wells by Peck (1976) indicates a late Miocene (N-17; Blow, 1969) to Pliocene (N-19/20) age for Tamiami deposits. Those late Neogene, sandy, clayey dolosilts were considered to be Hawthorn deposits using the redefined nomenclature of Wedderburn, et al. (1982).

Tamiami Formation

Sandy, shelly limestones from Lostman and Allen Rivers in southwest Florida were first described by Dall and Harris (1892; Table 1). Sanford (1909) named this unit the Lostman's River Limestone. Cooke and Mossom (1929) believed those deposits to be synonymous with the Miami Oolite, but described shelly limestones from the ditches of the Tamiami Trail in Collier and Monroe Counties which they mapped as part of the Caloosahatchee Formation.

Mansfield (1939) mapped the same flat-lying, thin (<25 feet), sandy, shelly limestones along the Tamiami Trail and

Table 1. Synopsis of names, ages, and disposition of rocks now assigned to the Tamiami Formation (modified from Peck, et al., 1979b).

AUTHORS	UNIT NAME(S) DISPOSITION	AGE
Dall and Harris, 1882	Described limestone samples from the head of Lostmans and Allens Rivers, Southwest Florida.	
Sanford, 1908	Named Lostmans River Ls.	Pleistocene
Cooke and Mosson, 1929	1) Abandoned Lostmans River Ls. (considered a facies of Miami Gullite); 2) Described a limestone on the Tamiami Trail: mapped as Caloosahatchee Fm.	Pleistocene Pliocene
Hansfield, 1931	Described fossils from a sand near Pinecrest along Tamiami Trail	Late Miocene or Pliocene
Hansfield, 1932	Described fossils collected by Cooke and Mosson (1929) along Tamiami Trail	Pliocene
Hansfield, 1939	1) Named Buckingham Ls. 2) Named Tamiami Ls. (superjacent to Buckingham Ls.)	Uppermost Miocene Basal Pliocene (below Caloosahatchee Marl)
Parker, 1942	Adopted name Tamiami Formation	Pliocene; younger or = Caloosahatchee marl
Parker and Cooke, 1944	Considered Lostmans River Ls. = Cooke and Mosson's limestone on Tamiami Trail = Tamiami Limestone of Hansfield (1939); also considered Tamiami Fm., Buckingham Marl, and Caloosahatchee Marl all gradational facies equivalents.	Pliocene
Parker, 1951	Included in the Tamiami Fm. the upper Hawthorn beds of Parker and Cooke, (1944) and all upper Miocene deposits of South Florida.	Late Miocene
Schroeder and Bishop 1953, 1954	Examined benthic foraminifers of the Tamiami Fm.	Late Miocene
Schroeder and Klein, 1954	Noted Tamiami Fm. (including Buckingham Ls.) lies unconformably below Caloosahatchee Fm.	Late Miocene
Parker, Ferguson and Love, 1955	Again defined Tamiami Fm. to include all late Miocene deposits of South Florida	Late Miocene (including Duplin Marl equivalents)
Dubar, 1958	Noted in outcrop Tamiami Fm. unconformably overlain by Caloosahatchee Fm.	Miocene
Furl and Vernon, 1964	Suggested Alva and Labelle clay mbrs.	Late Miocene (Alva Faunicones)
Olsson, 1964	Placed "Pinecrest beds" between Tamiami Ls. of Hansfield and Caloosahatchee Marl	Miocene (all units)
Hunter, 1966	Formally proposed five members of Tamiami Fm.: 1) Upper units (all time equivalent): Pinecrest Sand Mbr.; Buckingham limestone Mbr.; Ochopsee Ls. (=Tamiami Ls. of Hansfield); 2) Lower units: Murdock Station Mbr.; Bayshore Clay mbr.	Miocene
Furl and Vanstrum, 1971	Suggested Ortona sand mbr.	Late Miocene
Akers, 1974	Dated oololiths of Pinecrest Sand Mbr.	Pliocene
Hunter, 1970 in Oaks & Dubar, 1974	Tamiami Formation	Medial Miocene to Early Pliocene
Missimer, 1978a, b	Suggested Ft. Myers clay and LeHigh Acres sandstone mbrs. of Tamiami Fm.	Pliocene-Miocene
Peck <u>et al.</u> , 1979	Documented 5 units composed of 3 depositional environments	Pliocene-Miocene
Wedderburn <u>et al.</u>	Recognized restricted Tamiami Fm. of Ochopsee & Buckingham Ls.	Pliocene

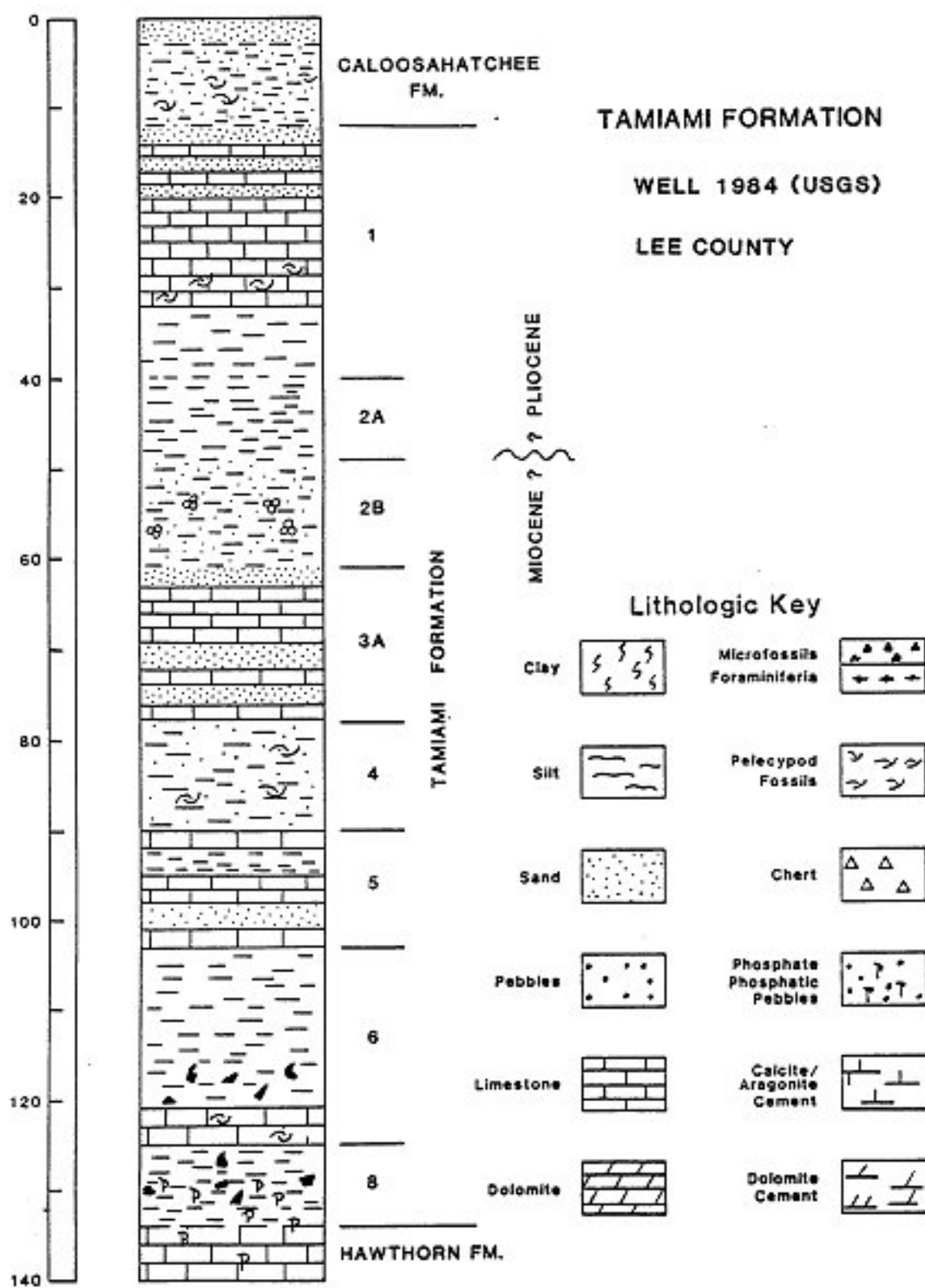
recognized it as the Tamiami Limestone. Cooke (1945) recognized that the Tamiami Limestone was identical to the Lostman's River Limestone, but did not revive the earlier name because the type section was submerged and it possessed no distinctive fossils (Peck et al., 1979b). Additionally, Parker (1942) recognized the unit as the Tamiami Formation because he considered it too sandy to be a "limestone".

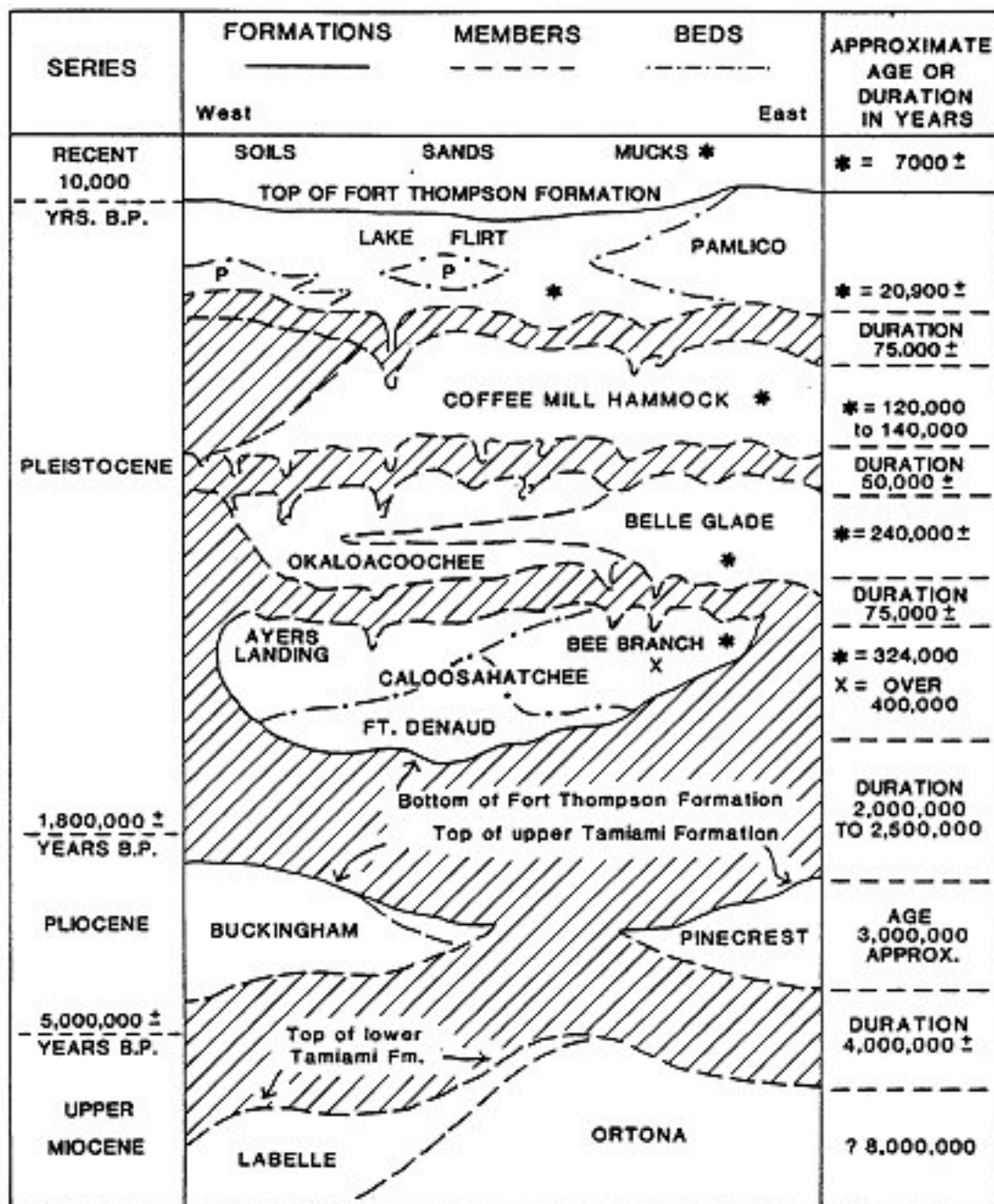
On the basis of macro-fossils from a Collier County quarry, Parker (1951) expanded the formation to include "all upper Miocene deposits in southern Florida". This redefined formation included; the Tamiami limestone, the Buckingham limestone (named by Mansfield, 1939) and a silty, greenish clay marl underlying the Biscayne Aquifer (previously considered part of the Hawthorn Formation).

Figure 7 illustrates the expanded Tamiami Formation as described by Peck, et al. (1979a/b; Lee County Well L-1984). This well log shows the variable lithologic nature of these deposits which range from shelly limestones, to quartz sands to phosphatic clays. Units 3-8 in Figure 7 is that portion which was added to the Tamiami Formation by Parker (1951).

Schroeder and Klein (1954) and Dubar (1958) recognized an unconformity between the Tamiami and overlying Caloosahatchee Formations. Missimer (1978) suggested that the base of the Tamiami Formation be placed at a pronounced unconformity mapped throughout Lee and Charlotte Counties.

Hunter (1968) identified upper and lower Tamiami members





with at least three concurrent macro-fossil assemblage zones (Figure 8). These unconformity-bounded, diachronous zones, and a late Miocene to middle Pliocene age were supported by diatom, foram, and calcareous nanno-fossil analyses by Akers (1974), Peck (1976) and Peck, et al., (1979b).

These studies and others by Hunter (1978) and Wedderburn, et al. (1982) all indicate a late Miocene age for the clayey and sandy units added to the Tamiami Formation by Parker, and a middle Pliocene age for the overlying Pinecrest, Ochopee, and Buckingham members (Hunter, 1978; Table 1). In light of these age assignments, Wedderburn, et al. (1982) have suggested that the Tamiami Formation be restricted to the Pliocene age sandy, shelly limestones.

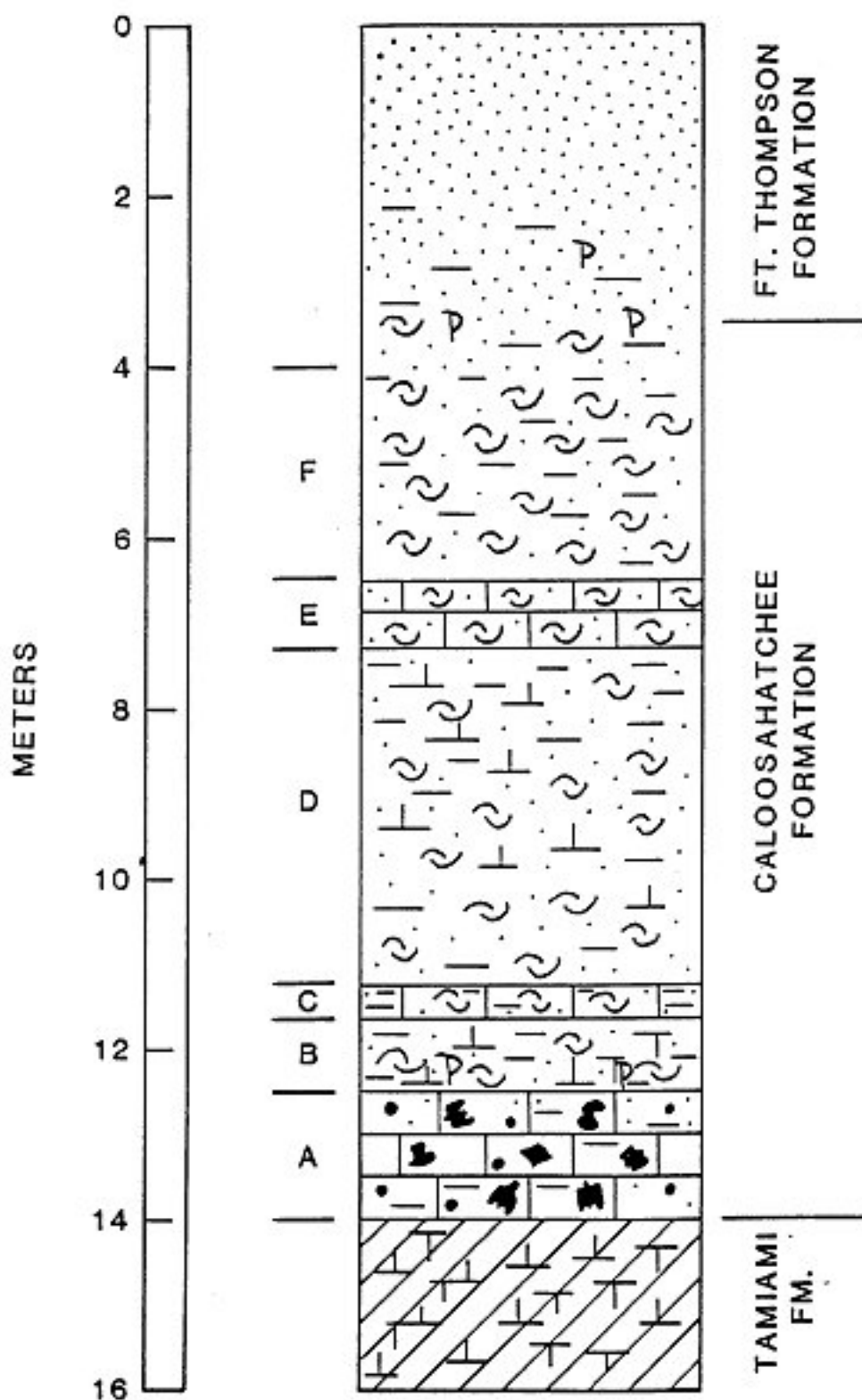
Quaternary Stratigraphy

Caloosahatchee and Fort Thompson Formations

A composite lithologic section of the Quaternary deposits outcropping along Shell Creek, Charlotte County, is presented in Figure 9. This section, redrawn from DuBar (1958), exaggerates the total thickness of the deposits by utilizing the maximum thickness of each described unit. The basal Quaternary deposits consist of a weathered, conglomeratic limestone (unit A) lying unconformably on

COMPOSITE QUATERNARY SECTION:
 Shell Creek, Charlotte County
 Redrawn from DuBar, 1958

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calcareous clays of the Tamiami Formation, and grading upward into a sandy, shelly, slightly phosphatic, calcareous mudstone (unit B).

The unit B mudstone is overlain by a thin, sandy, muddy limestone (unit C) which grades upward into unit D, which is a very shelly, sandy mudstone. Unit E is a shelly, sandy limestone similar to unit C and also grades upward into a sandy, shelly mudstone (unit F) which is unconformably overlain by clean to slightly muddy quartz sands (Fort Thompson Formation; Figure 9). A thin freshwater calcareous mudstone, termed the Lake Flirt Marl, usually overlies the Fort Thompson sands (Puri and Vernon, 1964) but is not present in the Shell Creek area (DuBar, 1958).

These individual units are usually thin (< 1 m), distributed discontinuously (Puri and Vernon, 1964; DuBar, 1974), and with the exception of the clean sands have diverse, well-preserved molluscan fossils (DuBar, 1958; Hunter, 1978). Extensive analysis of the molluscan fossil assemblages and mapping of the resulting, discontinuous bio-facies has led to a plethora of described beds, members and informal formations (see Puri and Vernon, 1964 for complete descriptions). A correlation chart showing the development of the Caloosahatchee and Fort Thompson nomenclature is presented in Figure 10.

Sandy, mollusc-rich limestones, fine quartz sands and marls outcropping along the Caloosahatchee River in Hendry,

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QUARTERNARY STRATIGRAPHIC CORRELATIONS: SOUTHWEST FLORIDA									
SELLARDS (1919)	COOKE & MOSSUM (1929)	PARKER & COOKE (1944)	DUBAR (1958)	BROOKS (1968)	UNIT	PERKINS (1969)	HUNTER (1978)	PLEISTOCENE	PLIOCENE
COFFEE MILL HAMMOCK MARL	COFFEE MILL HAMMOCK MARL	COFFEE MILL HAMMOCK MBR. (BED 6)	COFFEE MILL HAMMOCK MARL	COFFEE MILL HAMMOCK FM.	6	Q5	LAKE FLIRT	PLEISTOCENE	TAMIAHI FORMATION
FORT THOMPSON BEDS	FORT THOMPSON FORMATION	FORT THOMPSON FORMATION	OKALDA - KOOCEE MARL	FT. THOMPSON FM.	5	Q4	COFFEE MILL HAMMOCK MBR.	PLEISTOCENE	TAMIAHI FORMATION
			BED 5		Q3	OKALDA - KOOCEE BELLE GLADE MBR.			
			BED 4		Q2	CALOOSA - HATCHEE MBR.			
FORT THOMPSON BEDS	FORT THOMPSON FORMATION	FORT THOMPSON FORMATION	AYERS LANDING MBR.	FT. THOMPSON FM.	3	Q1	COFFEE MILL HAMMOCK FM.	PLEISTOCENE	TAMIAHI FORMATION
			BED 3		Q1	COFFEE MILL HAMMOCK FM.			
			BED 2		Q1	COFFEE MILL HAMMOCK FM.			
FORT THOMPSON BEDS	FORT THOMPSON FORMATION	FORT THOMPSON FORMATION	BEE BRANCH MBR.	FT. THOMPSON FM.	2	Q1	COFFEE MILL HAMMOCK FM.	PLEISTOCENE	TAMIAHI FORMATION
			BED 2		Q1	COFFEE MILL HAMMOCK FM.			
			BED 1		Q1	COFFEE MILL HAMMOCK FM.			
FORT THOMPSON BEDS	FORT THOMPSON FORMATION	FORT THOMPSON FORMATION	FORT DENAUD MBR.	FT. THOMPSON FM.	1	Q1	COFFEE MILL HAMMOCK FM.	PLEISTOCENE	TAMIAHI FORMATION
			BED 1		Q1	COFFEE MILL HAMMOCK FM.			
			BED 0		Q1	COFFEE MILL HAMMOCK FM.			
FORT THOMPSON BEDS	FORT THOMPSON FORMATION	FORT THOMPSON FORMATION	TAMIAHI FM.	FT. THOMPSON FM.	0	Q1	COFFEE MILL HAMMOCK FM.	PLEISTOCENE	TAMIAHI FORMATION
			BED 0		Q1	COFFEE MILL HAMMOCK FM.			
			BED -1		Q1	COFFEE MILL HAMMOCK FM.			

Glades and Lee Counties were described and assigned a Pliocene age by Heilprin (1887). Dall and Harris (1892) concurred with the Pliocene age and named the deposits the "Caloosahatchee beds of marl". Sellards (1915) recognized the Pleistocene age Coffee Mill Hammock Marl and Fort Thompson Beds, and Cooke and Mossom (1929) formally described the Fort Thompson Formation (Figure 10).

DuBar (1958) identified Pleistocene vertebrate fossils within the Caloosahatchee Formation and assigned the Formation an early Pleistocene age (previously Pliocene; Figure 10). Brooks (1968) used Th/U dated corals to separate a combined Fort Thompson-Caloosahatchee Formation into a series of six depositional sequences which he correlated with interglacial sea level high stands. Perkins (1968; 1977) used a variety of features indicative of subaerial exposure (i.e., soils, bored surfaces, freshwater limestones, and caliche) to differentiate five Quaternary depositional sequences.

Perkins (1977) interpreted barrier-beach, lagoon, and tidal-delta environments within each of the five, unconformity-bounded depositional sequences and correlated the sequences with chronostratigraphic units defined by amino acid epimerization (Mitterer, 1975). Although Perkins (1977) ignored existing stratigraphic terminology, Hunter (1978) recognized similar unconformity-bounded sequences and suggested that all Quaternary deposits of southwest Florida be placed within an expanded Fort Thompson formation. Much of the pre-existing formation and member nomenclature would be retained to

define chronostratigraphic and environmental subdivisions (Figures 8 and 10).

The historical development and regional correlation of the Hawthorn, Tamiami and Caloosahatchee/Fort Thompson Formations have followed similar patterns. Lithologically distinct and areally restricted beds were described and then raised to formation status in order to facilitate regional mapping. Continued areal correlation of the litho-types forced inclusion of a larger range of lithologies, which further diluted the character of the originally described formation or bed.

The inherent variability (i.e., lateral facies changes) and paucity of surficial exposures of these deposits has hindered consistent stratigraphic correlations and led to ongoing revisions and redefinitions of areally restricted members and formations. Although the reliance on litho-types for areal correlation is correct stratigraphic procedure, it often ignores described unconformities and has resulted in a proliferation of terminology that may or may not have paleo-environmental, hydrogeologic or chronostratigraphic significance.

Seismic Stratigraphy

Seismic investigations of the study area have been relatively rare. High resolution seismic reflection studies

have been conducted within the Myakka/Peace Rivers and upper Harbor area (Wolansky, et al., 1983) and the Caloosahatchee River/San Carlos Pass area (Missimer and Gardner, 1976) for the purpose of mapping the distribution of shallow aquifers and confining zones. Neither of these studies utilized a grid pattern of seismic profiles such that seismic sequences cannot be mapped in three dimensions from the straight line profiles.

The seismic profiles from both studies were correlated with subsurface lithologic data from wells adjacent to waterways and both indicated four lithologic sequences (from the bottom up): 1) phosphatic limestone, 2) sandy, shelly, limestone and/or sandstone, 3) gray to green calcareous clay with some sand and shell, and 4) fine to medium quartz sand with varying amounts of marl shell, clay and phosphate. Acoustic velocities, which varied from 1.495 m/msec for seawater to 3.350 m/msec for the phosphatic limestones, were calculated from the correlation of seismic reflectors and lithologic unconformities (i.e., depth in meters divided by travel time in msec = velocity in m/msec).

A grid pattern of seismic reflection profiles were collected by Olsen (1980) for a 23 km² area off Captiva Island for the purpose of locating beach compatible sand resources. Although this study covered a very small area, Olsen was able to correlate seismic unconformities with subsurface lithologies from adjacent cores. In water depths of 6-8 m, a shallow limestone was mapped consistently throughout the area at a subbottom depth of ~3 m (9-12 m total depth) and was correlated

with a "medium hard, solution-riddled, fossiliferous limestone". An underlying unconformity surface at 12-15 m was interpreted to have developed on an earlier Pleistocene limestone and contains several localized, erosion channels with depths to 18-25 m (based on acoustic velocities of 1.498 m/msec).

In addition to the few seismic studies conducted in southwest Florida, several studies from outside the immediate study area contain important background information. Mitchum (1978) conducted an extensive seismic study of post-Cretaceous deposits of the west Florida slope using a seismic sparker system with approximately three seconds of penetration and correlated the profiles with sedimentologic and faunal data from seven coreholes. His conclusions pertaining to Cretaceous and Paleogene deposits are summarized in a previous section.

Using standard seismic sequence procedures (i.e., Vail, et al., 1977), Mitchum identified four Miocene depositional sequences (lower, lower-middle, upper-middle, and upper Miocene). Corehole 33-48, which is approximately 170 km west of Charlotte Harbor (approximately 1000 m water depth), contained middle and upper Miocene sediments with a mixed fauna from middle-bathyal (450-900 m) and middle-neritic (30-50 m) sources indicating a significant off-shelf sediment transport.

Oligocene and early Miocene deposits consist of argillaceous to slightly argillaceous carbonates deposited as westward prograding clinoforms. Middle and late Miocene deposition became increasingly rich in terrigenous clays

(decreasing to the south) with late Miocene/Pliocene sequence geometry changing from prograding clinoforms to thin lenses overlying the clinoform slopes.

Mullins et al. (1987; 1988) have refined the work of Mitchum (1978) by creating a high density seismic network with associated piston cores and sediment samples on a small portion of the west Florida slope. Mullins et al. (1987) recognize only three Miocene sequences (composed of discrete, higher frequency depositional cycles) and attribute the mid-Miocene change from prograding clinoform to lense sequence geometries to an intensification of the Loop Current.

Southwest Florida Hydrogeology

The hydrogeology of southwest Florida has been evaluated many times, although usually on a local or site specific scale, i.e. Sproul, et al., (1972; Lehigh Acres area), Boggess (1974; Sanibel Island), Sutcliffe (1975; Charlotte County), Wilson (1977; Hardee/DeSoto Counties), Missimer (1978; Lee County), Wolansky (1983, Charlotte Harbor), and Wedderburn, et al. (1982; Lee County). The hydrostratigraphic classifications of some of those studies are presented in Figure 11.

The Floridan Aquifer (Parker, et al., 1955) or Floridan Aquifer System (FAS; Wedderburn, et al., 1982; Meyer, 1982) consists of several discrete producing zones within the Eocene to Miocene carbonates. Water resource investigations in

southwest Florida have usually ignored the lower Floridan units due to the low water quality (high salinity and total dissolved solids). Utilization of these aquifers for waste disposal and water storage will require extensive research on their geologic and hydrologic properties (Meyer, 1982).

The base of the Floridan Aquifer System is a sequence of impermeable Paleocene evaporites (the Cedar Keys Formation) which is overlain by the cavernous, early Eocene, Oldsmar Limestone. This unit, called the Boulder Zone Aquifer by Kohout (1967), is considered to be in hydraulic connection with the deep ocean along the southern and southeastern margins of the Florida platform (Kohout, 1967; Meyer, 1982).

Kohout (1967) and Kohout et al. (1979) have postulated an inland, convective flow of seawater in the boulder zone that is balanced by seaward, fresh and/or brackish flow in the overlying principal aquifer. This hypothesis is supported by temperature gradients in deep wells across peninsular Florida (Kohout, 1967) and by the existence of geothermal springs with elevated calcium-magnesium ratios along the southwest Florida coast (Kohout, et al., 1979; Fanning, et al., 1981).

These springs have been estimated to flow from depths of 700-900 m based on local geothermal gradients (Kohout, et al., 1979). Wells to similar depths have penetrated this warm (35-38°C), saline aquifer and must be plugged and capped to prevent contamination of overlying aquifers (Sproul et al., 1972). One such site is a free flowing well along the east shore of Charlotte Harbor (Burnt Store area; estimated flow of 2750 gpm;

DEFINED STRATIGRAPHIC UNITS

Sanibel Island

(Boggess, 1974)

Water Table Aq.
Shallow Artesian Aquifer

Charlotte County

(Sutcliffe, 1975)

Water Table Aquifer

Zone 1

Zone 2

Zone 3

Zone 4

Zone 5

Lee County

(Wedderburn, et al, 1982)

Water Table Aquifer

Surficial Aquifer

Sandstone Ag

Mid-Hawthorn Ag

Lower Hawthorn / Tampa Producing Zone

Suwannee Aquifer

Deeper Aquifer

South Florida

(Meyer, 1982)

Surficial Aquifer

Tamiami Aquifer

Upper Hawthorn

Tampa Limestone Upper Floridan

Suwannee Aquifer

Principal Aquifer

Intermediate Aquifers
Boulder Zone

Sarasota/Charlotte Counties

(Wolansky, 1983)

Surficial Aquifer

Tamiami/ Upper Hawthorn

Lower Hawthorn Upper Tampa

Floridan Aquifer

UNDIFF. QUATERNARY
TAMIAMI FORMATION
HAWTHORN FORMATION
TAMPA LIMESTONE
SUWANNEE FORMATION
OCALA GROUP
AVON PARK LIMESTONE
OLDSMAR LIMESTONE

Rosenau et al., 1977).

The middle and upper Eocene deposits consist of several permeable and impermeable zones that are in variable hydraulic connection with the underlying boulder zone and overlying Principal Aquifer through fractures and sinks (Meyer, 1982). Meyer also believes that these aquifers are connected to the ocean, but are not connected to updip areas of freshwater recharge due to the presence of gypsiferous beds under the south-central sinkhole lake region. These marine to brackish aquifers are overlain by impermeable chalky limestones, dolomites and localized evaporites which separate these "intermediate Oldsmar aquifers" from the overlying "Principal Aquifer" (Meyer, 1982).

The Principal Aquifers (or deeper aquifers of Wedderburn, et al., 1982) consist of porous limestones and dolomites of the Ocala Group and the Avon Park and Lake City Limestones. There may be two or more permeable zones within this unit which are combined due to its overall water-bearing properties and the paucity of data on specific aquifer-aquiclude boundaries (Meyer, 1982).

The Suwannee aquifer is separated from the underlying Principal Aquifer by a 15 m thick bed of clay or argillaceous limestone (Wedderburn et al., 1982). Permeability in the Suwannee aquifer is associated with moldic porosity of mollusc-rich zones and occasional sandstones which are separated by clays and/or calcilutite (Wedderburn, et al., 1982). The top of the aquifer is identified as a pronounced attenuation of

gamma profiles due to a confining clay bed (Wedderburn, et al., 1982).

Early and middle Miocene deposits overlying the Suwannee Aquifer consist of interbedded limestones, dolomites, clays, and phosphatic dolosilts that act as a major confining zone for the Floridan aquifer system. The hydrogeologic significance of these units was documented by Back and Hanshaw (1970). Their study compared the hydrogeology of the Yucatan and Florida peninsulas and found that the Yucatan, which is comprised of Cretaceous to Quaternary karstified limestones, has developed a very thin, unconfined freshwater zone over a saltwater aquifer.

As a result of the deposition of relatively impermeable clays, marls, dolosilts, etc., over Paleogene karstified limestones, peninsular Florida has developed relatively higher heads and a deeper flow system (as described above; Back and Hanshaw, 1970). Hanshaw, et al. (1971) have also suggested that the development of the flow systems in concert with fluctuating sea levels has caused dolomitization of calcite and aragonite and hence, the diagenetic alteration of the carbonate platform.

The transitional lithologies which comprise the confining beds of the FAS contain several independent hydrologic units (Figure 13; Parker, et al., 1955; Sproul, et al., 1972; Wedderburn, et al., 1982). Wedderburn, et al., (1982) and Sutcliffe (1975) each describe two producing zones in these strata for Lee and Charlotte Counties, respectively. Boggess (1974; Sanibel Island), Wilson (1977; Hardee/DeSoto Counties)

and Wolansky (1983; Charlotte/Sarasota Counties) indicate only one producing zone within the same deposits.

Permeability within these deposits is located near the formational contacts and is associated with the dissolution and/or reworking of molluscan carbonates (Wedderburn, et al., 1982; Meyer, 1982). Sutcliffe (1975) also states that these artesian systems leak upward in western Charlotte County and flow as springs in the southern part of the County.

Shallow artesian aquifers within the upper Hawthorn and/or Tamiami Formations have been identified in Lee and Charlotte Counties (Figure 11). These aquifers are composed of sands and sandy, shelly limestones that are locally confined by impermeable clay beds (Missimer, 1978; Wedderburn, et al., 1982). Aquifer porosities and permeabilities vary vertically and laterally with lithology and the highest porosities are associated with a Pliocene patch reef facies in southern Lee County (Meeder, 1979; Klinzing, 1982).

Few regional geologic and hydrologic studies have occurred in this area and it is not known if these different interpretations represent real geographic variations in hydrologic or geologic properties, differing stratigraphic interpretations and/or nomenclature, or different tendencies toward grouping or splitting of hydrogeologic units. It is clear that enhanced porosities and permeabilities are associated with specific zones of secondary dissolution, and that these zones are at least partially confined by subsequent changes in the depositional environments.

METHODS

Quaternary Sequences

The Quaternary sequences infilling the study area were obtained from 40 vibracores collected using the procedures and apparatus described by Lanesky *et al* (1979). Vibracore sites were picked to coincide with seismic profiles and locations were determined with LORAN C and fixed reference points. Water depths at core sites are referenced to Mean Low Water (MLW) by correction of measured depths with predicted values (National Oceanic and Atmospheric Administration, 1985).

In the laboratory, the cores were split and described with reference to: color (GSA rock color chart), macro-fossils, visually estimated grain size, bioturbation and root fragments. Seventy-five samples from 10 cores were analyzed for grain size distributions and percent carbonate. Analytical procedures followed Folk (1980) with sieving at 1 phi intervals (-1 to 4 phi) and pipetting at the 5, 7 and 8 phi intervals. Percent carbonate was measured by acidification and point counting of carbonate-quartz grains.

Percent organic carbon and $\delta^{13}\text{C}$ (PDB standard) were analyzed for 33 sub-samples selected from the preceding 75 samples. The subsamples were air dried, ground with a mortar and pestle and sieved to remove coarse CaCO_3 particles. One gram samples were acidified (10% HCl) and combusted following

the procedure of Sackett and Thompson (1963). The resulting CO₂ was measured manometrically to quantify the % organic carbon and analyzed in a Varian MAT spectrograph for the carbon 13/12 isotope ratio.

The depositional sequences defined from the vibracores are compared with modern depositional environments by the statistical interpretation of sedimentological data from 215 surface sediment samples collected and analyzed by Huang (1966). These data were re-analyzed using Q-mode factor and K-means cluster analyses in order to sedimentologically define the modern depositional environments and delineate their distribution. The parameters used for the statistical analyses are: % gravel-sand-silt-clay, mean grain size, standard deviation, skewness, kurtosis, % CaCO₃, % organic carbon, and % organic nitrogen. Analytical procedures and trend surface analysis of the data set have been published as Huang and Goodell (1967).

The factor and cluster analysis procedures of this study were computed using the SYSTAT, Inc. statistical software (SYSTAT, 1983). The principal components factor analysis was calculated from a Pearson correlation matrix and the resulting factors were sorted and rotated using a varimax option. The K-means cluster analysis was used to assess the similarity between stations and verify the groupings established by the factor analysis.

Twenty articulated Chione cancellata and Mercenaria campechensis shells were age dated using the amino acid

racemization (AAR) techniques of Belknap and Wehmiller (1980) and Wehmiller and Belknap (1982). Amino acids may occur in two stereo-isomers (enantiomers) designated as D or L. Living organisms construct protein almost exclusively from the L-enantiomers. AAR measures the amount of change in the ratio of D/L-enantiomers of amino acids found in fossils, which changes with geologic time, from an initial value near 0.0, to an equilibrium value near 1.0.

The rate of change, or racemization, is dependent primarily on temperature, but also varies to a lesser extent with genus and leaching. The D/L ratios can be used independently as a stratigraphic indicator (amino-stratigraphy), but in order to provide a date, must be calibrated with radiometric data. The AAR dates presented in this study are based on the calibrated kinetic model and regional isochrons of Wehmiller and Belknap (1982) which has an uncertainty of $\pm 20\%$.

Seismic Reflection Profiling and Analysis

The primary data base for this study is a network of high resolution seismic reflection profiles. These profiles are created by short duration acoustic pulses generated at set time intervals by a device towed from a moving vessel (Figure 12). The acoustic signals propagate through the water column and subsurface strata as compressional or P waves. A portion of the signal is reflected off the sea bottom and subsurface

horizons and is recorded as time of travel (TT; e.g., elapsed time from signal generation to return of reflected signal) by a towed receiving hydrophone (Figure 12).

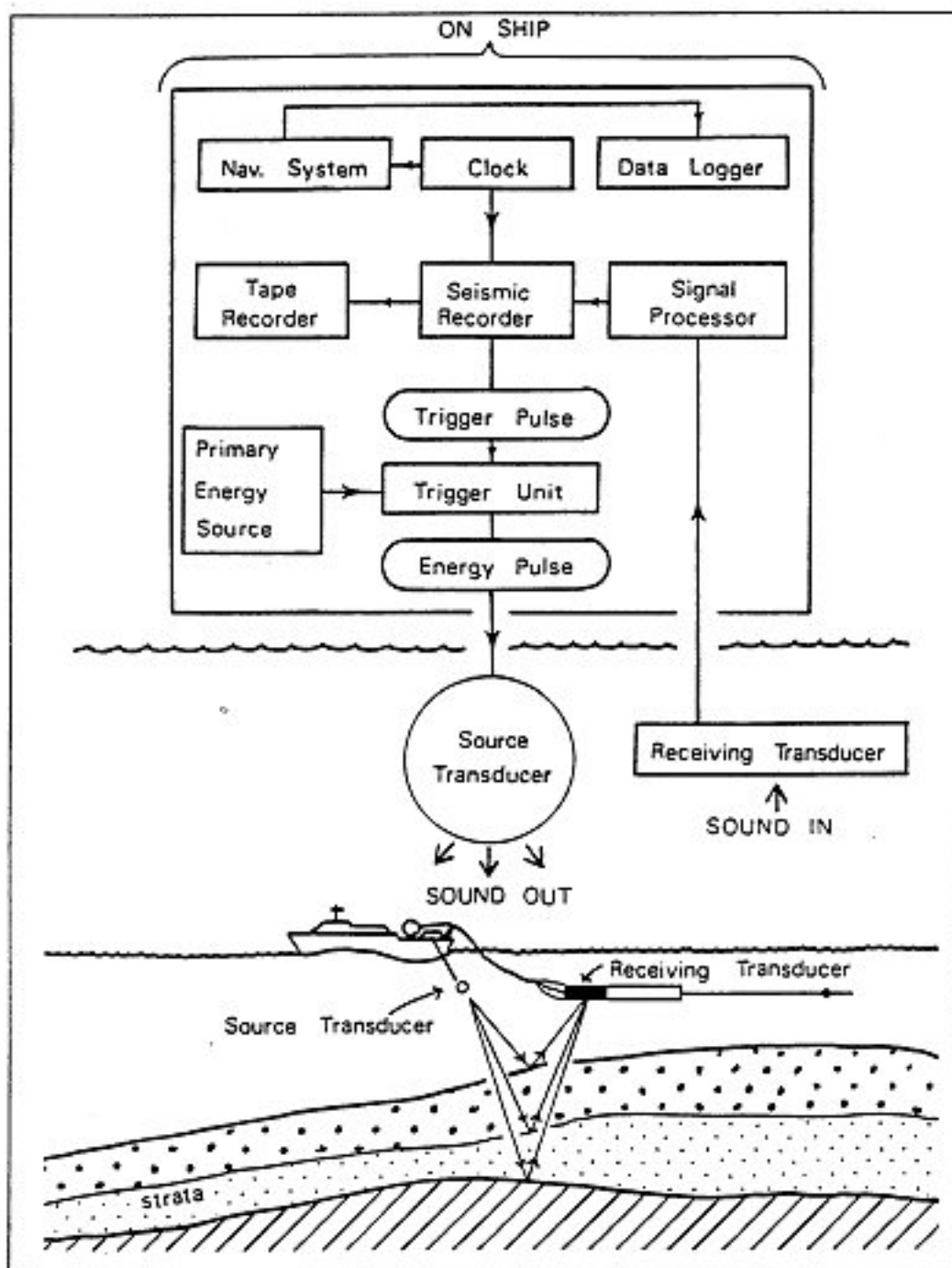
An acoustic wave propagates through water, rocks and sediments at different velocities depending upon the density and elastic properties of each medium (Hamilton, 1970):

$$V_p = [(K+4/3u)/\rho]^{1/2}$$

Where;	V_p = the velocity of the primary, compressional wave in non-porous rock (shear waves will not propagate through water);
	ρ = bulk density of medium;
elastic properties	K = bulk modulus, a measure of resistance to compression;
	u = shear modulus, a measure of rigidity.

When a propagating acoustic (pressure) wave encounters a boundary of different density and/or velocity at a relatively small incident angle, a small amount of the wave energy is reflected back to the surface where it is received by an array of pressure sensitive hydrophones or geophones. The remaining acoustic energy is transmitted through the boundary and continues to propagate downward. Eventually, the wave energy is reduced through multiple reflections and more importantly, by attenuation due to dispersion, heat loss, etc (Dobrin, 1976).

The magnitude and intensity of a recorded reflection event (or reflector) depends on the magnitude of the velocity*density contrast across the boundary. The velocity*density product (ρV) is called the acoustic impedance



and is a measure of the medium's capacity to propagate the acoustic wave. The impedance contrast across a boundary can be quantified as:

$$\text{Reflection coefficient, } R = \frac{p_2 \cdot V_2 - p_1 \cdot V_1}{p_2 \cdot V_2 + p_1 \cdot V_1} \quad \text{Equation 1}$$

where: $p_1 \cdot V_1$ is the impedance of layer 1,

$p_2 \cdot V_2$ is the impedance of an underlying layer 2.

R is a dimensionless quantity between +1 to -1.

An example of a strong reflection is the seawater-seafloor boundary which has an R value of 0.67. The air-water boundary has an R value of 0.99 (V_p air=330 m/sec; V_p water=1500 m/sec). This large R value results in strong reflections across that boundary which causes reverberations or multiple water-bottom reflections. These multiples overprint the real subsurface reflectors at various combinations of the water depth and must be factored out of the data interpretation.

Hamilton (1970) conducted a series of experiments on marine sediments to quantify the relationships between sound velocity and various sedimentological properties, i.e., density, porosity, grain size, mineralogy, etc. His results indicate that mean grain size is the best index property to sound velocity in continental shelf/slope sediments and that clay, porosity, and density are all empirically related to velocity and impedance.

The significance of these results is that the velocity

structure and impedance of marine sediments are directly, albeit empirically, related to commonly measured geologic properties. Therefore, with the exception of multiple reflectors or reverberations, acoustic reflecting surfaces have real geologic significance.

There are a variety of seismic reflection devices used for subsurface analyses. High-frequency tuned transducers use oscillating crystals to produce very narrow range acoustic pulses between 3-7 kHz. Boomer type systems employ high voltage electronic surges to rapidly force 2 metal plates apart and back together to generate an outgoing acoustic pulse. Other commonly used systems include sparkers which also use high voltage surges to vaporize water creating a large air bubble, and air guns which employ pulses of compressed air to create a low frequency acoustic signal.

The specific device used depends on the subsurface penetration and resolution required. High frequency acoustic waves are attenuated much faster than low frequencies such that low frequencies must be used deep penetration (i.e., 1-5 km for large air guns; Dobrin, 1976). The use of low frequencies limits resolution (or the ability to discriminate closely spaced reflectors).

Because $[\text{Frequency} = \text{Velocity}/\text{Wavelength}]$, low frequencies produce long wavelengths (Dobrin, 1976). Resolution is approximately equal to 0.125 to $0.250 \times$ wavelength (Sheriff, 1977). The boomer type devices used in this study produce a broad frequency acoustic pulse centered

around 4000 kHz. Assuming a velocity of 1700 m/sec for unconsolidated sediments, the theoretical resolution of these systems is 5-10 cm (frequency = V_p/WL , or $WL = V_p/freq.$; $1700/4000 = 0.425$ m/cycle $\times$ 0.125 to 0.250 = 0.053 to 0.106 m). The physical limitations of the graphic recorder and practical considerations of velocity structure and interpretation reduce the resolving ability to 50-80 cm.

Two different high resolution profiling systems were used in this study. An E.G. & G. "Uniboom" was used for profiles CH-01 to CH-08 and fired at a 200 joule power setting and a 0.10 second sweep. Profiles CH-09 to CH-28 were collected using an O.R.E. "Geopulse" at 175 joules and a 0.125 second sweep. Navigational fixes were obtained at course changes and/or 5 minute intervals using LORAN C and adjusted to fixed reference points (channel markers, etc).

The seismic data were interpreted using the stratigraphic procedures of Vail, et al. (1977), Brown and Fisher (1980), and Vail (1987). Lithology or impedance based seismic reflections are one of three possible types (Vail, 1987): 1) reflections that follow synchronous geologic time lines, 2) unconformities or downlap surfaces, and 3) rare reflections caused by fluid interfaces or diagenetic surfaces.

The basic analytical steps of seismic stratigraphy are; 1) sequence analysis-definition of sequence unconformity bounded depositional sequences, 2) integration of seismic data with geologic data from wells and outcrops, 3) facies analysis-reflector geometry, continuity, etc., within sequences, 4)

reflection character analysis, i.e., amplitude, frequency and
5) creation of a geologic model.

A depositional sequence is defined as:

-A stratigraphic unit composed of a relatively conformable succession of genetically related strata.

-The upper and lower boundaries of a depositional sequence are unconformities or their correlative conformable surfaces (Brown and Fisher, 1980).

Sequence bounding unconformities may be classed as:

-Type 1 sequence boundaries; a regional surface characterized by subaerial exposure and erosion associated with stream rejuvenation, a basinward shift of facies, a downward shift in coastal onlap, and onlap of overlying strata...

-Type 2 sequence boundaries; a regional surface marked by subaerial exposure and a downward shift in coastal onlap landward of the depositional-shoreline break; however, it lacks both subaerial erosion associated with stream rejuvenation and a basinward shift in facies (Vail, 1987).

These two types of boundaries lead to type 1 and type 2 sequences which are defined by the nature of the basal unconformity; i.e., a type 1 sequence is bounded at the base by a type 1 boundary. The sequences may also be subdivided into systems tracts or "contemporaneous depositional systems" (Brown and Fisher, 1980) based on changes in reflector terminations along a bounding unconformity (onlap and downlap), position within a sequence and stacking patterns (Vail, 1987). Although the system tract terminology, which is derived from the study of continental margins, is not specifically applicable to the paralic sequences defined in this study, the procedure does

indicate that sequences can be subdivided into consistent "tracts" which indicate a relative, landward or basinward shift of the sequence depocenter.

After the sequences have been defined and mapped (step 1), they must be calibrated with subsurface lithologic data and the seismic time-of-travel scale converted to depth via velocity interpretation (step 2). The procedures for calculating seismic velocities and lithologic description are presented in the Well and Borehole Data section. The third step in the analysis of seismic data is the definition and mapping of seismic facies (step 3). Seismic facies are defined as:

-An areally definable, 3-dimensional unit composed of seismic reflections whose elements, such as, reflection configuration, amplitude, continuity, frequency and interval velocity, differ from the elements of adjacent facies units.

-A seismic facies unit is interpreted to express certain lithology, stratification and depositional features of the deposits that generate the reflections in the unit.

(Brown and Fisher, 1980)

The four principal types of seismic facies are parallel/divergent, progradational, mounded/draped, and onlap/fills. Examples and the geologic implications of each of those types are described in detail by Vail et al (1977), Roskandic (1978), and Brown and Fisher (1980). In this study facies analysis consists of qualitative observations of the distribution of parallel, prograding, onlap, and channel fill

type units.

The fourth step involves reflection character analysis, or the examination of the frequency and magnitude of the individual waveforms which are not resolvable on high resolution records. Consequently, the fourth and final step in this study is the development of a geological model which integrates and correlates the seismic and lithologic sequences into a predictable pattern of depositional sequences (step 4). Although these procedures were developed for low frequency seismic exploration of continental shelf/slope systems, the principles are applicable to the high resolution analysis of coastal and inner shelf systems (Hine and Riggs, 1986).

Well and Corehole Data

Lithologic and geophysical logs from 22 sites around Charlotte Harbor are used to quantify the seismic model. Seventeen of these sites are from existing wells drilled and logged by the U.S. Geological Survey, the Southwest Florida Water Management District and the South Florida Water Management District. Three cores drilled by the Florida Geologic Survey were re-described from the archives at Tallahassee, Florida on the basis of color, visual grain size, and structure.

Two wells were drilled for this study using a rotary drill rig with a 16 cm bit. Cuttings from those wells were described during drilling and later re-examined with a binocular microscope for color, macro-fossils, gross mineralogy

(quartz sand, phosphatic sand, limestone etc.) and visual grain size . Graphic lithologic logs for all sites were redrawn from descriptive logs to ensure consistency of nomenclature and graphic symbols. Geophysical logs were evaluated at two foot (or half scale) intervals and replotted using Lotus 123 software to standardize horizontal and vertical scales.

In order to directly compare seismic and wellhole data it is necessary to translate the vertical time scale of seismic data to true measured depth (i.e., feet or meters). This conversion is possible if the velocity structure of the subsurface strata is known; $\text{distance (meters)} = \text{velocity (meters/second)} \times \text{time (seconds)}$.

If subsurface velocities and densities are known, it is possible to construct a "synthetic seismogram" from the pV profiles using equation (1) which defines the reflection coefficient R . Two of the wells used in this study were drilled for the express purpose of obtaining velocity (sonic) and density ($\gamma\gamma$) geophysical logs.

The sonic log records Dt , or the time required for a compressional sound wave to traverse one foot of formation versus depth. Dt is called the interval transit time and is the reciprocal of the sound wave velocity (Schlumberger, 1972a/b; Doveton, 1986). The sonic sonde employs two transmitters (and two receivers) which are alternately pulsed and averaged. In sedimentary formations, Dt depends on several parameters, principally, the matrix material and porosity (Doveton, 1986). Examples of log velocities and commonly used

seismic model velocities are presented in Table 2.

Weak signals are sometimes strong enough to trigger the near receiver but are not recorded at the far receiver. When this occurs, the sonic curve shows a large, abrupt excursion toward higher Dt values. This anomaly is known as cycle skipping and is likely to occur where the signal is strongly attenuated by unconsolidated formations, formation fractures, gas, aerated mud or rugose salt sections (Schlumberger, 1972a).

There is a wide range in the velocities of lithified rocks evaluated from sonic logs (i.e., sandstones, and limestones, 5400-7000 m/msec) and the unconsolidated deposits evaluated with laboratory instrumentation (i.e., sand and clay; 1490-1836 m/msec; Hamilton, 1970). These velocity differences are attributed primarily to the effects of burial pressure, lithification and cementation (Schlumberger, 1972a/b; Dobrin, 1976; Doveton, 1986).

The density or gamma-gamma log emits medium energy gamma rays (cesium 137 source) into the formation. These gamma rays collide with the electrons in the formation which causes an energy loss and back-scattering of the emitted gamma rays (Schlumberger, 1972a; Doveton, 1986). A detector within the density sonde measures the number of backscattered gamma particles which is related to the electron density of the formation. Electron density is related to true bulk density which is, in turn related to matrix material, porosity, and fluid density (Schlumberger, 1972b).

The influence of the mud column is minimized by

eccentering the sonde against the borehole wall. The sonde used in this study is not calibrated in standard API limestone porosity or bulk density values such that all density logs are presented as counts per second (cps) and represent uncorrected, relative density values.

Acoustic impedance logs were created from the sonic and density logs by dividing the density log (cps) values by the Dt values (reciprocal velocity). Reflection coefficient logs were then created from the impedance logs using equation (2) which compares the impedance from an upper unit with that of a lower unit;

$$R = (p_2 V_2 - p_1 V_1) / (p_2 V_2 + p_1 V_1) .$$

A true synthetic seismogram would "convolve" (or integrate) the reflection coefficient log with an estimate of the outgoing seismic signal in order to produce the seismic profile. In this study, the reflection coefficient logs are compared directly with adjacent seismic profiles and the mapped seismic sequences.

Natural gamma logs are also extensively used in this study because of their value in correlating formations in cased or uncased holes and because they have been routinely collected in various monitoring and test wells. The gamma log is a measure of the natural radioactivity of the formations (Schlumberger, 1972a). Gamma rays are bursts of high energy electromagnetic waves which are emitted by potassium isotopes and elements of the uranium and thorium decay series.

Uranium occurs via substitution, as a persistent, usually

minor component in the phosphatic deposits of peninsular Florida (Altschuler et al, 1964). The gamma rays are capable of penetrating steel or plastic casing. Although the casing attenuates the gamma rays by about 30%, the profile retains its shape and value for correlation (Doveton, 1986). Most of the logs used in this study are uncalibrated with respect to the API standards and record counts per second (cps) vs. depth.

Table 2. Acoustic Velocities of Selected Materials

Material	Velocity (m/msec)	Density	Dt (usec/ft)
Sandstones*	5488-5945	2.65 (Quartz)	55-51
Limestones*	6402-7012	2.71 (Calcite)	47.5
Dolomite*	7012	2.87	43.5
Freshwater* (including drilling mud)	1500	1.00-1.10	189
Clay (Marine)**	1490	1.40	--
Clayey silt ** (Marine)	1533	1.30	--
Sand ** (Marine; fine to coarse)	1711-1836	2.72	--

* Velocities calculated from sonic log velocities, all units lithified (SCHLUMBERGER, 1972).

** Velocities measured directly from unconsolidated sediments (Hamilton, 1970).

RESULTS

Seismic Stratigraphy

Regional Unconformities

The locations of the seismic profiles collected and interpreted for this study are presented in Figure 13. This figure also includes the locations of all profile sections presented as raw or interpreted data in the figures or appendices. Two regional unconformities (or reflectors) which bound six depositional sequences and five intervening disconformities have been identified from the seismic profiles. The depositional sequence nomenclature to identify these features is presented in Table 3.

Reflectors which can be correlated over the entire study area and show distinct erosional unconformity are noted by an "R" number with the sediment-water interface being R-1. Underlying regional reflectors are labeled as R-2 and R-3 with an open-ended classification which can be appended with later studies. Depositional sequences are labeled with letters and begin with A as the lowermost sequence following geological convention. Disconformities separating sequences do not truncate underlying strata and are numbered from the bottom up, i.e., 3-1, 3-2, 3-3, (indicating relationship with the R-3 unconformity).

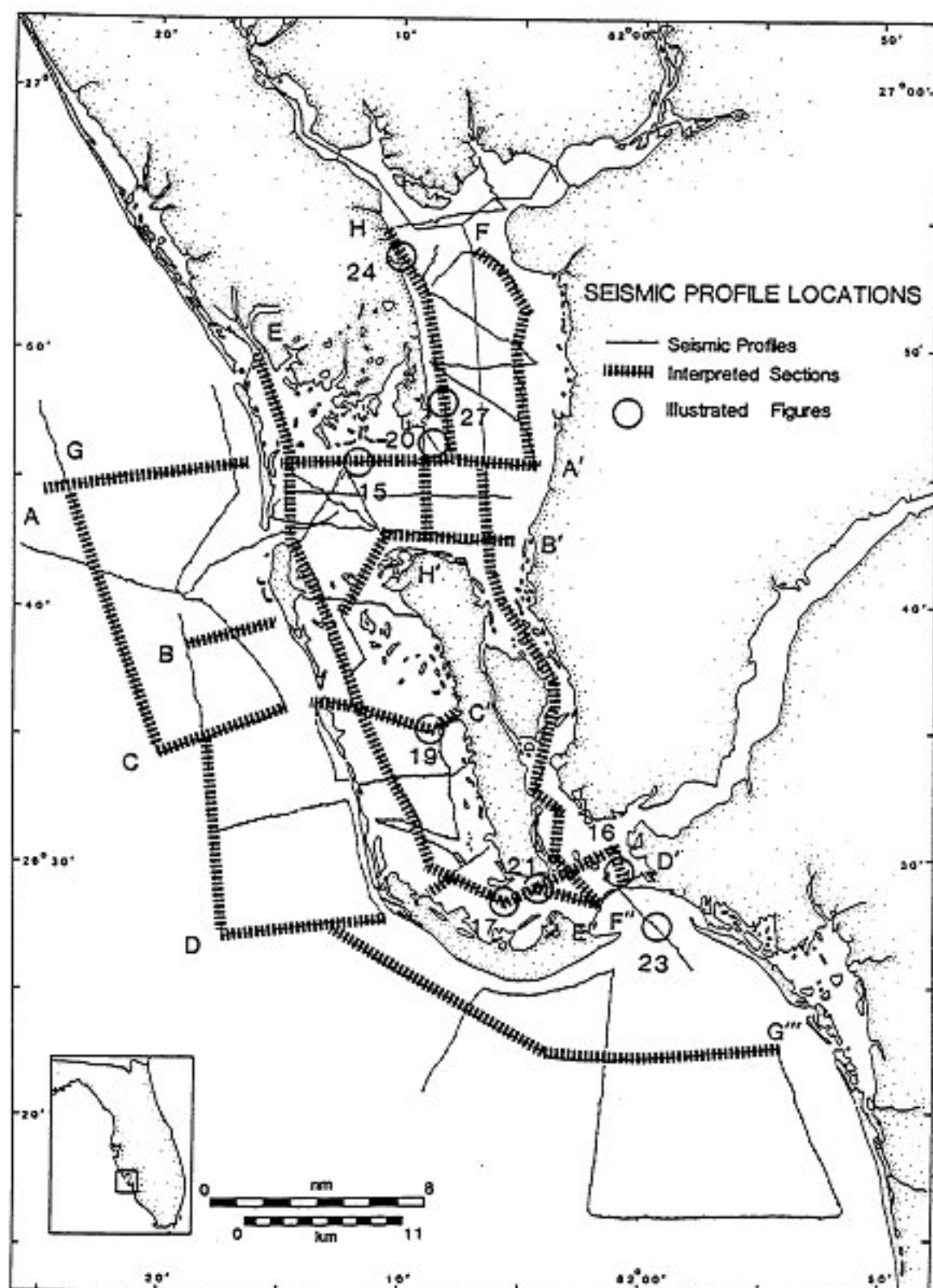


Table 3. Sequence and Reflector Nomenclature

<u>Regional Unconformities</u>	<u>Local Disconformities</u>	<u>Sequences</u>
R-2 - - - - - - - - -		
	3-5	F
	3-4	E
	3-3	D
	3-2	C
	3-1	B
R-3 - - - - - - - - -		A

The lowermost regional unconformity (R-3) represents the limit of acoustic penetration throughout most of the study area. Figure 14 is a structure contour of the R-3 surface with depth denoted as two-way travel time (2-way TT) in milliseconds (msec). Using an assumed velocity of 2.0 m/msec (Table 2; 1.7 m/msec for unconsolidated Quaternary deposits), and dividing the 2-way TT by 2, for 1-way TT, the depths in 2-way TT are equal to depth in meters (i.e., 100 msec x 2.0 m/msec x 0.5 = 100 m).

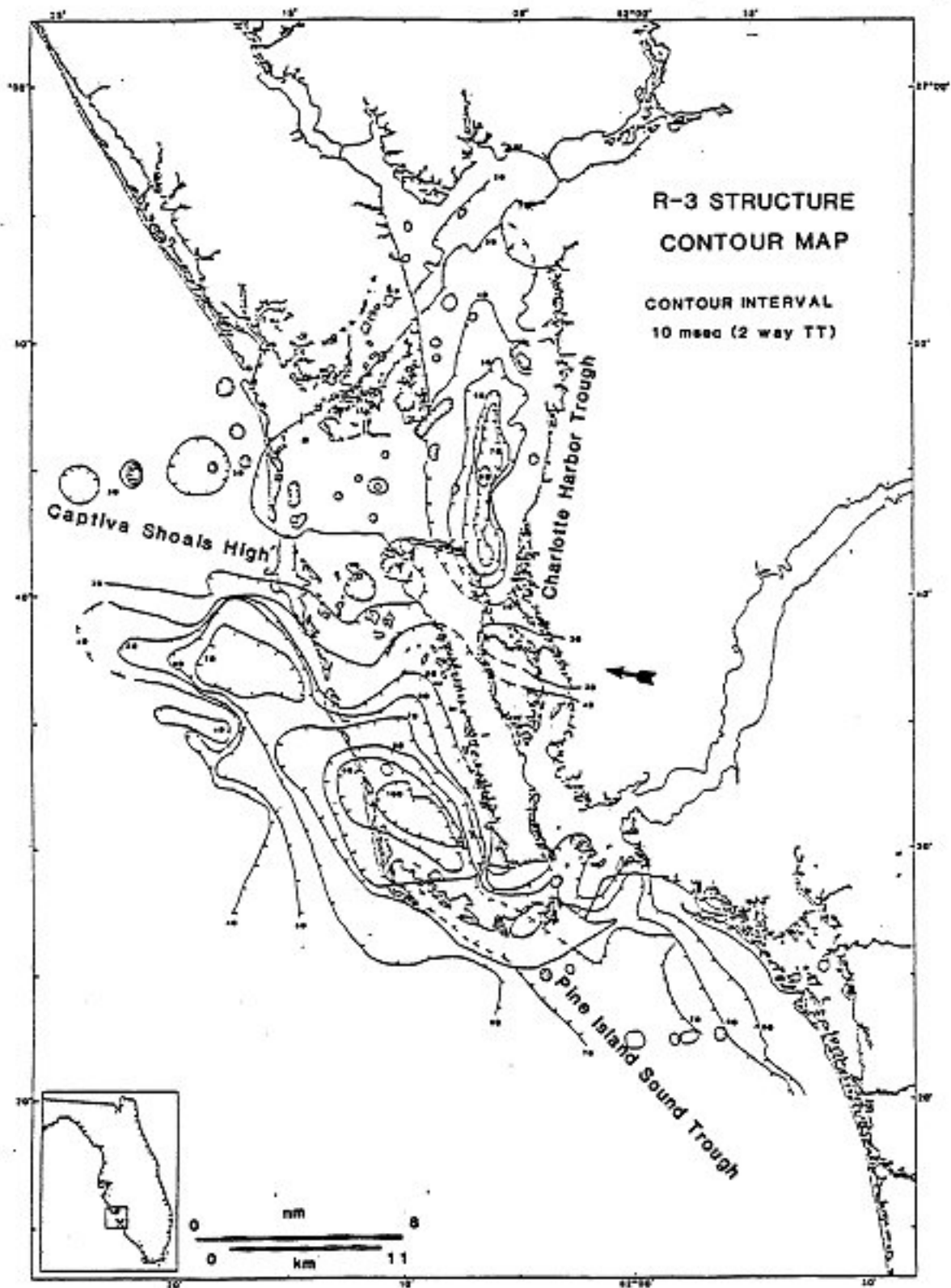
The depth of the R-3 unconformity varies from less than 20 msec to more than 100 msec (Figure 14). The lows in the R-3 structure contour are aligned as a broad system of NE-SW and NW-SE trending troughs. The trough system reaches its maximum depth under Pine Island Sound (115 msec 2-way TT) and continues offshore to the NW under Sanibel and Captiva Islands and extends southeasterly under southern Sanibel Island and San

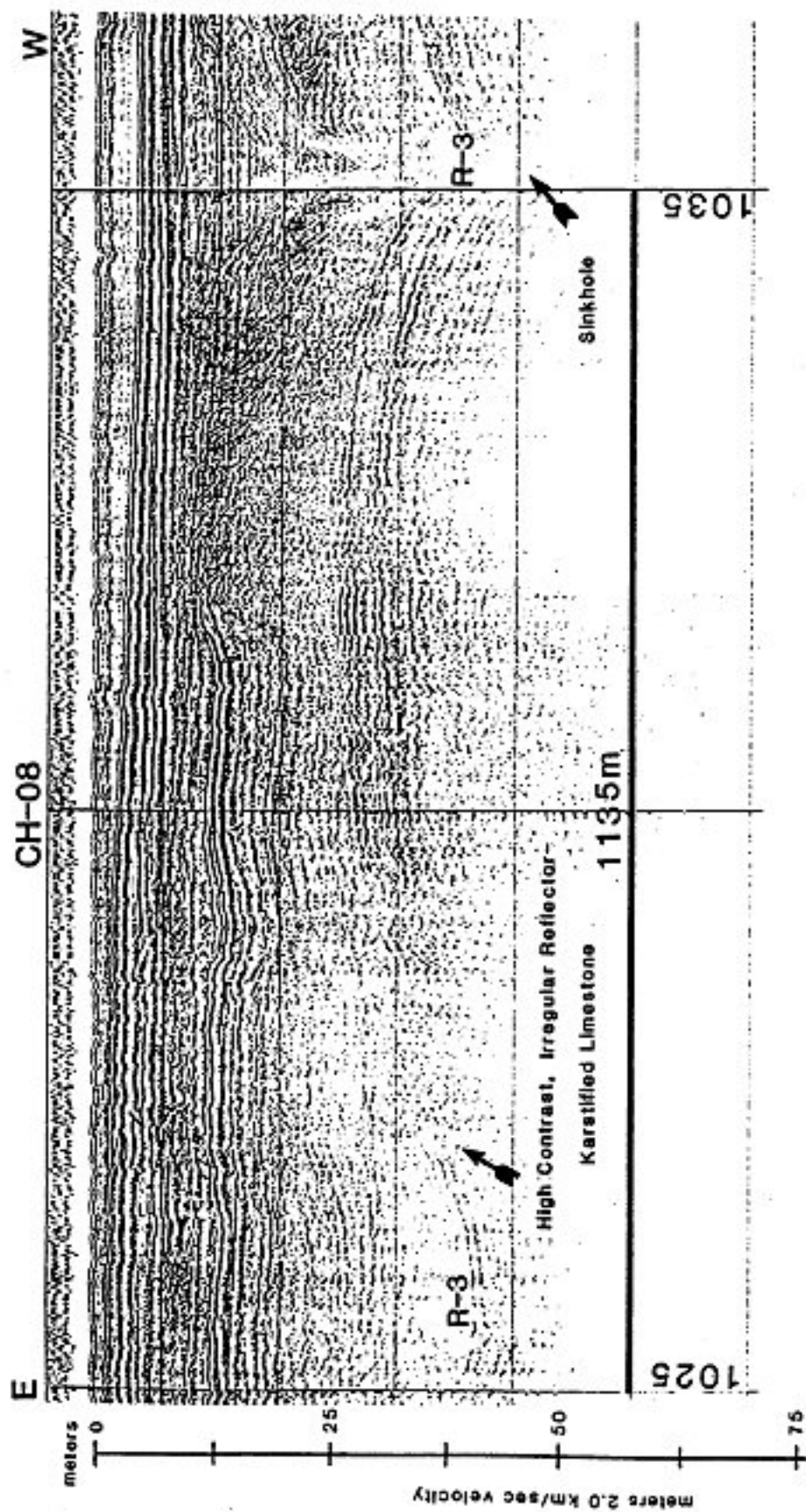
Carlos Bay.

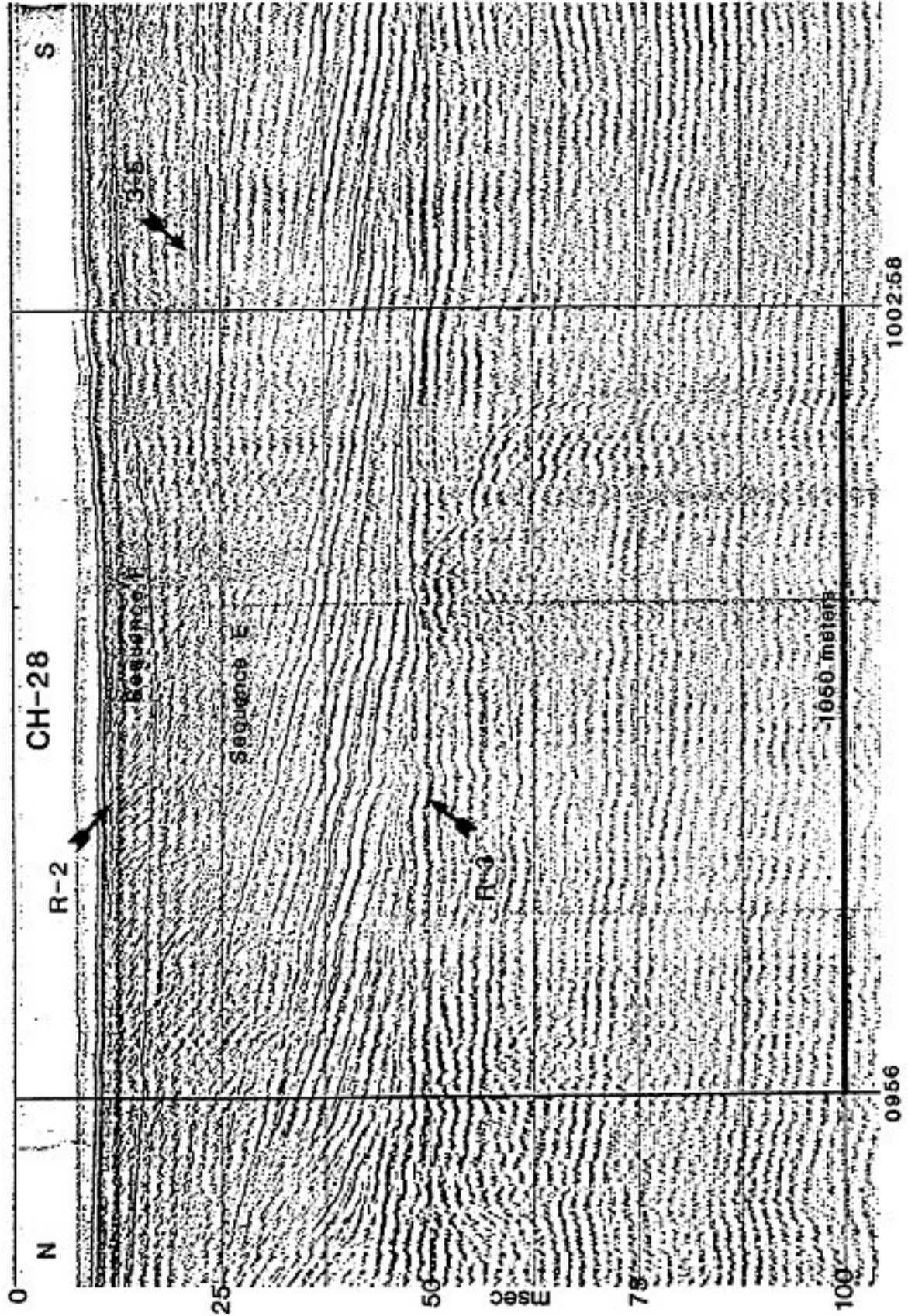
A north to northwest trending trough also underlies the upper and lower Harbor area with maximum observed depths of 85 msec. A structural high apparently separates the Pine Island Sound trough from the Harbor trough. Although the Pine Island land mass reduces the water-based seismic coverage in this area, the structural high is present on both sides of Pine Island and is coincident with the Captiva shoals and Little Pine Island.

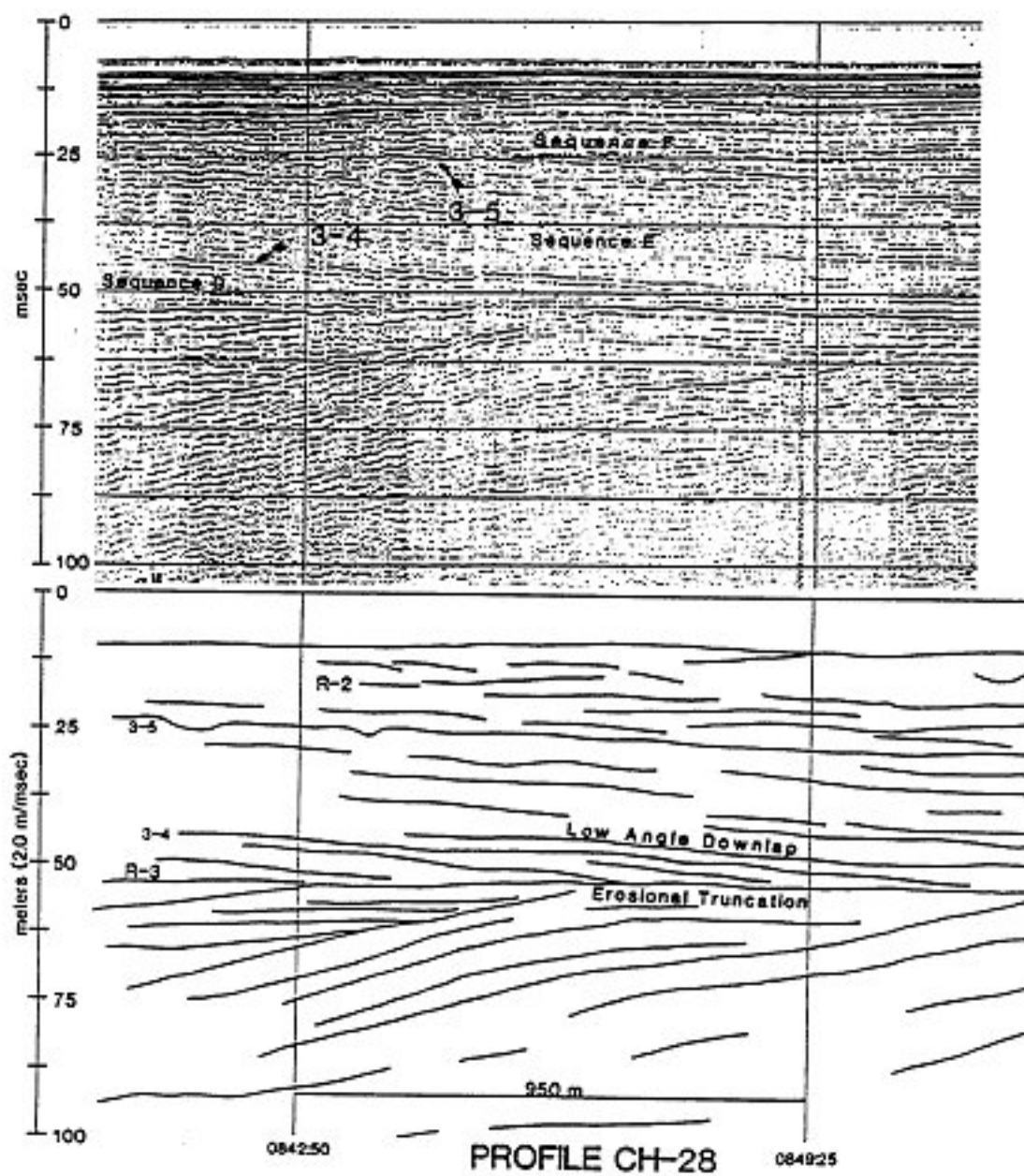
The character of the R-3 reflector is also very different between the high and low areas. Figure 15 is a portion of seismic profile CH-08 from the lower Harbor which shows the R-3 unconformity as a strong, irregular surface with no penetration of underlying strata. Overlying stratal reflectors downlap at relatively high angles onto the R-3 such that the unconformity is easily correlated between profiles.

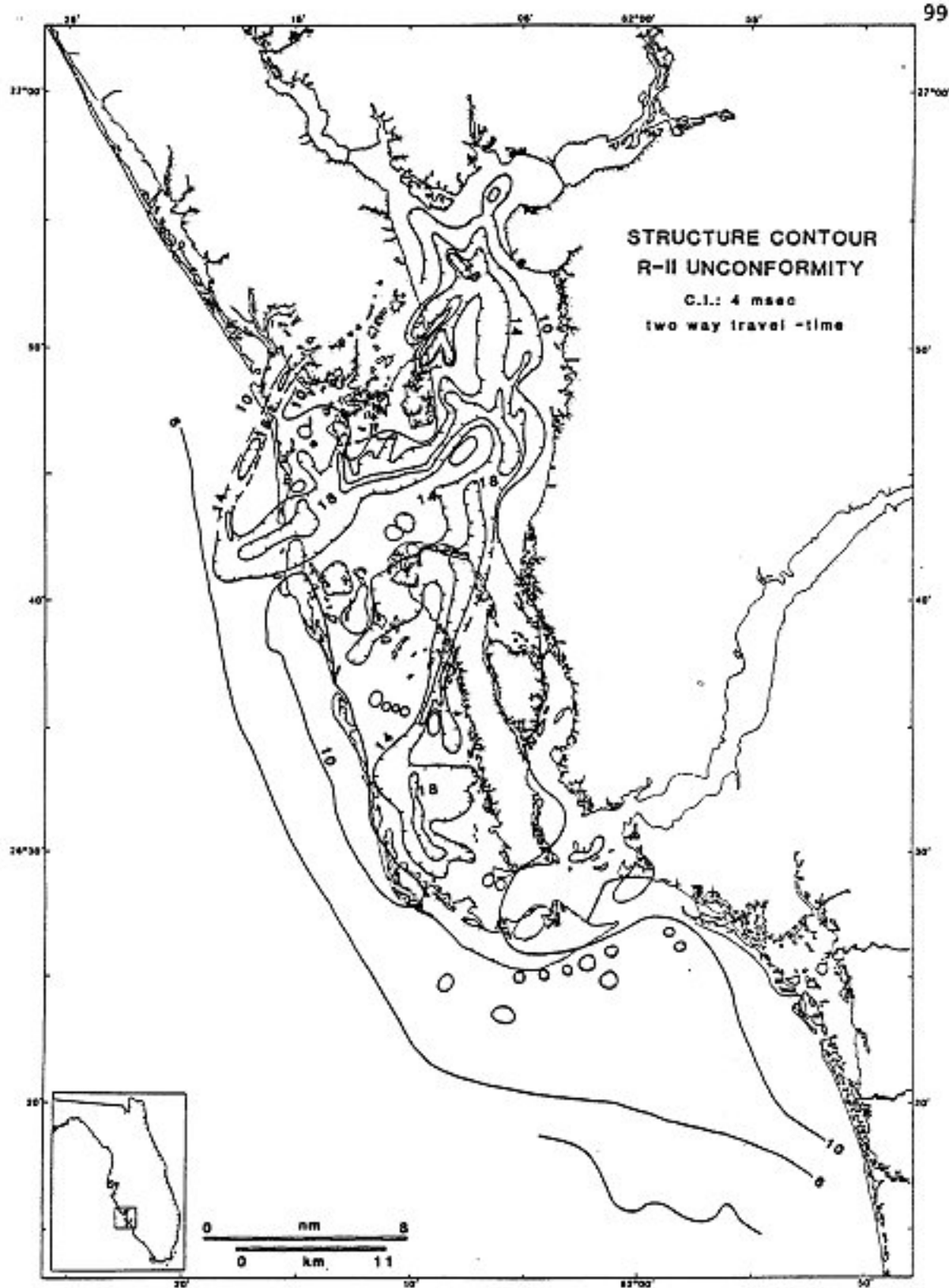
Figure 16 shows the R-3 surface within a structural low (CH-28; southern Pine Island Sound). Here, the R-3 reflector is continuous and moderately strong with the overlying stratal reflectors showing flat-lying, continuous aggradation onto the R-3 surface. Despite the very low angle offlap or aggradation of sequence E reflectors within the structural lows, the R-3 surface still represents a strong erosional unconformity as indicated by the angular unconformity in Figure 17 (CH-28; southern Pine Island Sound).

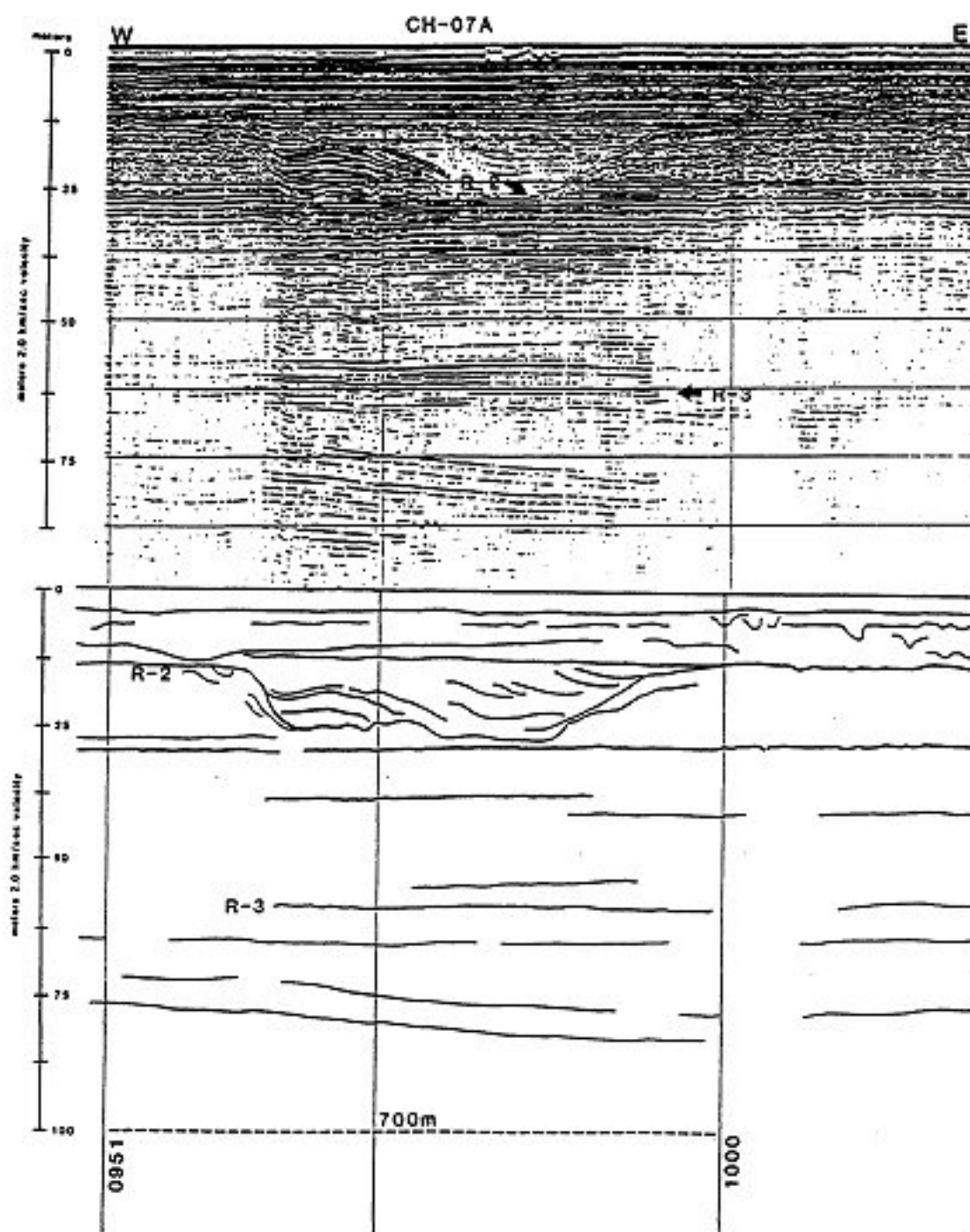












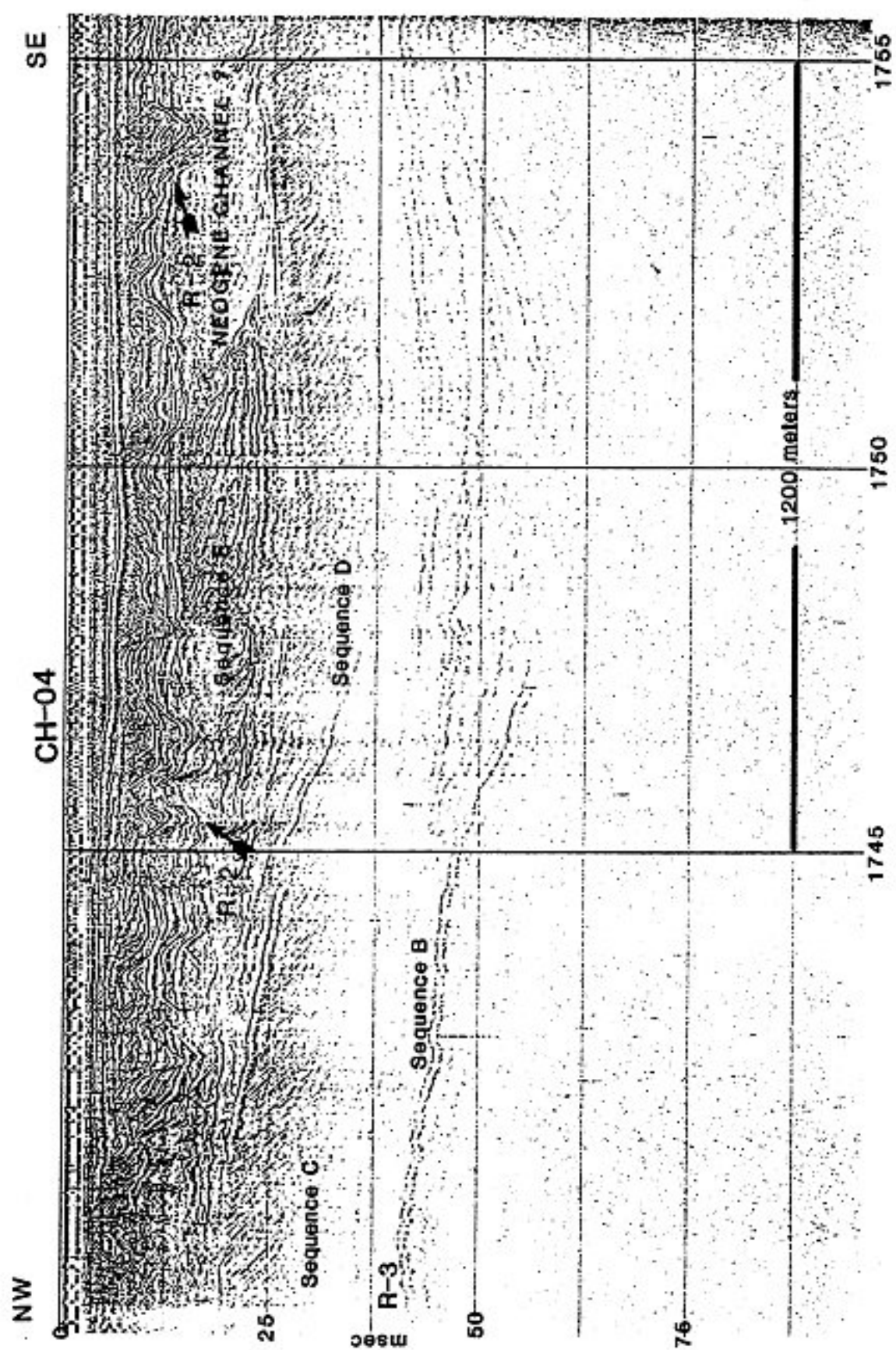


Figure 18 is a structure contour map of the R-2 unconformity and indicates that structural lows of this surface mimic the overall trend of the underlying R-3 surface. The broad troughs of the R-3 surface have been mostly infilled and the remaining relief is present as narrow channels (note the change in scale between Figures 14 and 21, from 10 msec contours to 4 msec contours). The channelized appearance of this relief from the structure contour is indicated by Figure 19 (CH-07; central Pine Island Sound) which shows the R-2 surface as a distinct downcutting channel.

Within the upper Harbor the R-2 surface is represented by multiple channel structures that cannot be individually correlated with the present profile grid (Figure 20, CH-04; upper Charlotte Harbor). These channel structures probably represent a combination of synchronous channel erosion and migration, and aperiodic, sea level controlled, channel reactivation.

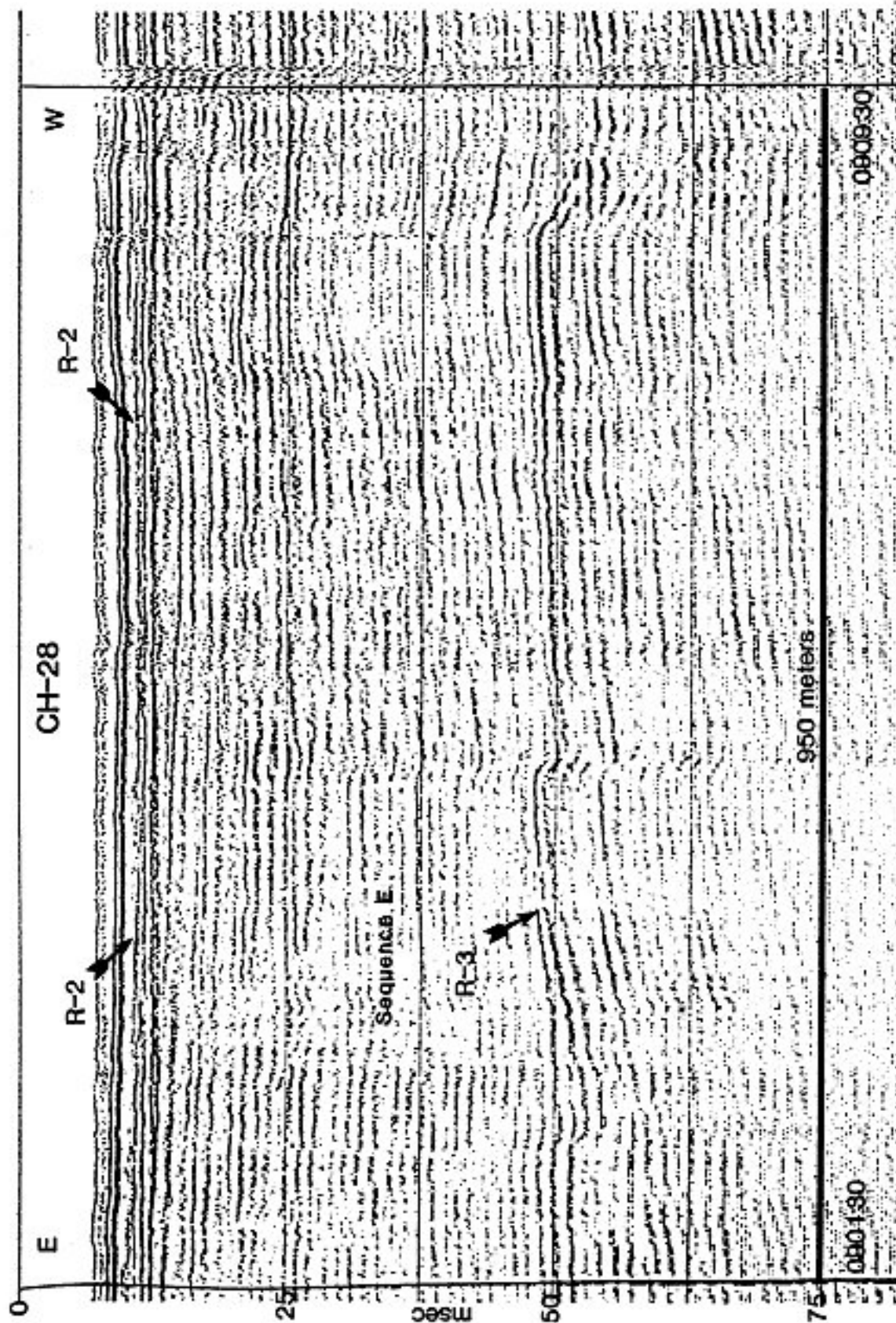
Seaward of the modern barrier system, the R-2 surface occurs as a strong, continuous, flat lying unconformity which sharply truncates the underlying sequences (Figure 21; CH-09, seaward of Boca Grande Pass). Using a seismic velocity of 1.7 m/msec for the unconsolidated Quaternary deposits (Table 2), this portion of the R-2 unconformity can be correlated with a "medium hard, solution riddled, fossiliferous limestone" described by the U.S. Department of the Army (Corps of Engineers, 1969). The geometry of this unconformity suggests that it represents a ravinement or zone of shoreface erosion

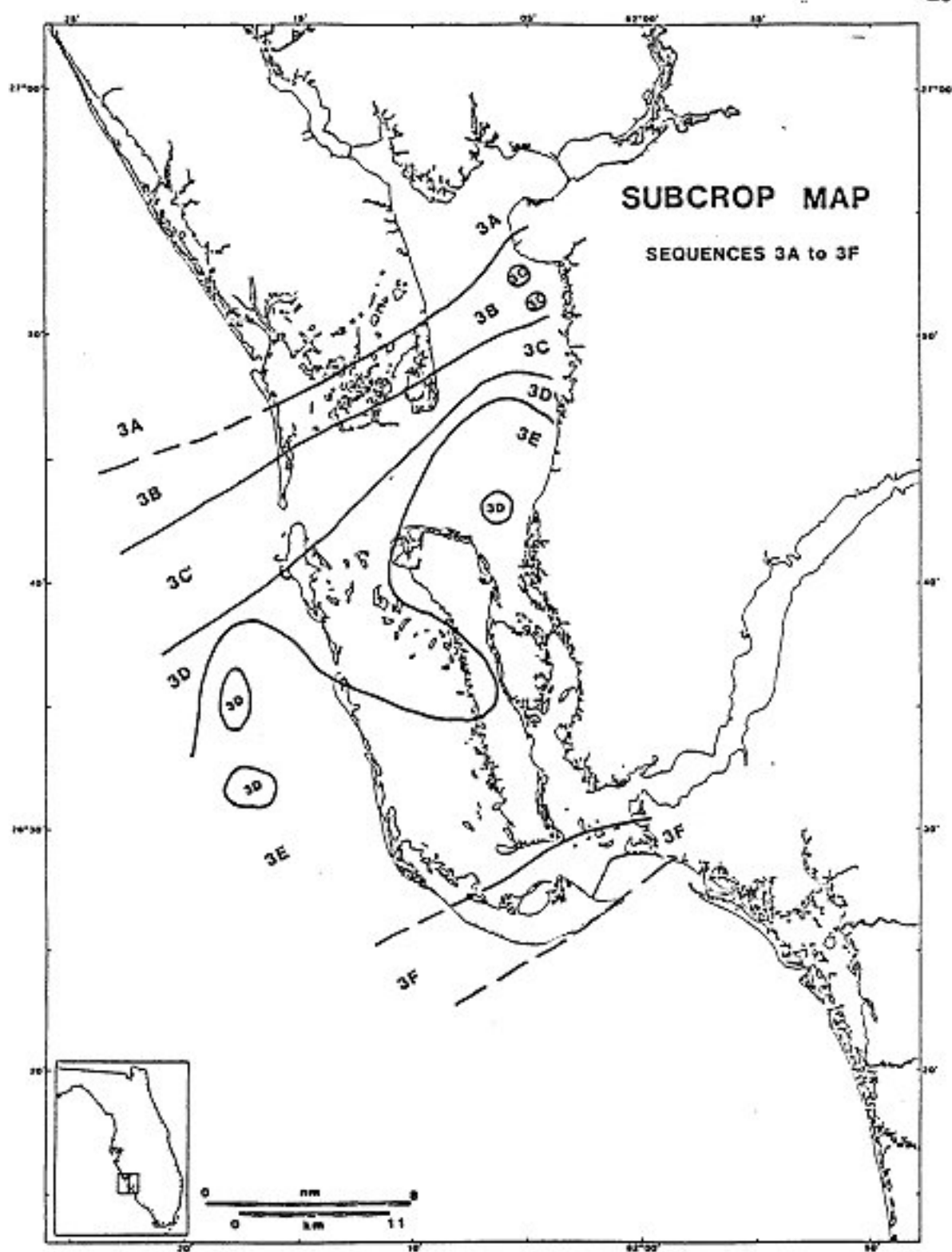
(Nummedal and Swift, 1987).

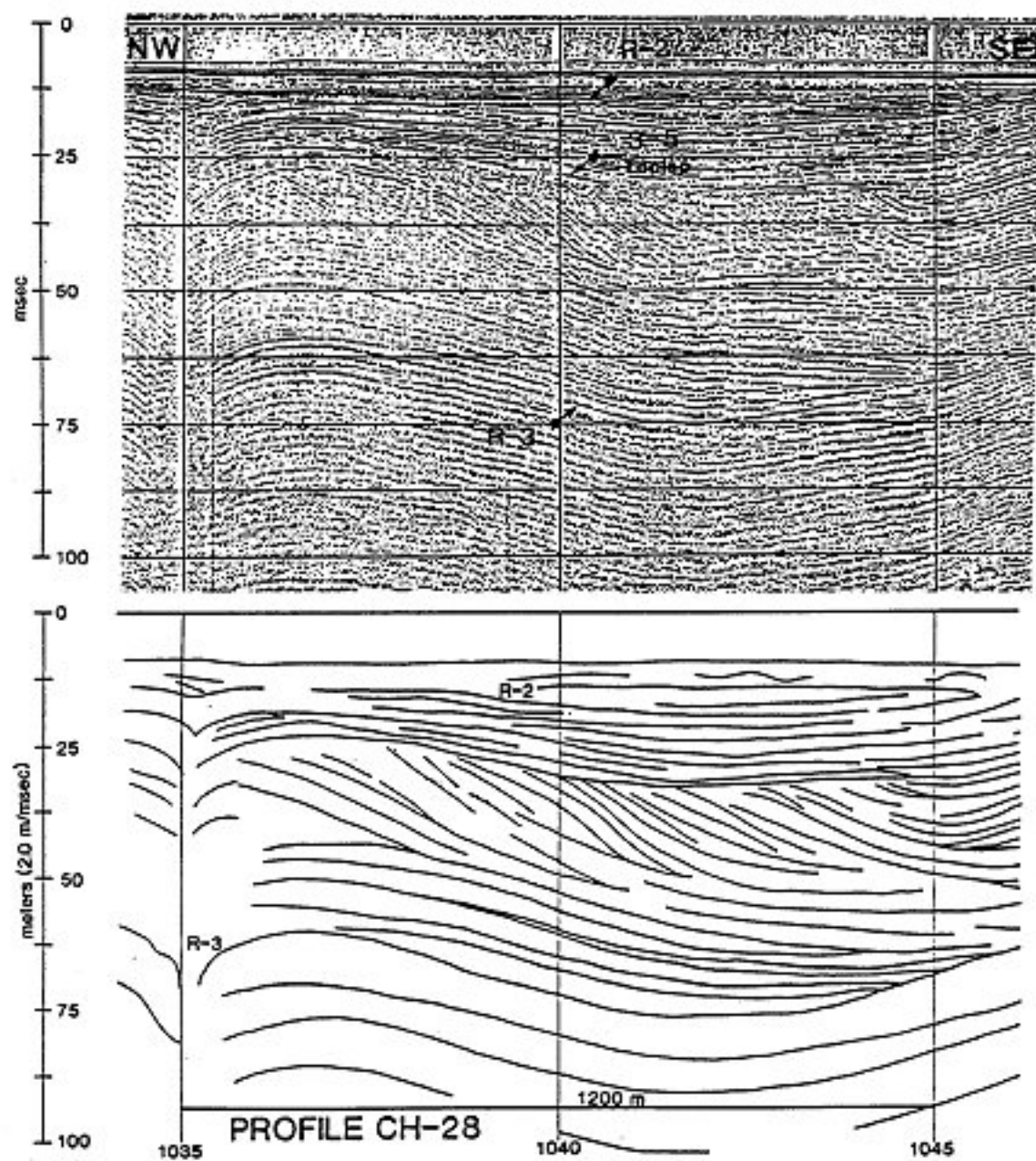
Depositional Sequences

Six depositional sequences have been identified between the two regional unconformities. The updip limit of each of these sequences (labeled 3A-F; from the bottom up) is presented as a subcrop map in Figure 22. These units, and the stratal reflectors in each sequence, dip to the southeast with the exception of Sequence 3E. Selective deposition within the structural lows and/or subsequent erosion from the Captiva Shoals/Little Pine Island high has resulted in an anomalous subcrop distribution for Sequence 3E, relative to the generally linear pattern of the other sequences (Figure 22).

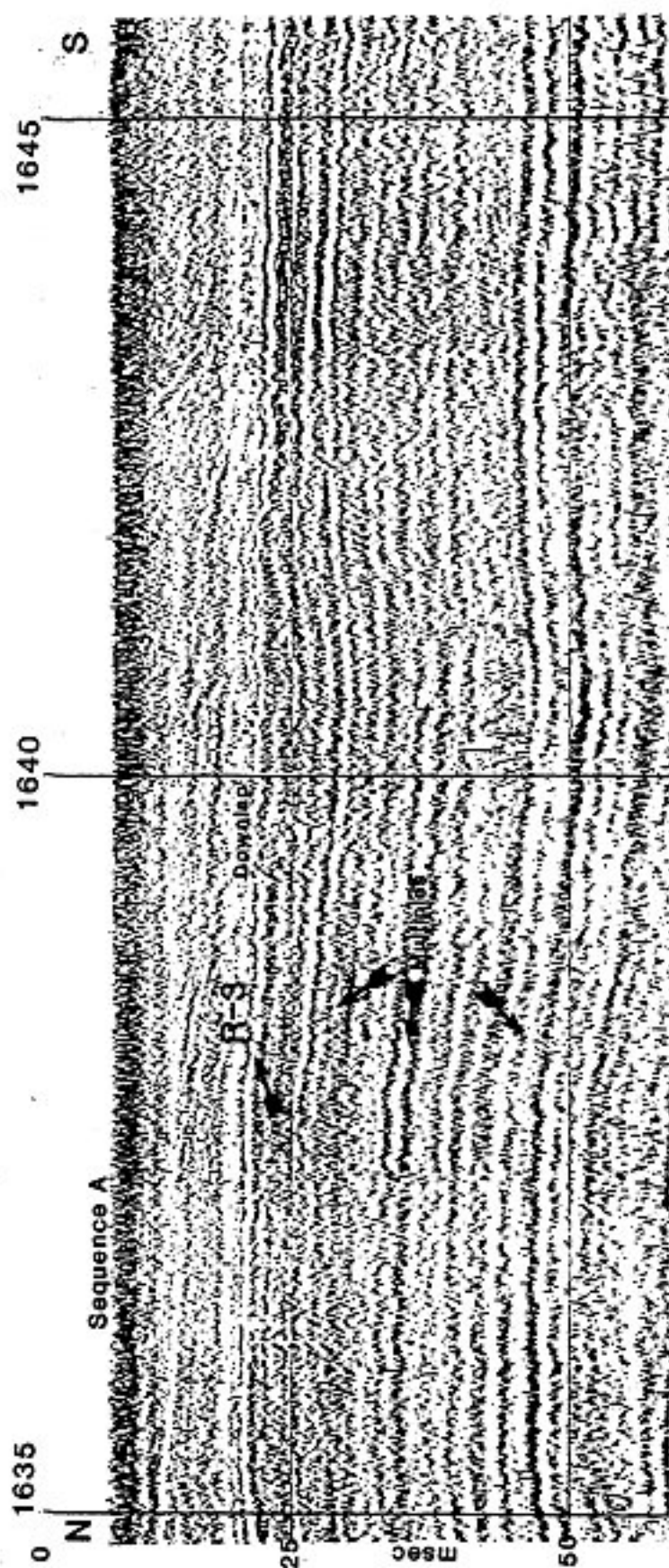
It is possible, and likely, that additional sequences occur between the two regional unconformities outside of the study area. The recognition of Sequence 3F is based on non-conformable sequence boundaries on short sections of only three profiles (CH-21, 27 and 28), and can only be tentatively mapped. The updip limit of this sequence is illustrated in Figure 23, and indicates a pattern of opposing reflector orientations and terminations relative to the underlying 3E unit (i.e., short steep clinoforms-sequence F vs. long, low angle clinoforms-sequence E). However, the major body of the 3F unit probably continues to the south and may underlie additional depositional sequences.

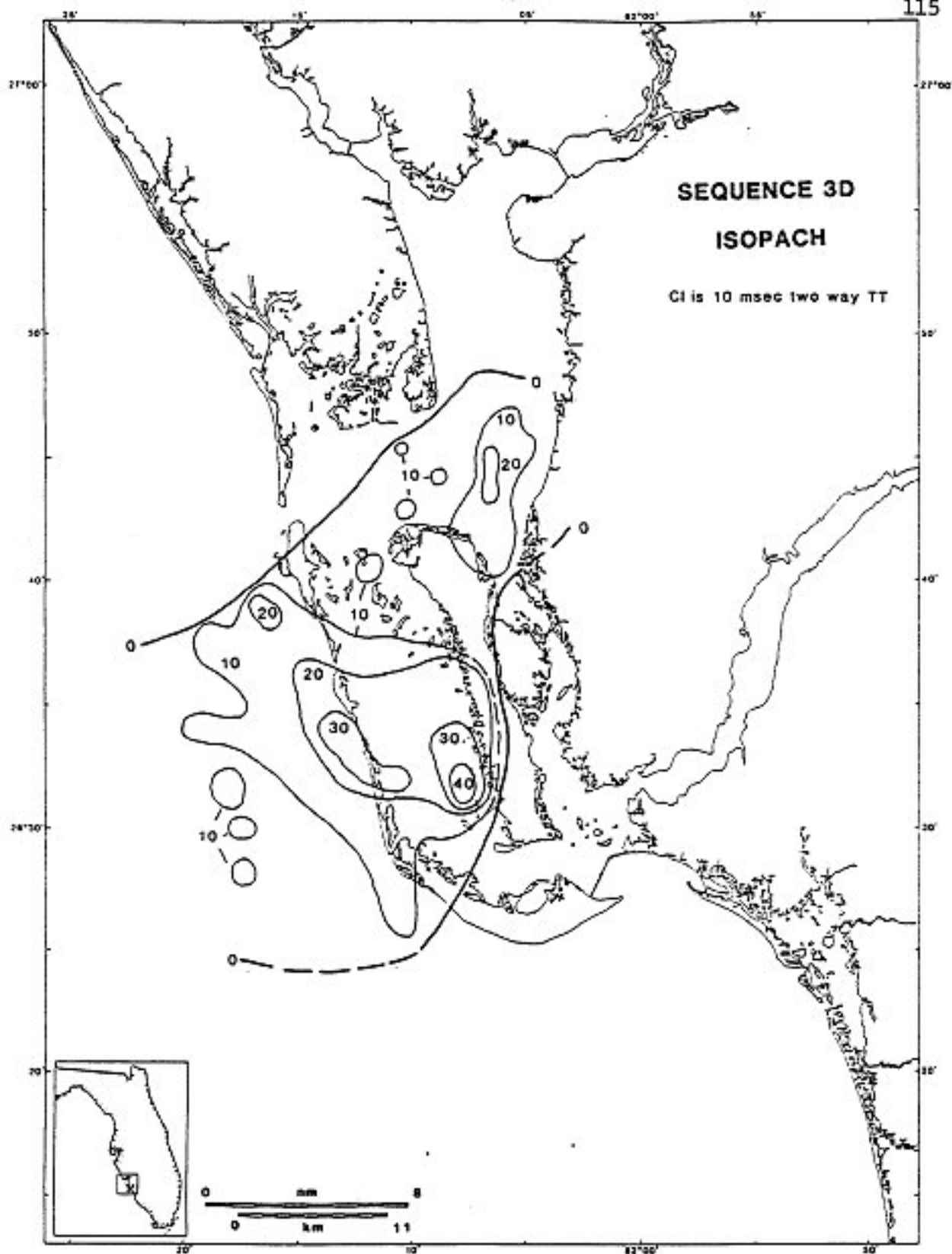


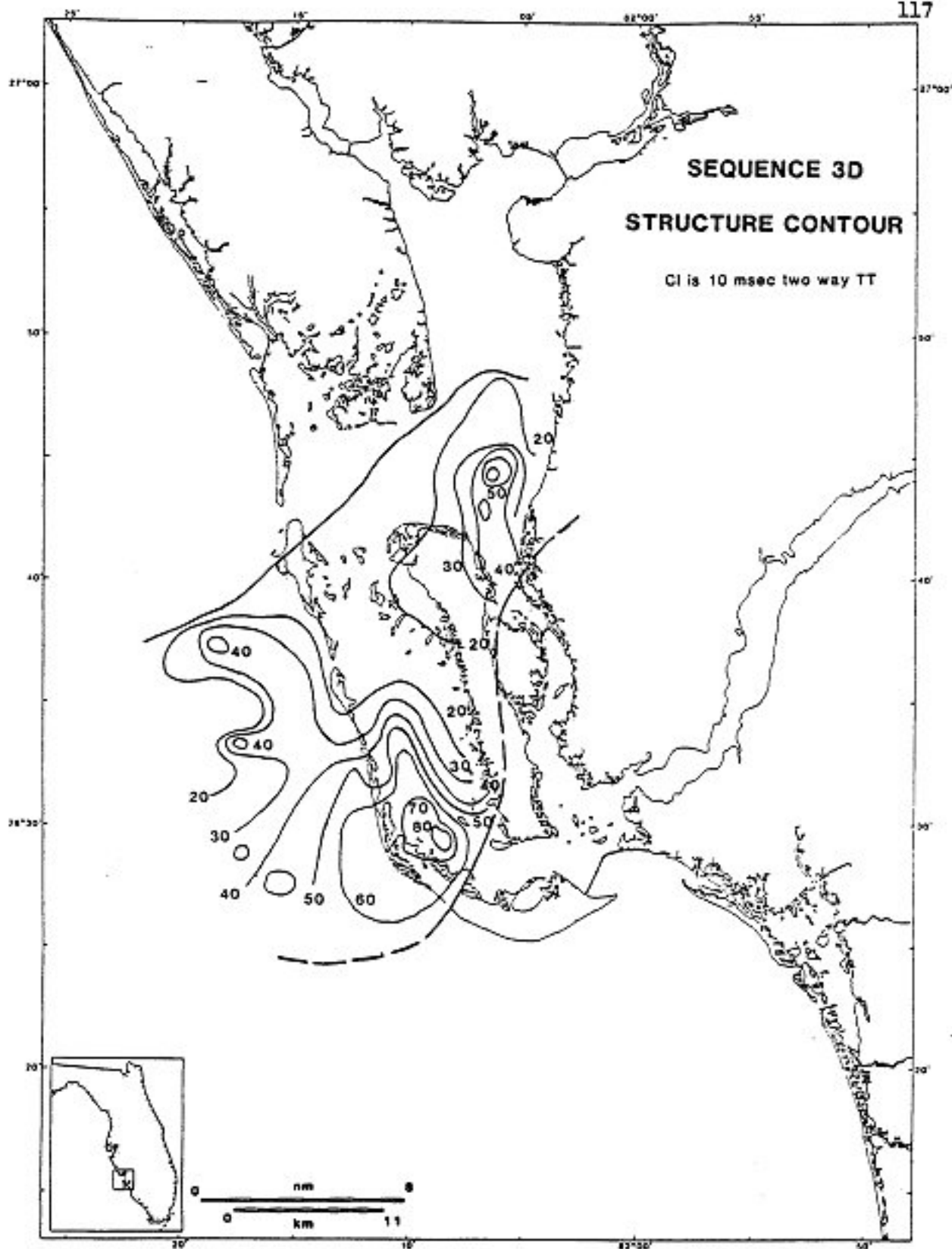




CH-01







Likewise, there is no assurance that Sequence 3A represents the initial depositional unit on the R-3 unconformity. However, some portions of the R-3 unconformity underlying Sequence A occur at a subsurface depth of only 16-20 msec (Figure 24; profile CH-01, 1635-1645). Also the Sequence A reflectors downlap directly on the R-3 unconformity with no underlying sequences as in Figures 15 and 20 (Figure 24). Although these factors indicate that Sequence A is thinning and approaching the land surface in a northerly direction, they do not prove that Sequence A is the basal depositional sequence on the R-3 unconformity.

Structure contour maps and isopach maps of the individual sequences are included as Appendix II. The structure contour maps show the top of each sequence (and the bottom of the overlying sequence) and consequently represent the disconformity surfaces separating each depositional sequence. Figures 25 and 26 illustrate the relationship between depositional sequences (Isopach map Sequence D; Figure 25) and the upper bounding disconformity (structure contour map Sequence D; Figure 26).

Depositional thickening of up to 40 msec (m) of Sequence D occurs within the troughs (Figure 14) while thicknesses of <10 msec occur outside the antecedent lows. Elongation of Sequence D down the trough axes also reflects depositional control by the antecedent topography. The upper bounding disconformity of Sequence D (structure contour, Sequence D; Figure 26), indicates that the Pine Island Sound/Captiva

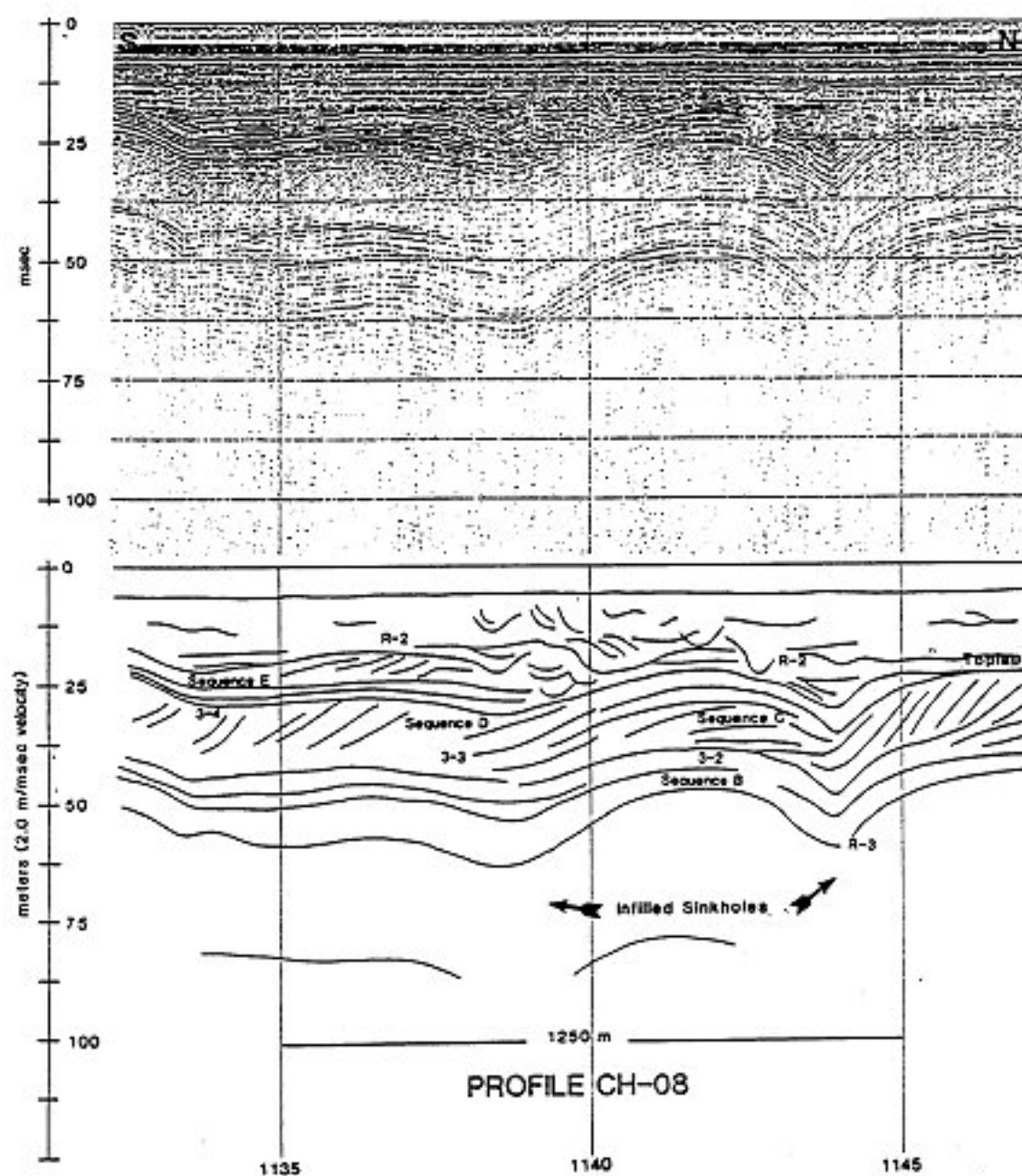
Shoreface and Lower Harbor troughs were only partially infilled with Sequence D deposition.

Figure 26 shows that those troughs were subsequently infilled by Sequence E deposition and remain as isolated topographic depressions in the Sequence E structure contour map. That SE trending trough, which extends into San Carlos Bay, was not completely infilled by Sequence E and is the site of the limited distribution of Sequence F (Figure 22 and Appendix II). In combination with the structure contours of the regional unconformities (Figures 14 and 18), these maps (Appendix II) represent a depositional sequence model which must be correlated and calibrated with direct lithologic data in order to create an integrated geologic model.

Seismic Facies

Four seismic facies have been identified from the seismic profiles. The R-2 unconformity separates thin, predominantly transparent, and channel infill facies from an underlying series of prograding and aggrading sigmoid to oblique clinoforms. Post R-2 sequences are generally thin, acoustically transparent and strongly overprinted with the bubble pulse and bottom multiples and consequently cannot be mapped or correlated between profiles (Figures 16, 19).

Where channel incision has eroded into underlying sequences (i.e., Figures 18 and 19) a distinct channel fill facies is present. The channel facies possess an asymmetric, aggrading pattern of infill indicative of lateral channel



migration (Mitchum et al., 1977). Many of the structures contain superimposed channels indicating multiple infilling/erosion episodes or extensive channel migration (Figure 20; CH-04, 1745-55). Concentrations of infilled channels east and north of Cape Haze indicate that the paleo-Peace/Myakka Rivers bisected the Cape Haze Peninsula in the vicinity of Turtle Bay and Whidden Creek (Figure 18).

Depositional sequences below the R-2 unconformity are composed of a prograding oblique seismic facies which is characterized by an upper toplap relationship (Brown and Fisher, 1980). The updip limit of Sequence D in Figure 27 (CH-08, 1130-40) illustrates the toplap configuration, which is also common to the updip limits of Sequences B and C. These sequences include both oblique tangential and oblique parallel reflector morphologies (Figures 19, 23 and 27). Within the topographically low troughs the stratal reflectors of Sequences D and E become parallel and vertically aggradational (Figures 16 and 17) and assume a sigmoid morphology (Brown and Fisher, 1980). As these sequences infill the Pine Island Sound trough, Sequence E develops an oblique morphology and laterally progrades to the southeast (Figure 22; Seismic section E-E', Appendix I).

Although these Sequences A-F are comprised primarily of only one seismic facie, there is a great range in the lateral and vertical dimensions of the stratal reflectors. The stratal reflectors of Sequence D (Figure 15) have a lateral extent of as little as 100-250 m and vertical extent of 5-10 msec (2-way

TT). The stratal reflectors of Sequence E occur over distances of 1000-1500 m with vertical relief of more than 30 msec (Figure 17).

The thin, discontinuous and/or channelized facies comprising the post-R-2 units are fundamentally different from the seismic facies underlying the R-2 unconformity with respect to continuity, distribution and the internal and external geometry of the stratal reflectors. Additionally, there is a change from oblique to sigmoid morphology as the sequences prograde into and infill the antecedent topographic lows.

Quaternary Sequences

Modern Depositional Environments

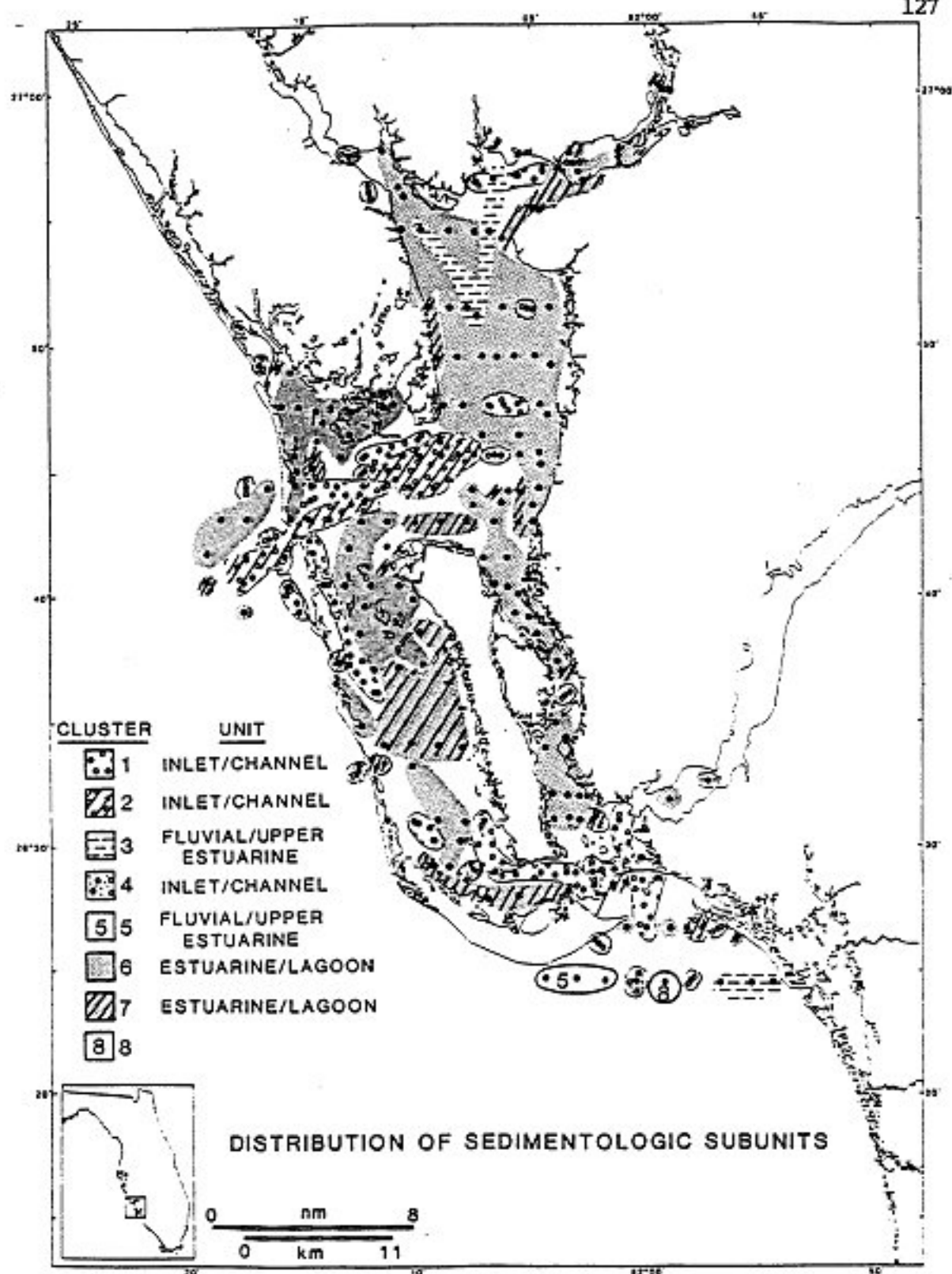
Six modern depositional units and sub-units have been identified from the Q-mode factor and K-means cluster analyses. Summary statistics for these units are listed in Table 4 with their distributions presented in Figure 28. Compositionally, the sediments of this estuarine system have three end members; a mature, well-sorted, fine to very fine quartz sand, a molluscan shell gravel, and a silt dominated clayey mud. The seven environments listed in Table 4 represent the three end members plus three additional sub-units (the seventh environment represents the combined shelly sub-units) which contain differing proportions of the end-members, i.e., a sandy, shelly sub-unit, muddy shelly sub-unit, and a muddy sand sub-unit.

The distribution of the units/sub-units in Figure 28 indicates that the units can be genetically interpreted on the basis of proximity to tidal inlets/channels and the fluvial systems. Five of the six units/sub-units are mapped and interpreted in this particular procedure. It must be noted that the various statistical procedures produced a uniform grouping of stations, however, the sedimentological parameters exhibited a gradational continuum between end members. Consequently, the boundaries of each sedimentological unit have some variation between analyses but still maintained discrete groupings.

Figure 28 shows that the mud dominated and shell dominated end members have discrete distributions within the estuarine system. Mud dominated samples are found at the confluence of the Myakka and Peace Rivers and at the mouth of the Caloosahatchee River in San Carlos Bay. Shell dominated samples are found adjacent to the inlets and tidal channels and extend throughout the lower Harbor and southern Pine Island Sound. The restricted distribution of the shell and mud dominated lithologies are environmentally diagnostic and can be used in the interpretation of vertical sequences. There is, however, no clear cut sedimentologic criteria to distinguish estuarine from lagoonal deposits because both environments produce sand dominated, slightly shelly, muddy deposits.

Table 4. Summary statistics for depositional units as defined by K-means cluster analysis. Unit/sub-unit titles are based on the grouped sample distributions.

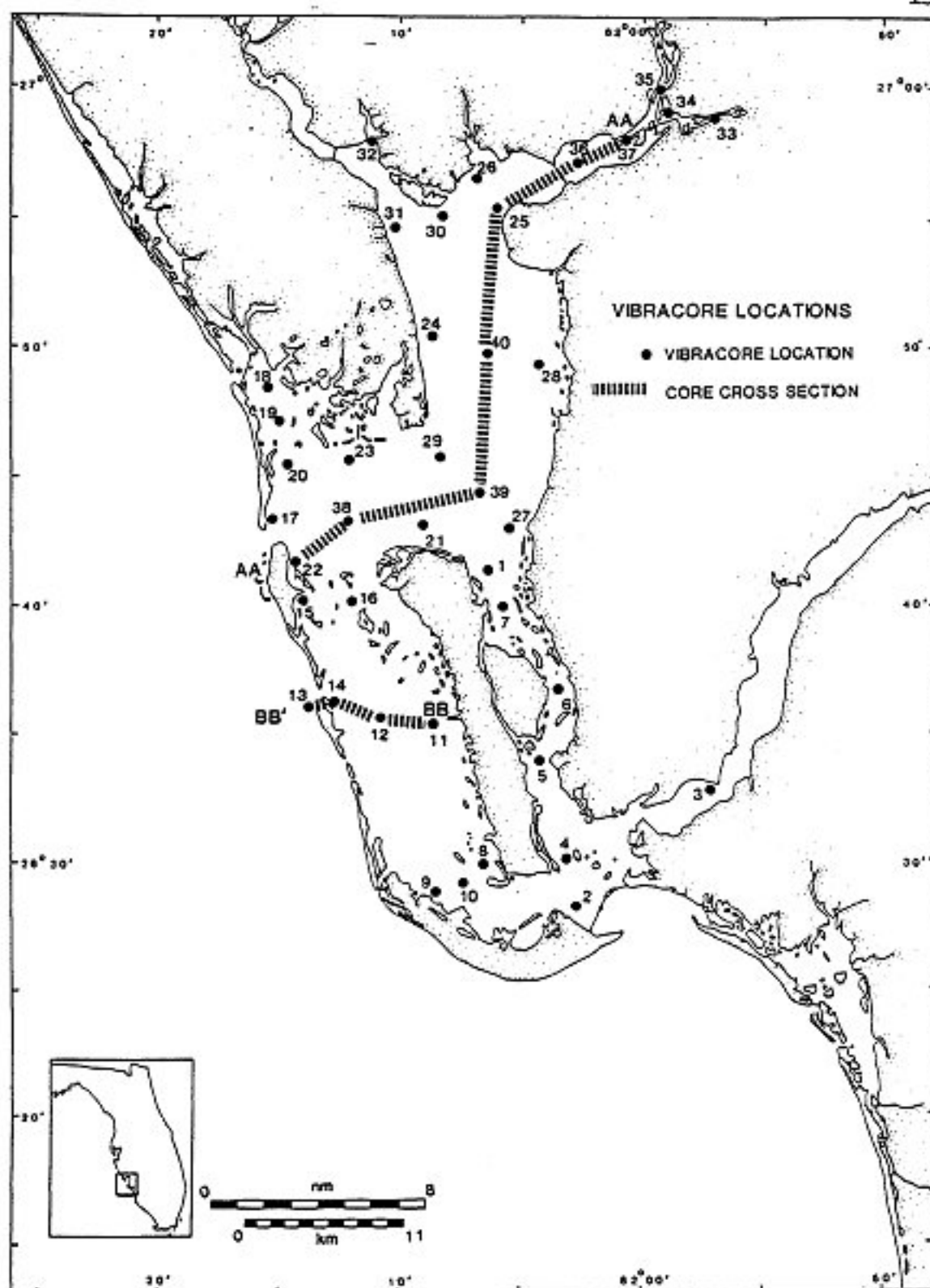
		Fluvial Up. Estu. Unit	Est./ Lagoon Unit	Sandy Lagoon	Inlet/ Chan. Unit	Tidal Delta	Shelly Channel	Muddy Channel
XSRAVEL	mean	5.49	0.67	0.09	7.73	7.31	28.36	4.12
	min	0.00	0.00	0.00	0.00	1.04	10.87	0.00
	max	18.86	7.54	0.95	48.60	16.02	48.60	20.18
	std-dev	7.25	1.42	0.21	5.60	4.18	12.12	5.14
XSAND	mean	47.32	56.17	97.82	85.76	89.43	63.08	88.01
	min	5.10	89.22	95.39	41.01	76.41	41.61	75.88
	max	67.20	100.00	99.63	99.85	97.97	78.41	99.85
	std-dev	19.85	2.91	1.18	6.89	5.73	12.31	6.51
XSILT	mean	28.75	2.41	1.55	4.84	2.63	6.33	5.74
	min	11.72	0.00	0.28	0.00	0.00	0.00	0.00
	max	57.23	7.76	3.59	32.67	11.08	17.98	15.18
	std-dev	12.98	2.04	0.95	4.41	2.23	6.19	4.50
XCLAY	mean	11.77	0.53	0.53	1.62	0.63	2.23	2.64
	min	2.74	0.00	0.00	0.00	0.00	0.00	0.00
	max	29.83	3.76	1.54	10.54	2.54	9.77	9.63
	std-dev	7.72	0.89	0.34	1.80	0.66	3.02	2.18
MEAN GRAIN SIZE (phi)	mean	4.12	2.82	2.95	2.10	1.59	0.37	2.69
	min	2.15	1.41	2.13	0.31	0.38	-0.54	1.15
	max	6.03	3.84	3.84	4.68	2.69	1.83	3.84
	std-dev	1.04	0.38	0.37	0.65	0.32	0.77	0.70
SORT (phi)	mean	2.54	1.05	0.82	1.77	1.72	2.42	1.67
	min	1.55	0.35	0.35	0.62	1.15	1.64	0.62
	max	3.95	1.78	1.11	3.63	2.56	3.55	2.43
	std-dev	0.73	0.29	0.15	0.42	0.37	0.67	0.40
XCaCO ₃	mean	22.82	3.26	2.42	43.54	62.96	87.22	16.57
	min	2.88	0.37	0.37	1.09	40.29	81.23	1.09
	max	48.77	12.54	8.97	93.87	87.45	93.87	33.30
	std-dev	16.42	2.70	2.10	9.34	14.45	4.59	7.49
XORG CARBON	mean	1.91	0.46	0.30	0.79	0.75	1.10	0.75
	min	0.69	0.04	0.13	0.02	0.28	0.02	0.18
	max	3.06	1.53	0.74	2.74	1.85	1.81	1.74
	std-dev	0.67	0.32	0.13	0.45	0.46	0.60	0.42
XORG N	mean	0.13	0.04	0.03	0.07	0.06	0.14	0.06
	min	0.07	0.01	0.01	0.01	0.02	0.04	0.01
	max	0.25	0.13	0.09	0.57	0.10	0.57	0.15
	std-dev	0.05	0.02	0.02	---	0.02	0.16	0.03
SKEW	mean	0.18	0.61	1.12	0.20	0.09	0.45	0.22
	min	-0.29	-1.09	-0.19	-1.59	-0.52	-0.13	-1.59
	max	0.63	1.95	1.95	1.60	1.03	0.95	1.60
	std-dev	0.26	0.66	0.51	---	0.40	0.37	0.62
KURT	mean	0.10	9.72	18.04	2.60	1.64	1.11	3.39
	min	-1.10	-0.28	12.86	-1.07	-0.25	-0.63	-0.82
	max	4.07	45.36	45.36	12.85	5.84	4.66	12.85
	std-dev	1.27	5.97	6.41	2.53	1.66	1.92	3.11
XPHOSP	mean	4.15	1.50	1.17	1.76	1.88	0.57	1.92
	min	2.01	0.00	0.00	0.00	0.14	0.00	0.00
	max	6.37	9.03	2.32	7.12	7.12	1.50	4.98
	std-dev	1.52	1.01	0.71	1.16	1.54	0.49	1.08

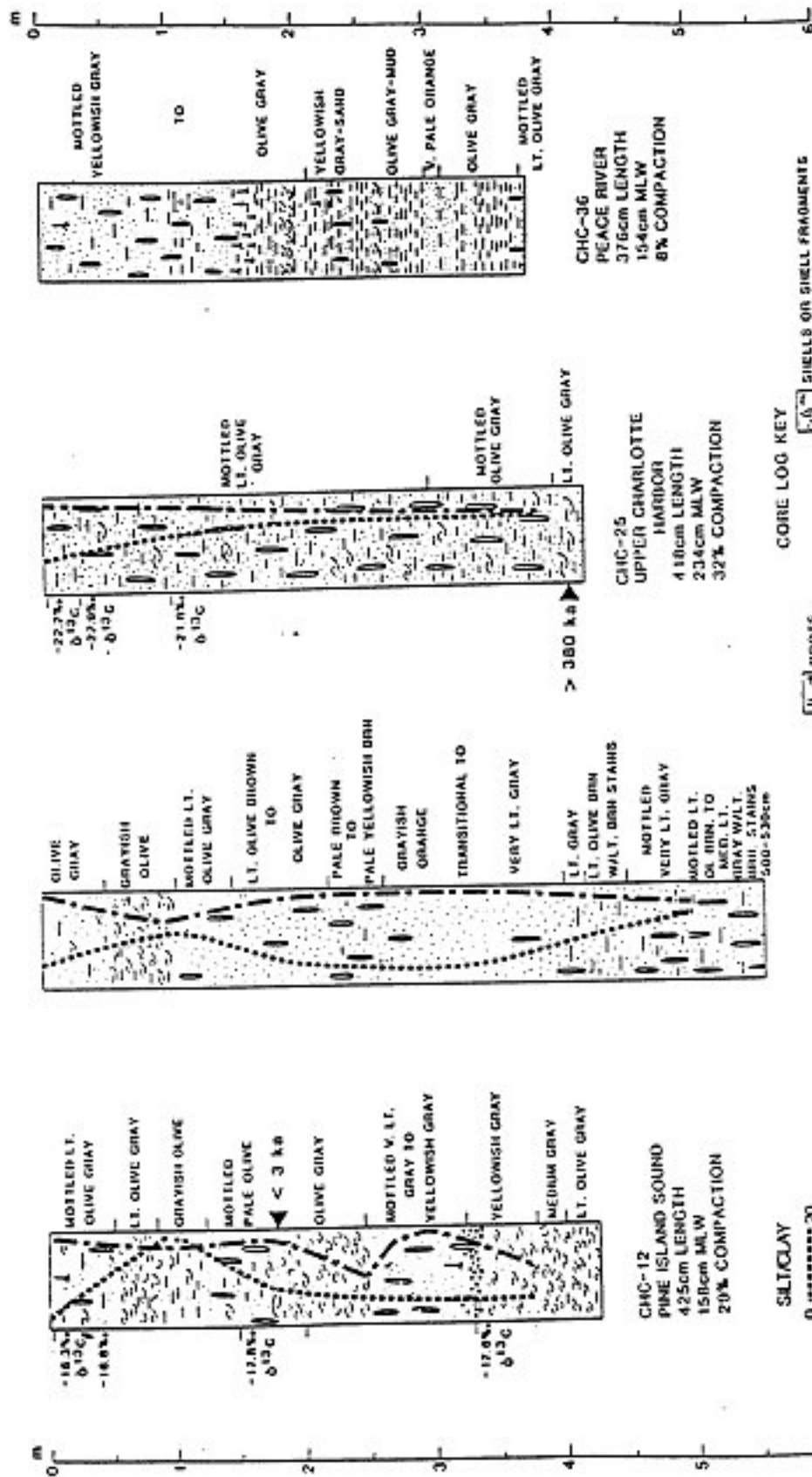


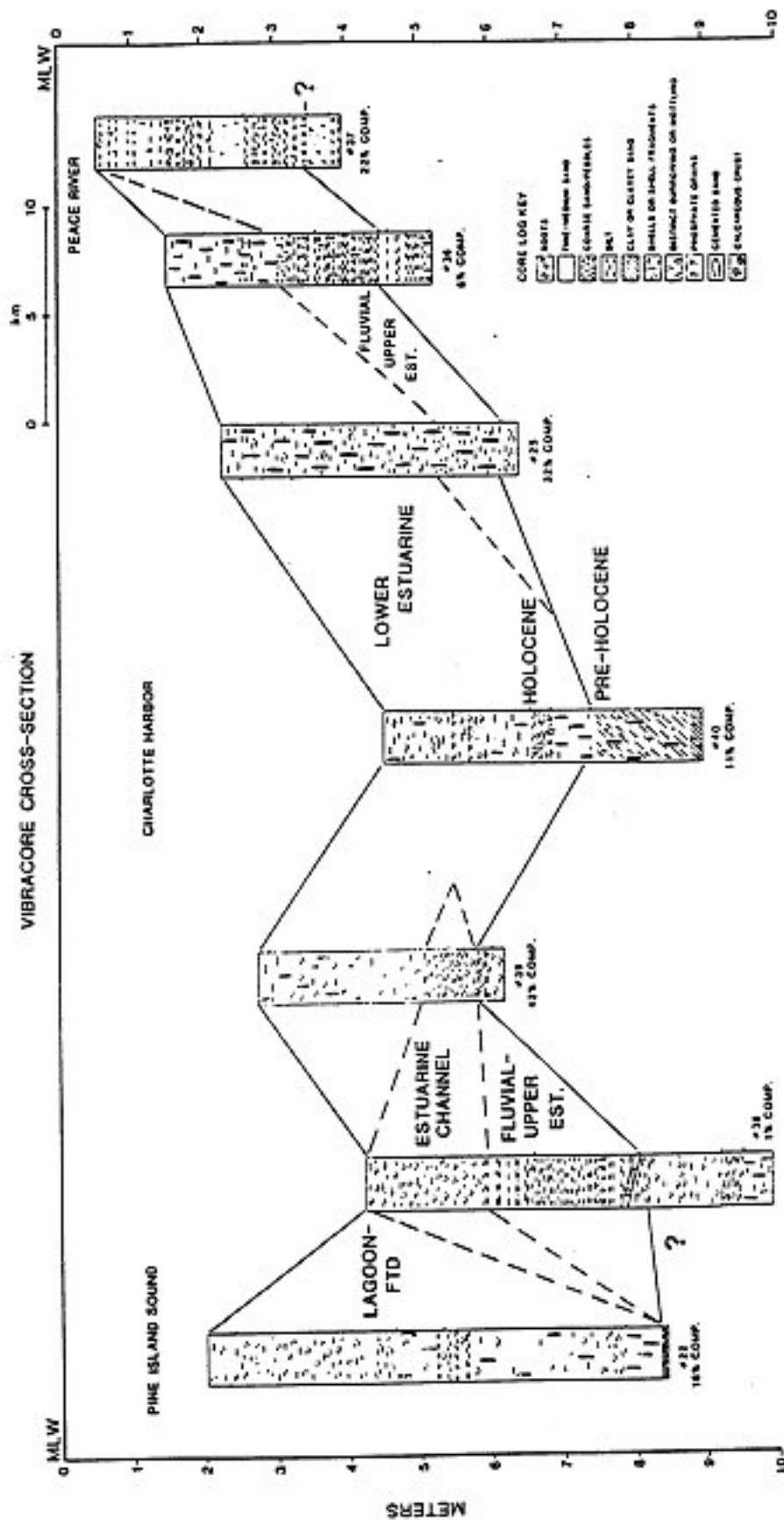
Quaternary Sequence Interpretation

Three characteristic litho-sequences are recognized from 40 vibracores taken throughout the Charlotte Harbor system (Figure 29). Representative core logs of the vertical sequences are shown in Figure 30 (core logs for all cores are included as Appendix V). The three sedimentologic end members described in the previous section are readily apparent. Core 36 (Figures 30 and 31) contains fining upward sequences (FUS) beginning with coarse quartz sands and phosphate pebbles which grade upward into laminated sand and mud and then into laminated muddy clay. The coarse sand/pebble lag is the only Quaternary unit not recognized from the surficial sediment analysis. This lithology is found only in the subsurface within the estuarine/lagoon system and consequently was not sampled or statistically recognized (Figure 28).

The coarse sand/pebble sequences were found only in the upper estuarine portions of the study area and underlying other sequence types in the central Harbor area. Although two coarse quartz/pebble lags occurred near the base of cores 12 and 15 in Pine Island Sound, the general location of these sequences, the presence of thick mud laminae (described only from the fluvial portion of this system), and the non-fossiliferous coarse quartz/phosphate pebble lag all indicate that this type of FUS is a result of fluvial channel and point bar deposition (Barwis, 1978; Taylor and Woodyer, 1978; Smith, 1987). This interpretation indicates that the mud-dominated environment as depicted in Figure 28 represents erosional exposure of relict







fluvial deposits, which is supported by the cross section of Figure 31 and the location of the deep channels in Figure 2.

Cores 25 and 36 (Figures 30 and 31) contain a gradual transition from the laminated sand and mud of the point bar FUS to a mottled muddy sand. Portions of core 25 (300-385 cm) appear laminated and become increasingly mottled upward as a result of increased bioturbation and/or a decreased rate of deposition. Large ghost/mantis shrimp mounds (10-15 cm high and 20-30 cm diameter) are common throughout the upper Harbor area (personal observation and J. Culter, personal communication) and support the idea of widespread bioturbation.

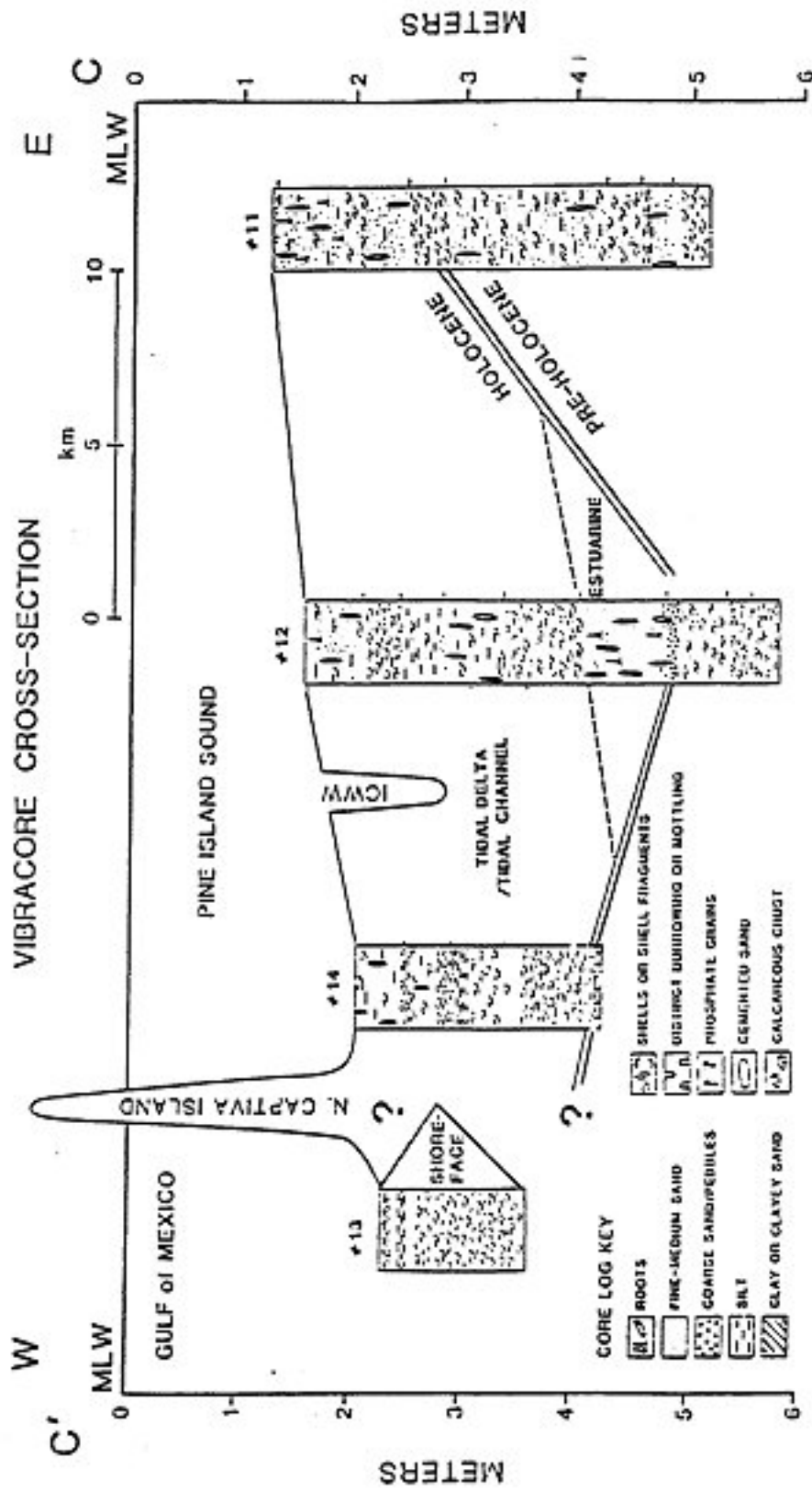
Cores 22, 38, 39 and 40 (Figure 31) all possess complete or portions of FUS beginning with coarse shell lags which fine upwards to shelly sand and then a shelly, slightly muddy sand. These sequences appear to become smaller and more numerous from the estuarine to lagoon portion of Figure 31. This change is consistent with the change from large tidal-channels in the lower Harbor to more ephemeral tidal inlets within the lagoons.

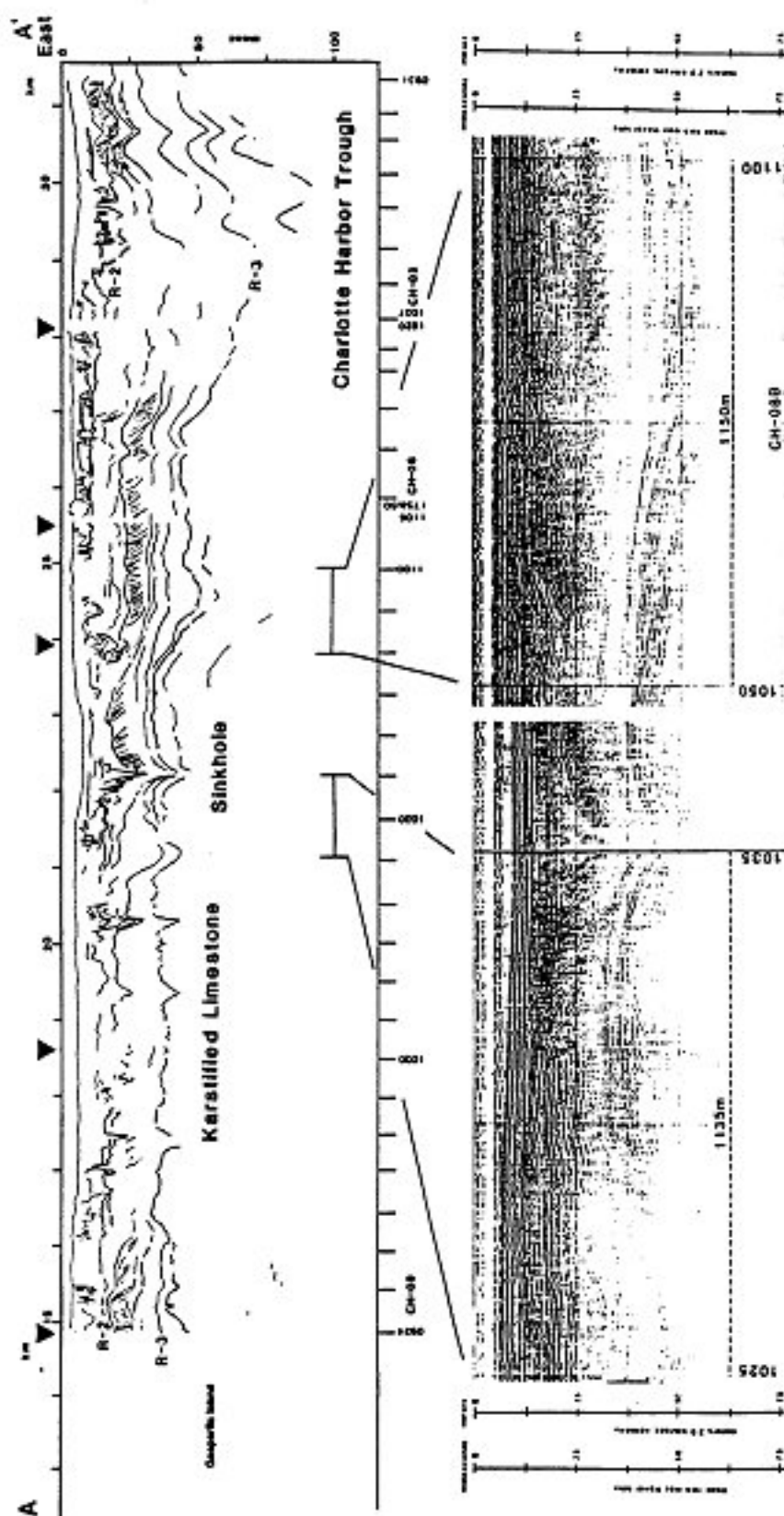
Figure 32 illustrates a core cross section across Pine Island Sound. The lagoonal sequences are composed almost entirely of approximately one meter thick shelly FUS, which supports the above observation of increasing sequence frequency and decreasing size from estuarine to lagoonal cores. Note also the overall increase in carbonate shell material near the barrier/inlet system. The E-W seismic section in Figure 33 shows numerous, discontinuous reflectors with distinct channel morphologies. These structures cannot be mapped or correlated

due to limited lateral continuity and extensive reworking.

The surficial distribution of coarse shell lags and shelly sands along the barrier/inlet trend and within the lower Harbor and southern Pine Island Sound (Figure 28) and the discontinuous, channelized seismic reflectors indicate that the shelly FUS are the product of flood tidal delta and tidal-channel migration and infilling (Moslow and Tye, 1985). Although it is not possible to sedimentologically differentiate overwash from tidal channel deposits, the origin of these 0.5 to 1.5 m thick sequences as baywide, catastrophic overwash events (Bland, 1985) should produce continuous mappable reflectors as opposed to the discontinuous channel structures of Figures 20 and 33.

Stable carbon isotope ratios are presented in the core logs of Figure 30 (and Appendix V). The surficial sediment analyses did not permit differentiation of lagoon and estuarine deposits. Comparison of $\delta^{13}\text{C}$ data from cores 10, 12, 22, and 25 indicates that the "estuarine" deposits of core 25 have $\delta^{13}\text{C}$ values between -21.8 to -22.9 while the "lagoonal" deposits of cores 10, 12, and 22 have values between -16.0 and -18.2. The more negative upper estuarine values indicate the contribution of organic material from terrestrial and/or wetland vegetation (mangroves; -25 to -28, *Juncus*; -24 to -26 and terrestrial dicots; -22 to -26; Evans, *et al*, 1985). The less negative lagoonal values reflect the contribution of seagrasses (-13 to -15 $\delta^{13}\text{C}$) in combination with marine phytoplankton (-18 to -21).





Chronostratigraphy

Table 5 presents the results of 20 amino acid racemization (AAR) analyses on articulated Chione cancellata or Mercenaria campechensis shells. Aminostratigraphy is based on a monotonic increase in D/L ratio with age, at a rate determined primarily by temperature and genus of the fossil. Chione and Mercenaria are closely related species and demonstrate little difference in AAR kinetics (Belknap, 1979; Wehmler and Belknap, 1982). Kinetic model ages are determined from the AAR ratios using assumptions about temperature history. The quoted model ages represent the Wehmler and Belknap (1982) isochron model ages, with an uncertainty of $\pm 20\%$. AAR aminostratigraphy and kinetic age modeling are useful in that the method requires only 1 gram samples and extends well beyond the range of ^{14}C dating, but lower precision and leaching did effect some results (similar problems with alteration also effect radiocarbon dating).

In Table 5, data in parentheses should be interpreted cautiously due to analytical problems (potential error is still within the $\pm 20\%$ uncertainty). Also ratios greater than about 0.85 D/L leucine are in equilibrium and are quoted as "greater than" minimum ages. Data less than 0.2 represent poorly resolved Holocene samples. Kinetic model ages are guides only: more repeats will be required to clarify groupings, and the presentation of samples by isotopic stages is the best level of accuracy possible (Figure 34). The raw age assignments range from <3 KA to > 380 KA, with half of the

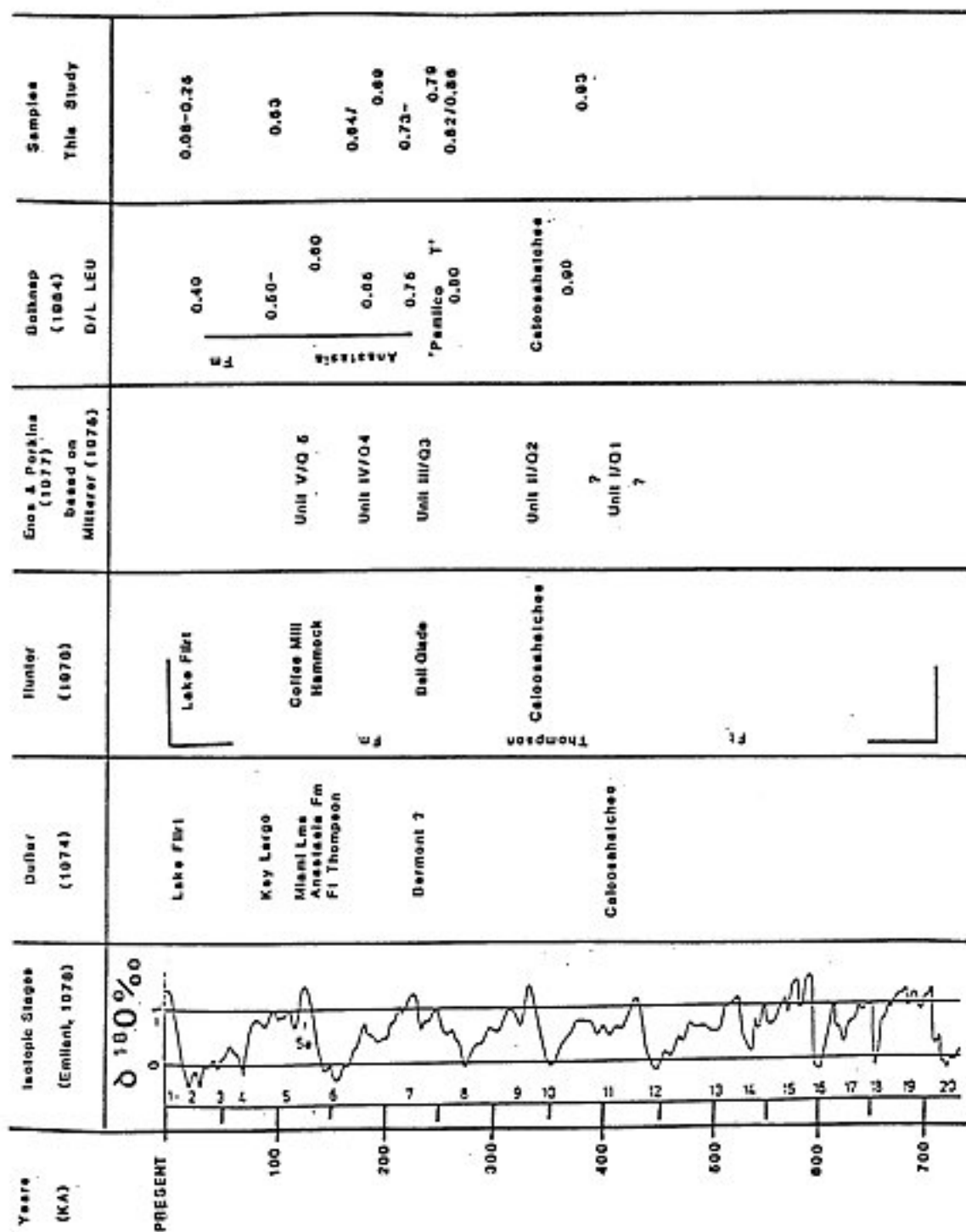
samples representing isotopic stage 7 (180-250 KA; Figure 34). Six of the 20 samples have Holocene age assignments (<8 KA) and three may represent stage 5 (with possible distinction of stage 5a and 5e). Two samples gave only minimum dates, while a third was too leached to be reliable.

Three sets of samples (cores 9, 11 and 28) produced anomalous results with older samples overlying younger samples. For core 9, the two data points are not analytically well resolved nor are they resolvable in the kinetic model. They represent indistinguishable Holocene age shells. In core 28, the same analytical uncertainty holds, but if the ratios are correct, the assigned ages would overlap. The dated samples from core 11 show the basal lag of one FUS dated at >275 KA overlying another FUS dated at 218 ± 42 KA. Although it is possible that the older, upper sample was reworked from some other depositional unit, the use of articulated samples makes it unlikely. The upper sample is not reliable, probably due to leaching.

In addition to the age-dated amino acid samples, most of the cores contained pronounced unconformities such as leached, indurated shell beds, iron or humic stained paleosols, or blue green sandy, phosphatic clays. These lithologic features, which can only be used to separate Pleistocene/Holocene deposits, are listed in Table 6 along with the compaction corrected depth in each core. The interpreted Holocene/Pleistocene contact is considered to be the base of the depositional sequence immediately overlying the dated

Table 5. Summary of amino acid dated samples. Species represented are Chione cancelata (C.c.) and Mercenaria campechensis (M.c.). Analyses in parentheses indicate leached or degraded samples and require cautious interpretation. All samples used articulated and/or matching adjacent valves.

Core #	Depth (in core-cm)	Species	D/L leucine	(X _e -X _i)/X	Kinetic age KA BP 22°C
5	155	C.c.	0.81	0.10	240+/-48
7	120	M.c.	(0.78)	0.13	(214+/-42)
7	120	C.c.	0.64	0.22	132+/-26
7	251	C.c.	(0.78)	(0.12)	(220+/-44)
8	395	C.c.	0.53	0.31	82+/-16
8	475	C.c.	No Return - (Dld, leached ?)		
9	343	C.c.	(0.25)	(0.60)	<8 KA
9	495	C.c.	(0.13)	(0.77)	<5 KA
11	306	C.c.	(0.86)	(0.08)	(275 KA)
11	365	C.c.	0.78	0.12	218+/-42
12	176	M.c.	0.08	0.86	<3 KA
15	317	C.c.	0.15	0.74	5+/-2
15	708	C.c.	0.92	0.10	270+/-54
18	165	C.c.	0.18	0.79	4+/-2
19	393	M.c.	0.08	0.86	<3 KA
19	417	C.c.	(0.69)	(0.18)	(160+/-32)
25	402	M.c.	(0.93)	(0.04)	(>380 KA)
27	375	C.c.	(0.79)	(0.12)	(220+/-44)
28	205	C.c.	(0.76)	(0.14)	(200+/-40)
28	230	C.c.	0.73	0.16	180+/-36



samples and/or the lithologic indicators (Figures 30, 31 and 32).

The summary of south Florida Quaternary stratigraphy that is presented in Figure 34 indicates general agreement on the ages of Pleistocene deposition which are closely correlated with glacial cycles interpreted from the oxygen isotope record (Emiliani, 1978). The dated depositional sequences from this study consist primarily of shelly FUS or shell lags because such deposits contain the species most amenable to dating. However, the occurrence of similar lithologic sequences of different ages within the lagoon system indicates that the interpreted tidal channel/tidal delta environment has consistently occupied this area over several transgressive/regressive cycles.

Quaternary Sequence Distribution

Figure 35 presents an isopach map of all post R-2 deposition (based upon an acoustic velocity of 1700 m/sec). Sediment thicknesses which include Holocene, Pleistocene and possibly some Pliocene deposits are up to 18 m thick in southern Pine Island Sound and in the infilled channels of the lower Harbor. These unconsolidated deposits are between 3 and 6 m thick along the shoreface and have been removed by shoreface incision 5 to 7 km offshore. There has also been significant erosion in Boca Grande, San Carlos and Matlacha Passes. There are insufficient data points to construct a

Holocene isopach, however the tabulated sediment thickness and depth to Pleistocene values of Table 6 correspond to Figures 18 and 34 (respectively).

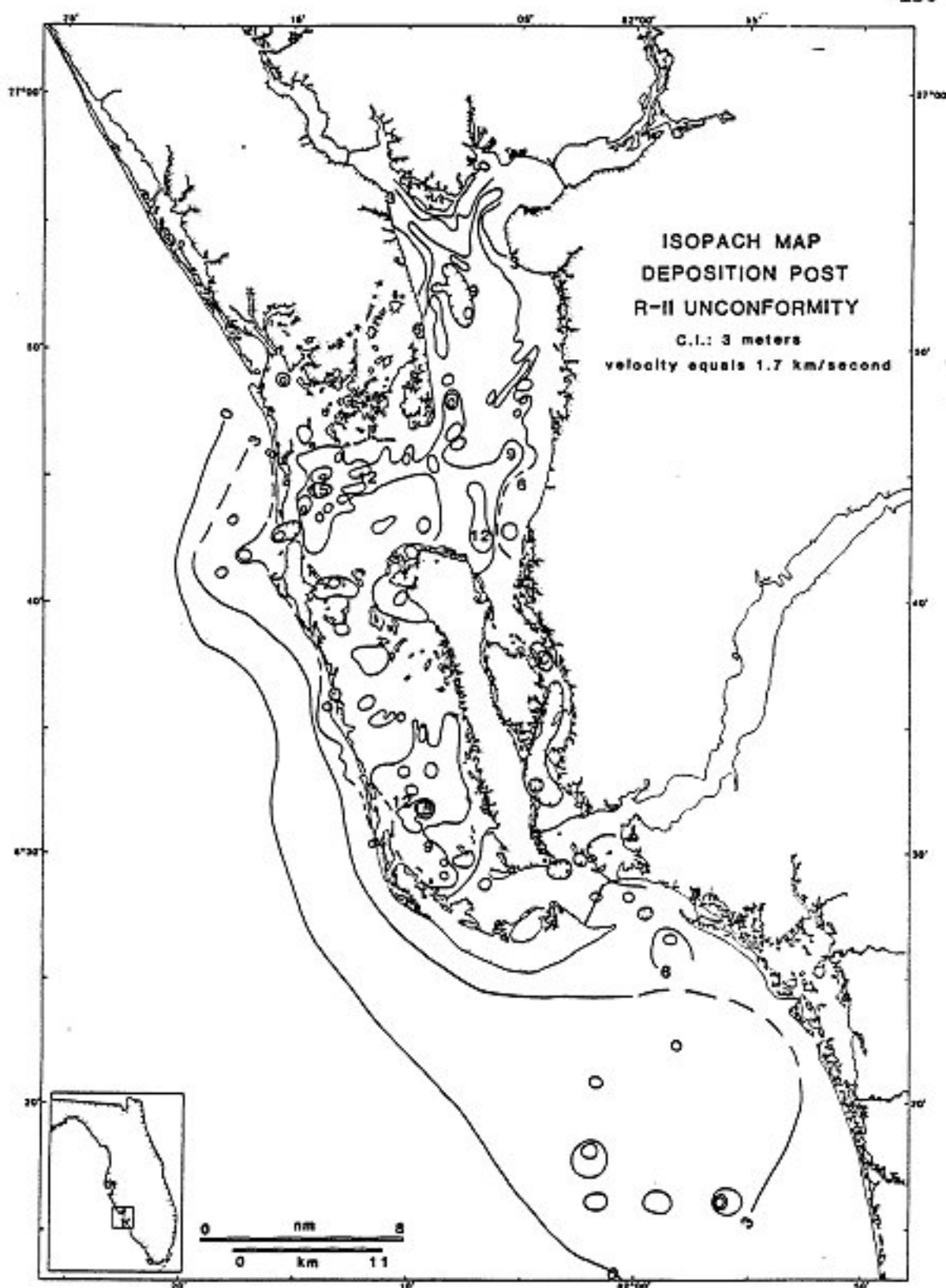
The cores in Table 6 are listed from landward to seaward, beginning with the fluvial/upper estuarine cores down through the upper Harbor, lower Harbor, Matlacha Pass and ending with the lagoonal cores. The Holocene sequences are listed with number 1 at the Pleistocene/Holocene contact and increasing upward. The Pleistocene deposits underlying the unconformity usually consist of leached, indurated shell lags (primarily Chione cancellata) and resemble the basal lag of the inlet/tidal channel FUS. Estuarine and point bar sequences of probable Pleistocene age were also observed (cores 36, 32, 31, 27 and 20), however, these sequences are incomplete and only tentatively interpreted.

Cores from the fluvial/upper estuarine portion of the study area (3, 34, 33, 37, 36, 32, and 26; Figure 29) are mostly composed of the point bar FUS. Cores 3, 32 and 26 taken from near the mouths of rivers or creeks are composed of the transitional estuarine sequence. Estuarine conditions prevail in those areas that, morphologically, would be considered part of the fluvial regime. Cores from the upper Harbor also consist of the estuarine sequence with the exception of 28 and 40 which are composed of tidal channel FUS (28) and tidal channel overlying an estuarine unit (40).

The lower Harbor cores consist of tidal channel FUS with the exception of cores 27 and 23 which are located on the

Table 6. Summary of Pleistocene/Holocene unconformities and Holocene litho-sequences. Sequence indicators are described in the text as: Point bar (PB), Estuarine (Est), and Tidal channel (and/or tidal delta; TC). Thicknesses and water depths are in cm with water depths referenced to MLW.

core	sequence indicator	depth cm-MLW	Holo.thick cm	total depth-cm	Holocene sequences		
					1	2	3
<u>Fluvial/upper estuarine cores</u>							
3	phos.clay	135	162	198	EST		
34	?	86	410(?)	496	PB	PB	
33	humic stn	26	275	301	PB		
37	?	59	422	481	PB	PB	PB
36	Fe stn	154	324	478	PB	EST	
32	Fe/humic stn	180	354	534	EST		
26	phos.clay/ humic stn	57	594	651	EST		
<u>Upper Harbor</u>							
31	humic stn	169	280	449	EST		
30	indur.surf.	107	123	230	EST		
25	AA date	234	528	762	EST		
28	AA date	233	249	582	TC	TC(?)	
40	phos.clay/ Fe stn	457	333	790	EST(?)	TC	
24	indur.surf.	197	175	372	EST		
<u>Lower Harbor</u>							
39	indur.surf.	274	391	665	TC		
29	?	176	549(?)	725(?)	TC	TC	
27	AA date	214	360	574	EST		
21	indur.surf./ humic stn	62	531	593	TC	TC	TC
23	Fe/humic stn	175	158	333	EST		
38	?	427	576(?)	1003	EST	TC	TC
<u>Matlacha Pass</u>							
1	indur.surf.	230	134	364	EST		
7	AA date	178	116	294	EST		
6	indur.surf.	128	306	434	EST		
5	indur.surf./ AA date	169	119	288	EST		
4	indur.surf./ phos.clay	110	330	440	TC	EST	
<u>Gasparrilla/Pine Island Sounds</u>							
18	?	238	256	594	TC		
19	AA date	241	512	753	EST	TC	
20	Fe/humic stn	214	268	482	EST	TC	
17	?	212	278	590	TC		
22	indur.surf.	197	742	939	EST	TC	TC
15	AA date/ phos.clay	119	451	570	PB	EST	TC
16	Fe stn	148	262	410	EST	TC	
14	indur.surf.	256	260	516	TC		
12	indur.surf.	158	426	584	PB	TC	
11	AA date	124	175	299	TC		
9	AA date/ humic stn	31	593	624	EST	TC	
10	?	136	318	454	EST	TC	
8	AA date/ indur.surf.	93	252	345	TC		
2	indur.surf.	179	430	609	TC		



margins of the Harbor away from channel activity (Figure 2). Cores from Matlacha Pass are composed of estuarine sequences which overlie an indurated surface at relatively shallow depth (288-440 cm; Table 6). Cores from the lagoons (Gasparilla/Pine Island Sounds) are composed of multiple tidal channel FUS directly overlying an indurated surface or an estuarine sequence.

Several amino acid dates (cores 9, 15, and 19) and/or indurated surfaces within or underlying the estuarine sequence indicate that it is a Holocene unit (although the age of the Point Bar unit in core 15 is questionable). The depth to Pleistocene and Holocene thickness data in Table 6 indicate that sediment thicknesses increase in a seaward direction across the lagoons. The exception to that pattern is core 12 (Figure 33) which contains at least one meter of fluvial to estuarine sediments underlying lagoonal deposits.

Neogene Sequences

Borehole Lithologies

The lithologic and borehole geophysical data analyzed for this study comes from a variety of sources (Table 7). Approximately one-half of the borehole sites (Figure 36) are based on wells drilled and logged by the South Florida Water Management District (SFWMD; Wedderburn et al, 1982). These wells were primarily logged by one person (M.S. Knapp) and provide a level of consistent data acquisition that is not

provided by other data sources. The remainder of the data were collected by a number of geologists from the South West Florida Water Management District (SWFWMD), the U.S. Geological Survey (USGS), and the Florida Geological Survey (FGS).

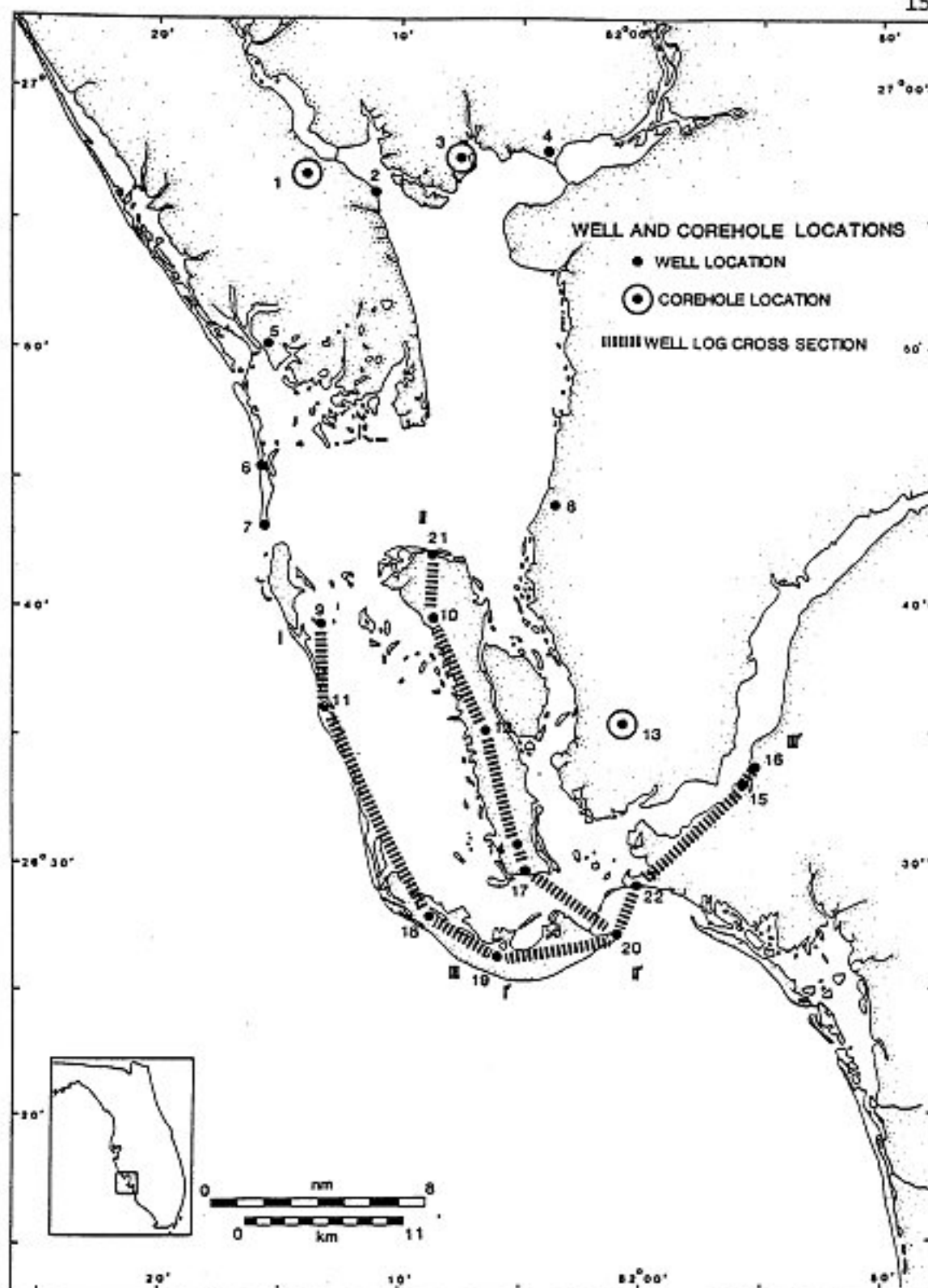
Lithologic descriptions varied with the different geologists and must be liberally interpreted. Three cores were personally examined and logged by the author (From the FGS archive collection) as well as cuttings from wells 21 and 22 (LE-036 and 37, SFWMD). Most of the gamma logs were measured with a 0-500 counts per second (CPS) scale, however, some of the older logs possessed no quantitative scale and represent chart scale only (as indicated on logs).

The lithologies and stratigraphic compositions as illustrated in the core logs of Appendix III seem numerous and complex. There are however, only three lithologic end members within the upper Hawthorn and Tamiami Formations in the study area: Quartz sand (including phosphatic sand and gravel), calcilutite, and a variably indurated, molluscan/foram rich calcarenite or dolomite. The lithologic complexity of the described formations is derived from the varying proportions of the end members, variable diagenetic alteration and induration of the carbonate components, and at least partially, from different descriptive terminology of various geologists.

The calcilutite units are primarily composed of clay to silt-sized calcite with variable amounts of dolomite, aragonite and quartz (Peck et al, 1979b). These deposits have alternately been described as dolosilts-e.g., wells 2 and 8;

Table 7. Source and location of lithologic and geophysical borehole data used in this study. Well locations are illustrated in Figure 36.

Well no.	latitude	longitude	SEWARD	SHEPHERD	FLGS	USGS	elevation	depth	casing	lithology	gamma	neutron	sonic
1	26°56.25	82°13.83		ROMP 3-1	W-15332		10	603.5	247	XX	XX		
2	26°56.20	82°11.07		Coastal Petrol.	Cattle Dock Pt.		5	1406	68	XX	XX		
3	26°57.50	82°07.50		ROMP 2-1	W-15333		2	689	142	XX	XX		
4	26°57.98	82°04.02			xx		8	405	162	XX	XX		
5	26°50.28	82°15.62			Placida #8		7	413	84	XX	XX		
6	26°45.40	82°15.60		Gasparilla Is. Water Assoc.	Injection well			1057	121	XX	XX		xx
7	26°43.38	82°15.57		LE-024	W-7342		4	570	141.5	XX	XX		
8	26°44.30	82°15.42		LE-005			6	195	n/a	XX	XX		
9	26°39.73	82°12.72		LE-015			n/a	345	n/a	XX	XX		
10	26°40.03	82°09.00		LE-014			3	963	361	XX	XX		xx
11	26°35.92	82°13.03		LE-021			5	300	n/a	XX	XX		
12	26°35.50	82°06.85					n/a	270	n/a	XX	XX		
13	26°35.52	82°01.35		LE-029	L-665		6	662	217	XX	XX		
14	26°31.28	82°05.17		LE-006	W-15487 Cape Cora		8	650	n/a	XX	XX		
15	26°33.27	82°05.25		LE-018			5	620	n/a	XX	XX		
16	26°32.62	82°04.97			L-1849		n/a	155	n/a	XX	XX		
17	26°29.58	82°05.35		W-035			5	1420	624	XX	XX		
18				LE-026	W-14011		2	635	n/a	XX	XX		
19				LE-004			4	774	660	XX	XX		
20	26°27.04	82°01.05			L-2401		5	891	470	XX	XX		xx
21	26°28.48	82°00.24		LE-036			5	300	-	XX	XX		xx
22	26°29.00	82°00.65		LE-037			3	300	-	XX	XX		xx



clays-wells 5 and 13; dolomite-well 15; and commonly called dolomitic or limey marl. Scott (1988) describes the clays of the synonymous Peace River Formation as calcitic and/or dolomitic with montmorillonite and palygorskite comprising the primary clay minerals.

The clay or calcilutite in core 13 (Figure 37) occurs between two strong unconformities, both noted by blackened, indurated limestones. The clay unit is approximately 20 m thick and is laminated with fine sand/silt layers that vary from a few millimeters to 10 cm in thickness. The clay unit grades downward into a sand unit containing phosphate pebbles and growth oriented oyster shells, which directly overlie the basal unconformity.

The quartz sand dominated units (Figure 37) occurs in most holes at least once and commonly as two or three discrete units, e.g., Wells 3, 11, 17 and 20. Sandy clays and sandy limestones are probably more common than the sand unit of Figure 37 and at least partially represent sample homogenization from the rotary drilling procedure (Maher, 1964).

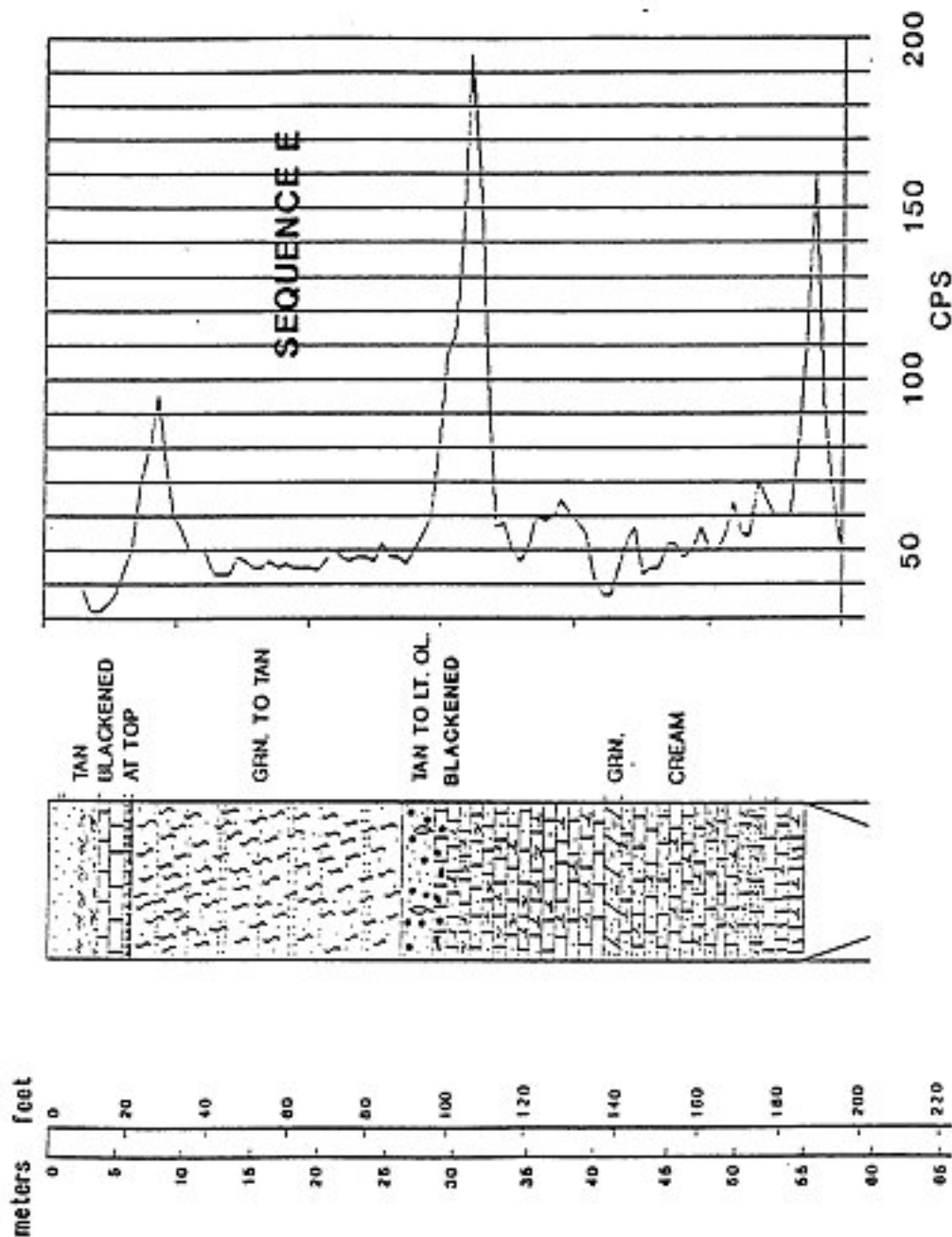
The calcarenite units consist primarily of unconsolidated to poorly indurated, sandy, foram-rich molluscan shell hash (sandy, skeletal grainstones) at depths of 0 to approximately 30 m. Deeper limestone units grade into poorly to moderately indurated packstones with an increased proportion of forams and lime mud and a corresponding decrease in molluscan skeletal particles.

Dolomites and dolomitic units are common in the recorded lithologic logs (i.e., Wedderburn et al., 1982; Sutcliffe, 1975) and the graphic logs of Appendix III. The actual proportion of dolomite in these deposits may or may not be reflected in those logs. The identification of dolomite on geologists logs is controlled by a combination of grain size, color, induration, and acid effervescence. Individually, these factors are not quantitative indicators of dolomite abundance (Maher, 1964), but collectively, probably do represent the relative abundance calcite and dolomite. The complex stratigraphy of this area and possible multiple origins of dolomite (Back and Hanshaw, 1970; Baker and Burns, 1985), indicate that detailed mineralogical study should be pursued.

The problems of interpretation notwithstanding, the dolomites in this area are present in two forms; as disseminated grains (or cements) in the calcilutite units (e.g., wells 1, 2, 8, and 10), and as moderate to well indurated, white to light orange beds with thicknesses of 3 to 20 m (e.g., wells 1, 9, 10, 11, and 17). The bedded dolomites are often associated with unconformities; e.g., wells 1, 3, 8, and 9. The orange color (Fe stain ?) and dolomitic composition may represent phreatic conditions and diagenesis.

Unconformities are common in the lithologic record and include the previously described blackened, indurated limestones and sandy phosphate pebble lags. Chert horizons are also present as unconformities (wells 5 and 9) and are probably alternate descriptions for the blackened limestones.

WELL #13 CAPE CORAL (Core #15487 FGS)



LITHO-LOG

GAMMA LOG

Most of the observed unconformities develop on top of the limestone units (e.g., well 13; Figure 37). This apparent sequential juxtaposition of unconformities on limestone units may result from the limestones occupying a terminal position in a transgressive-regressive couplet, or because unconsolidated sands and calcilutites erode upon subaerial or submarine exposure leaving a relatively resistant limestone.

Figure 37 also shows the correlation of gamma peaks to unconformities. The gamma logs, which measure natural gamma decay, is strongly influenced by the phosphate-rich lag deposits at the unconformities. If the lithologic unconformities produce a measurable acoustic impedance anomaly, the gamma peaks should be correlative with seismic reflectors.

Borehole Geophysics

The borehole geophysical logs for wells 21 and 22 are included in Appendix III. Figures 38 and 39 show those logs along with the graphic lithologic logs and the derived seismic model for each site. The seismic models were created from the seismic sequence structure contour maps which show travel time to each sequence in milliseconds. Vertical depth scales are generated by assuming acoustic velocities of 2.0 m/msec as in the previous section.

The uncalibrated sonic velocities of wells 21 and 22 range between 2.5 and 5.5 m/msec and average between 4.2 and 4.5 m/msec (Appendix III). Consequently, sonic velocities from

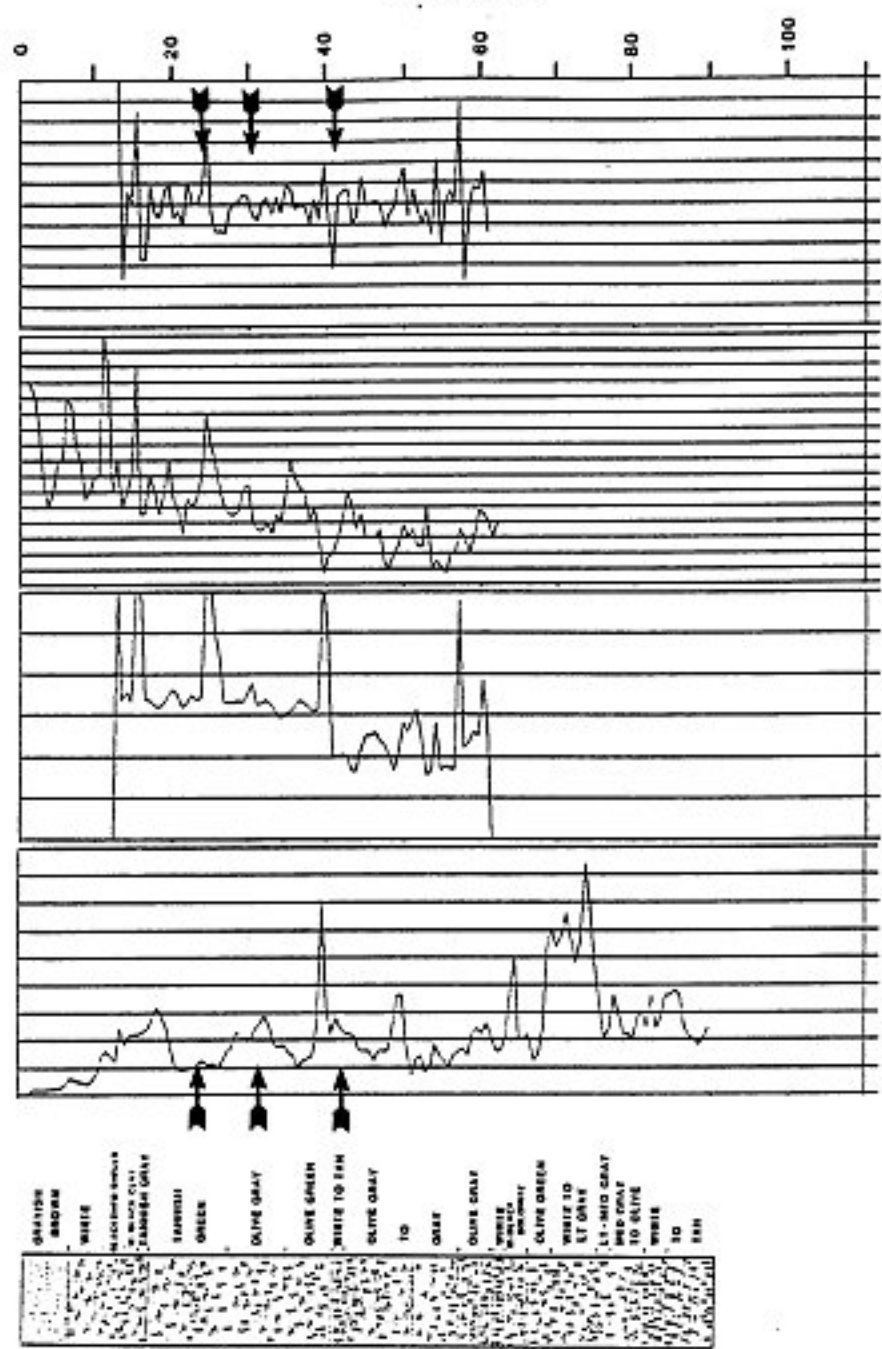
the borehole logs are used as dimensionless estimates of velocity in the calculation of the impedance and reflection coefficient logs.

Figure 38 contains the geophysical log data from Well 21 on Bokelia Island. The seismic model, generated from the sequence structure contours and isopach maps (Appendix II), indicates that this section should contain 3 sequences between the R-3 unconformity at 42 msec and the R-2 unconformity at 12 msec. Each of the geophysical logs of Figure 38 (gamma, velocity, density and RC) shows a tripartite set of peaks in the 44 to 10 m range.

The gamma and density logs correspond with the predicted seismic model with peaks at 42-30 m, 30-22/20 m and 22/20-12 m. The velocity and RC logs show a similar tripartite set of peaks but at 40-25 m, 25-15 m and 15-12 m. Because each of the logs responds most dramatically to different geophysical parameters, offsets in the major peaks, possibly due to "cycle skips" of the sonic log, are to be expected (Doveton, 1986).

Figure 39 shows the same seismic-geophysical data for Well 22 which is located very close to the water on the Sanibel Causeway (Figure 36). The seismic model predicts two sequences (Sequences E and F) in this area between 44 and 10 msec. The gamma and density logs again show remarkable correlation with major peaks at 44-20/18 m and 20/18-10 m. The sonic log is inconclusive at this site due to obstructions and loss of data at about 27 m. That portion of the record available does indicate a major peak at 44 m and uniform readings up to 27 m.

Timing diagram for MSEC 2 Way TT (2000m/sec Velocity). The diagram shows three sequences: Sequence E (0 to 20), Sequence D (20 to 30), and Sequence C (30 to 40). The diagram is divided into sections R-2 and R-3. The x-axis is labeled 'MSEC 2 Way TT (2000m/sec Velocity)' and ranges from 0 to 100.

REFLECTION
COEFFICIENT LOG

DENSITY LOG

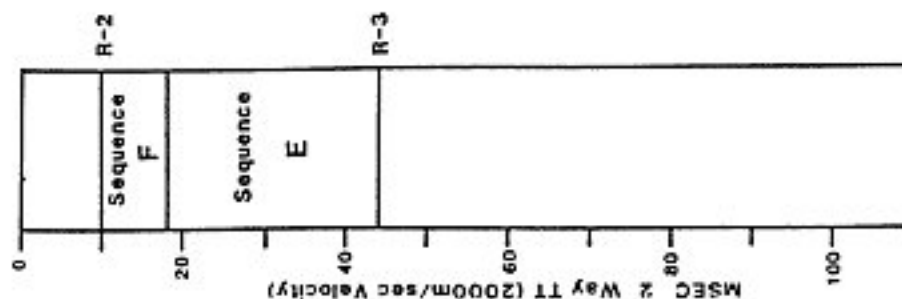
VELOCITY LOG

GAMMA LOG

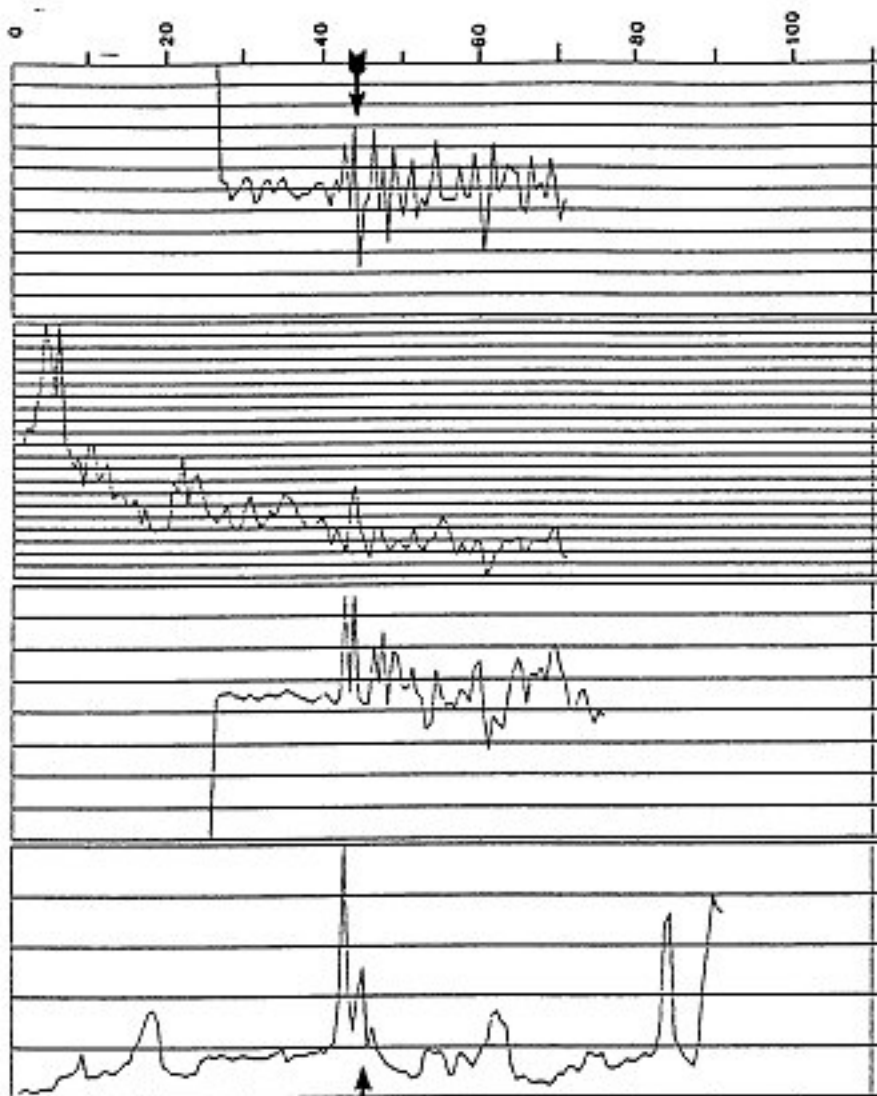
LITHO-LOG

SEISMIC MODEL

SEISMIC-LITHOLOGIC SEQUENCE CORRELATION Well #22 (Sanibel Causeway)



MEDIUM GRAY
 LIGHT GRAY
 WHITE
 TAN
 MEDIUM GRAY
 WHITE
 WHITE TO TAN
 LIGHT GRAY
 LIGHT GRAY
 WHITE
 TAN
 MEDIUM GRAY
 WHITE
 WHITE TO TAN
 LIGHT GRAY



REFLECTION
COEFFICIENT LOG

DENSITY LOG

VELOCITY LOG

GAMMA LOG

LITHO-LOG

SEISMIC MODEL

These figures, based upon an estimated seismic velocity of 2.00 m/msec, show that the seismic models and seismic sequences can be correlated with borehole geophysical data. It is also significant that the gamma log peaks consistently occur at lithologic unconformities. Although this correlation has been well established (Altschuler et al, 1964; Peck et al, 1979b; Wedderburn et al, 1982), it is especially important for the determination of unconformity-bounded depositional sequences.

Neogene Sequence Distribution

The lithologic cross sections in Appendix IV integrate the borehole lithologic data with the seismic sequence model. These sections include; a north-south line with holes from Useppa, North Captiva and Sanibel Islands (I-I'), a second north-south line from Bokelia, Pine Island and Sanibel Island (II-II') and an east-west line from Sanibel Island to the mainland (III-III'; Figure 36). Sequence boundaries were located on the lithologic sections using described lithologic unconformities, discrete peaks on gamma logs and the collective data comprising the sequence model.

Line I-I' crosses a portion of the Pine Island trough and shows a distinct increase in the thickness of the post R-3 depositional package. This contains a distal, thinning portion of Sequence C and a major portion of Sequence D (excluding the

most proximal and distal units. The medial portion of Sequence D is represented by 38 m of Well 11 (North Captiva). The basal part of that section is a slightly shelly, phosphatic, sand/gravel lag that grades upward into a 30 m thick, greenish gray, dolomitic calcilutite. Sequence D rapidly thins to the south, although the section and bounding unconformity are missing from the adjacent well 18.

Continuation of Sequence D to Well 19 (Sanibel Island) is based on the seismic model and tentatively correlated to a phosphatic, dolomitic, sand. This sand unit may alternatively be identified as the basal portion of the overlying sandy clay to dolomitic interbeds of Sequence E. The updip portion of Sequence E is present in Well 11 as a 5 m thick, gravelly sand to sandy limestone couplet and Well 18 as a 29 m thick, grayish olive calcilutite to 9 m thick white to light orange dolomite.

Cross section II-II' illustrates a series of well logs beginning with Well 21 on Bokelia and going south down Pine Island (Wells 10, 14 and 17) and then to Well 20 on southeastern tip of Sanibel Island. This section contains portions of Sequences C, D, E and F. The distal portion of Sequence C is contained in Wells 21 and 10 as 10 m units of olive green shelly clay and grayish, dolomitic sand (respectively).

Sequence D is present in all wells along this section beginning with sand or sandy clay to dolomitic clay couplets in wells 21 and 10. The medial portion of Sequence D in Well 14 consists of a 28 m thick, olive colored dolomitic clay overlain

by 5 m of thick, light orange, dolomitic limestone which thins rapidly to a 6 m thick unit of shelly, clayey sand in Well 17 and less than 4 m of clayey, phosphatic sand in Well 20.

Sequence E is somewhat problematic along Section II-III' as indicated by the subcrop map of Figure 22. That figure shows the updip limit of Sequence D extending as a southeastward trending tongue from southern Cayo Costa Island across Pine Island Sound to central Pine Island. At northern Pine Island and Bokelia, Sequence E is present at well 21 (Figure 38) as a sandy, clayey shell hash and at well 10 as a sandy limestone. Sequence E thickens under southern Pine Island to 32 m in thickness and is present as dolomitic and/or limey clay overlain by dolomitic to limey sand (wells 14 and 17).

At well 20 on southeastern Sanibel Island this sequence is approximately 34 m thick and contains all three end-member lithologies; Phosphatic clayey sand, overlain by shelly, sandy clay, in turn overlain by poorly to moderately indurated, sandy limestone. The Sequence E transitions described at wells 14, 17 and 20 indicate a general sand-clay-limestone sequence with the exception of a missing sand facies in Well 14. This general transition is masked by differing descriptive terminology which shows the clay facies as clayey dolosilt in well 14, sandy to limey clay in well 17 and poorly indurated, clayey limestone in well 20.

The east-west cross section III-III' connects the two north-south sections via wells 19 and 20 on Sanibel Island and

continues east with well 22 on the Sanibel Causeway and wells 15 and 16 along the Caloosahatchee River. This cross section includes Sequences D, E and F with the distal portion of Sequence D previously described from wells 19 and 20. Sequence D in well 20 is lithologically similar to the basal part of the overlying Sequence E, i.e., a clayey, phosphatic sand. A sequence boundary is placed within that sand unit on the basis of the seismic model (Appendix II) and the well 20 gamma log (Appendix III) which has large peaks at 63 m and at 58 m.

Sequence E is also described from wells 19 and 20 and continues to the east with no apparent thinning. The east-northeast orientation of the cross section parallels the subcrop boundaries (Figure 22) and represents a strike section of Sequence E. In wells 15 and 16 Sequence E is composed of a gravel/sand lag gradationally overlain by a calcilutite described as either dolomite or clay.

DISCUSSION

The results of this study indicate that the late Neogene to Quaternary stratigraphic development of the Charlotte Harbor area has occurred as a series of lithologically heterogeneous, unconformity-bounded depositional sequences. Although the Neogene sequences are poorly constrained with respect to absolute dating, they occur as a morphologically distinct sequences of prograding oblique clinoforms bounded by regionally mappable unconformity surfaces (Figure 22).

In contrast, the Quaternary sequences are relatively well constrained chronologically by radiometrically calibrated amino acid dates (Evans, et al., in press). Morphologically, the Quaternary sequences consist of laterally discontinuous, downcutting channels and thin, chaotic to transparent seismic facies (Figures 20 and 21). This discussion will: 1) examine the antecedent topography, chronostratigraphy, and paleoenvironmental interpretations of each of the distinct sets of sequences, 2) compare the respective sequence morphologies, lithologies and overall stratigraphic development of the Charlotte Harbor area, and 3) briefly discuss the hydrogeologic significance of interpreting southwest Florida stratigraphy as eustatically controlled, lithologically heterogeneous depositional sequences.

Late Neogene Sequences

Antecedent Topography

The high resolution seismic reflection profiles of this study have defined two regional unconformity surfaces that can be mapped throughout the study area. Structure contours of the lowermost of those surfaces (R-3; Figure 14) indicate that the Charlotte Harbor area is underlain by a rectilinear pattern of troughs that are at least 50-60 km in length, 8-10 km in width and 40-50 m deep. Sinkholes of 10's to 100's of meters diameter and depths of 10 to >50 m are also present throughout the area and may, in some cases, be aligned with the troughs. The NE-SW and NW-SE orientation of the troughs correlates with the regional pattern of fracture/lineament trends (White, 1970; Culbreth, 1988).

The structure or relief comprising the antecedent topography of the R-3 surface is attributed to karstic dissolution of the underlying late Paleogene and Neogene limestones. The enclosed basins or troughs have developed a rectilinear pattern due to fracture selective dissolution and would be classified as uvalas using karst terminology (Bögli, 1980; Trudgill, 1985).

Culbreth (1988) mapped surface lineaments (i.e., linear patterns of vegetation, sinkhole alignment, stream channels, etc.) throughout south Florida. On the basis of gravity anomalies, Culbreth attributed the origination of the linear surface features as zones of basement weakness and the NE-SW and NW-SE orientations as an artifact of the Earth tide stress

field acting on those zones of weakness.

Similar rectilinear orientations and patterns of karst distribution have been reported by Hutton, et al., (1984), and Popenoe, et al., (1984) for west central and northeast Florida (respectively). Hutton, et al., describe the evolution of Holocene paralic environments on a surficially karstified Eocene limestone with streams and marsh islands localized by fracture controlled dissolution. Popenoe, et al., (1984) use seismic reflection data to map the distribution of open and infilled sinkholes off the northeast Florida coast and attribute their locations and alignment to regional fracture patterns.

Evidence of widespread subaerial or meteoric exposure necessary for carbonate dissolution is common in both the seismic and lithologic data of the Charlotte Harbor area (i.e., Figures 15 and 23). Vertically offset (and laterally adjacent) sinkholes also indicate that dissolution has occurred episodically (Figure 15). Alternating periods of carbonate deposition and dissolution are a direct result of sea-level fluctuations on a low lying carbonate platform such as peninsular Florida (Perkins, 1977; Popenoe et al., 1984; Choquette and James, 1987).

The creation of karst requires that: 1) the meteoric fluids must be undersaturated with respect to CaCO_3 , and 2) a fluid flow system must exist to transport dissolution products and aggressive fluid to and from the reaction sites. With these constraints there are several environments where karstic

dissolution can occur (Lohmann, 1987): 1) direct subaerial or vadose exposure and dissolution by CO_2 enriched rainwater, 2) exposure to meteoric phreatic waters with enhanced CO_2 due to bacterial reduction of soil organics, and 3) the coastal margin of meteoric phreatic systems which contains the seawater-freshwater mixing zone.

There is direct or indirect evidence that each of these environments has occurred within the depositional framework of the study area. Subaerial exposure is indicated by the very sharp, irregular seismic reflections of Figure 15 (Popenoe et al., 1984; Rasmussen and Neumann, 1987), and the blackened limestones of Wells 5, 13, 21, and 22 (Shinn and Lidz, 1987). Additionally, orange colored Fe stains indicative of paleosols or oxic water-table conditions are common in most of the well logs.

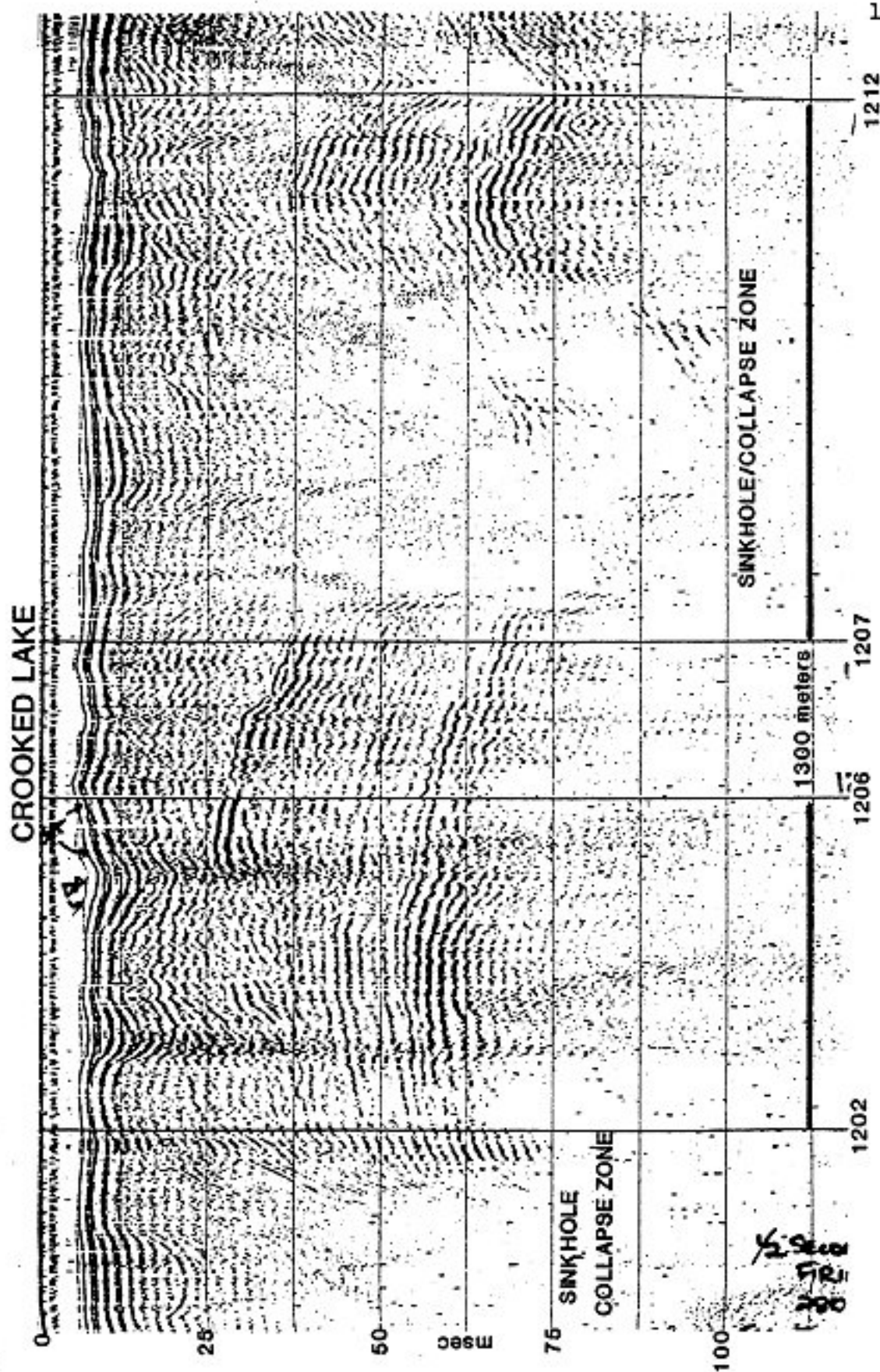
Wells 3, 5, 9, 10, 19, and 21 contain dark brown to gray zones that probably represent organic-rich deposits capable of enhancing meteoric CO_2 . Although karst dissolution does not necessarily follow the occurrence of these beds, in Wells 3, 10 and 21 these dark, peat-like deposits overly unconformable limestone contacts and in Well 10 distinct, macroscopic root material is noted on the log at ~70 m.

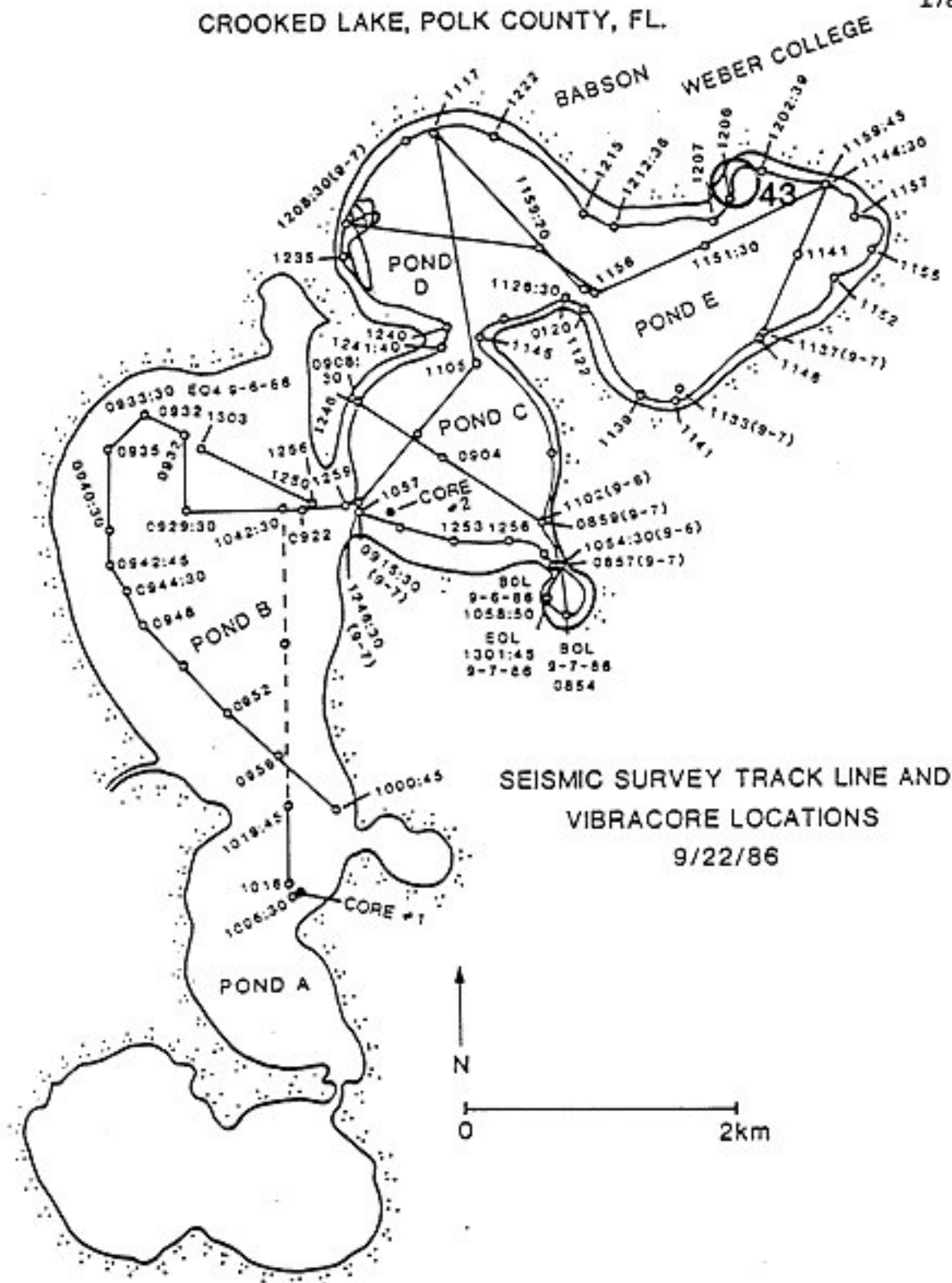
Evidence of saltwater-freshwater mixing zone dissolution is indirect but definitely indicates the presence of fresh and saltwater zones. The geometry and distribution of the seismic sequences indicates multiple, depositional cycles. Likewise the lithologic data indicates gradational deposition of marine

to paralic sediments that prove that marine environments migrated periodically through the area. The identification of estuarine and fluvial paleoenvironments (Missimer and Gardner, 1976; Peck, et al., 1979a) indicates that freshwater meteoric environments must also have been present, and consequently the seawater-freshwater mixing zone migrated through the area in concert with sea-level fluctuations (Peck et al., 1979b).

The R-3 unconformity probably does not represent the primary or initial karst episode responsible for the formation of the trough system. The 100 to 120 m depth of the R-3 unconformity within the troughs is near the limit of penetration of boomer type devices (E.G. & G. Uniboom Manual) and although that surface represents a strong erosional unconformity (Figure 17), no karstified surfaces or collapse zones were observed within the troughs.

Seismic data from Crooked Lake, Polk County (Figure 40) shows that large karst shafts do occur on the Florida Platform and are overlain and infilled with sequences of prograding clinoforms that are geometrically and lithologically analogous to the late Neogene sequences of Charlotte Harbor (Hine, et al., 1986). The morphology and orientation of Crooked Lake (Figure 42) is strikingly similar to the R-3 structure contour (Figure 14) which indicates that the antecedent topography of both systems may be created from similar fracture selective karst processes.





Carbonate dissolution has occurred in the study area as evidenced by both seismic and lithologic unconformities. Although the exact process of that dissolution (i.e., subaerial exposure, mixing zone dissolution, etc.) cannot be determined without detailed mineralogic and isotopic data, the stratigraphic data indicate that any or all of the three processes of karstification were environmentally possible.

Alternate explanations for the antecedent topography of the southwest Florida area have invoked dip-slip faults as recent as late Pliocene or Pleistocene (Tanner, 1965; Bond, et al., 1981) and/or differential subsidence or compaction caused by "tensional basal displacement" (Brooks, 1972; Missimer and Gardner, 1976). The bases for these interpretations are: 1) the rectilinear orientations of stream beds and shorelines (Tanner, 1965; Bond et al., 1981), 2) rapid lateral and vertical variability of seismic facies (Missimer and Gardner, 1976), and 3) an inability to correlate adjacent boreholes due to rapidly thickening and thinning of lithologic units (Brooks, 1972).

There is no indication of active or recently active offsets due to dip-slip faulting in the seismic profiles. Bond, et al. (1981) have postulated vertical offsets of up to 10 m in the southwest Florida area. The >100 m vertical relief of the R-3 unconformity, which is an order of magnitude greater than previously known, and the topographic closure of the lows argue against Neogene faulting. Additionally, the small size of the enclosed troughs (60 km x 10 km) and the overall

thickness of the underlying Paleogene and Neogene carbonates make basement structural adjustments or displacement difficult to support.

The seismic data of Missimer and Gardner (1976) are used to suggest that strata between the R-3 and R-2 surfaces (this study) are folded by "basal tensional displacement". Those profiles and continuations of those profiles from this study do not allow sufficient penetration to evaluate "basal structure".

Widespread mapping of the "folded" units indicates a consistent facies change from depositional locations outside the troughs to within the troughs. Such a facies change would not occur if the thickening and thinning were due to post-depositional folding.

The results of this study can only evaluate structural faulting and displacement where other investigators have proposed such structures in the late Neogene section. The data do not support such displacement but do indicate that fracture selective dissolution is responsible for the antecedent topography of the study area. The possibility of deeper structures is not precluded by this data and the coincidence of deep structural features (i.e., Klitgord, et al., 1984; Culbreth, 1988) with the karst troughs suggests that Cretaceous or Paleogene structures and the Earth tide stress field may be responsible for establishing the regional fracture pattern which is propagated laterally and vertically by fracture selective karst dissolution.

The distribution and preservation of the lithologic

sequences illustrated in the well logs of Appendix 3 are significantly modified by the karstified antecedent topography. For example, the linear troughs localized the developing fluvial systems such that the sand/gravel channel lags are effectively limited to the vicinity of the troughs and sand dominated units thicken dramatically to the east (Peck et al., 1979b).

Similarly, the calcilutite, estuarine deposits thicken within the troughs and effectively infill them during the D and E depositional cycles (i.e., wells 11, 14, and 18). In wells where the sand/gravel lag is missing, it appears that downcutting and erosion by estuarine channels may have removed the sand (if it was present; i.e., wells 2, 5, 8 and 21). The variable distribution of the end-member lithologies indicates that the antecedent topography defined by the R-3 structure contour has played a major role in controlling depositional environments.

Seismic and Lithologic Sequence Correlation

The controlling variable in the seismic-lithologic correlations of Figures 38 and 39 and Cross Sections I, II, and III is the velocity used to translate seismic time-of-travel (TT) to vertical distance measurements (m). The velocity measurements derived from the sonic logs are too high by a factor of ~2x according to published velocities (Table 2) and seismic and sonic log data from other studies in southwest

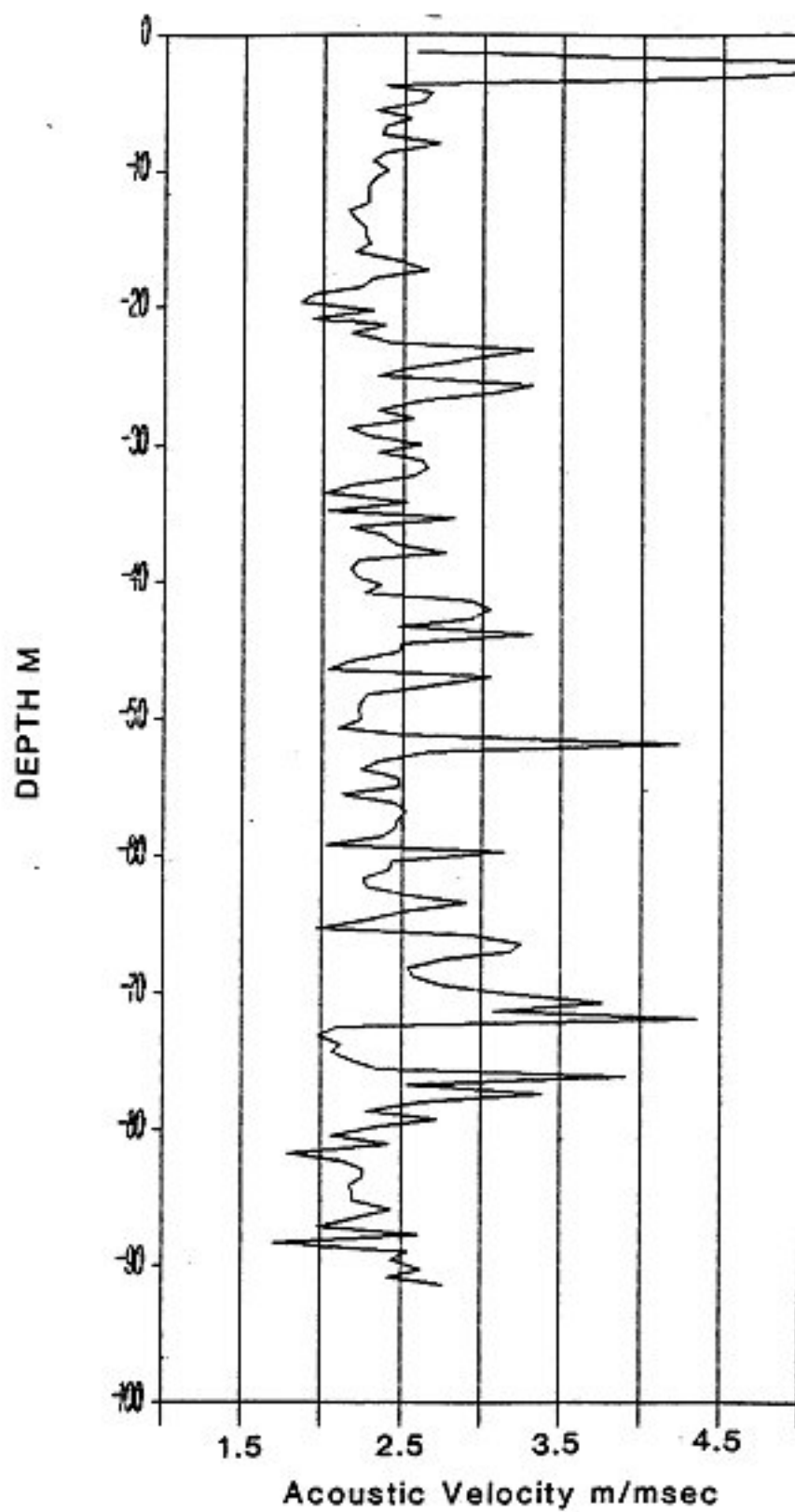
Florida. Figure 42 is a sonic log from a well in Nokomis, Florida (approximately 60 km north of study area). This log exhibits velocities that range from 2.0 to 4.0 m/msec and average ~2.3 m/msec as compared to the ~4.5 m/msec of the well 21 and 22 sonic logs (Appendix 3).

Missimer and Gardner (1976) correlated seismic reflectors to lithologic unconformities for sites along the Caloosahatchee River and derived velocities of 1.98 to 2.59 m/msec for the Tamiami Formation. Ball *et al.* (1988), calculated velocities of 2.2 m/msec for the upper second of penetration on the west Florida shelf and ~3.8 m/msec for Paleogene carbonates from deep penetrating common depth point seismic data.

Each of these calculated or estimated velocities indicates that the late Miocene deposits of south Florida have sonic or acoustic velocities between 2.2 and 2.5 m/msec. Lower velocities of the water column (~1.5 m/msec) and the unconsolidated Quaternary deposits (~1.7 m/msec) should reduce average velocities to near the assumed value of 2.0 m/msec. The essential test of that assumption is whether seismic unconformities can be correlated with lithologic unconformities.

The direct sonic log velocities and reflection events calculated from those logs are limited to two sites and by an obstruction at one of those sites. While recognizing the limitations of those direct velocity analyses, area-wide correlations of the predicted seismic unconformities with lithologic unconformities (cross sections I-III) validates the

VELOCITY LOG WELL TR5-2 NOKOMIS, FLORIDA ¹⁸⁴



assumed 2.0 m/msec velocity and indicates that the mapped seismic sequences represent real unconformity-bounded lithologic sequences.

The correlation of lithologic unconformities with seismically mapped regional unconformities and local disconformities indicates that lithologically similar, laterally and vertically adjacent, depositional units occur between sequence bounding unconformities. Because these units are bounded by mappable unconformities and/or disconformities, they represent discrete depositional cycles such that Walther's Law of Correlation of Facies is not valid between adjacent or superjacent depositional sequences.

Although the definition of lithostratigraphic formations implies no chronostratigraphic consistency within a formation, the presence of identifiable and mappable unconformities within defined formations precludes paleoenvironmental interpretations based on lateral facies transitions and indicates potential problems when correlating borehole data.

Chronostratigraphy

Using lithostratigraphic terminology the unit represented by the depositional sequences A-F is the Tamiami Formation of Peck et al. (1979a/b), the Hawthorn Formation of Wedderburn et al. (1982) or the Peace River Formation of Scott (1988). Cross section III-III' contains Well 16 (L-1849) which was biostratigraphically analyzed by Peck (1976). Several late

Miocene indicator fossils were identified from that site, including: Discoaster quinqueramus, D. berggrenii, D. brouweri, Reticulofenestra pseudumbilica, and Globorotalia acostaensis. These species are indicative of foram zones N.17/N.18 and Discoaster quinqueramus and D. berggrenii are limited to those zones (Blow, 1969).

Figure 43 is a portion of a sea-level and sequence chart from Haq et al. (1987). That portion of the chart corresponding to foram zones N.17/18 is the late Miocene Tortonian Stage. The eustatic curve for this time period shows three third order sea-level high stands separated by major low stands at 5.5 and 10 MA. The high stands range from near present sea-level to approximately 50 m greater than present.

The depositional sequences defined in this study cannot be correlated to specific sea-level events due to a lack of biostratigraphically or otherwise dated deposits. A general, late Miocene Tortonian age is reasonably well established for these deposits (Schroeder and Bishop, 1953; Peck, 1976; Peck, et al., 1979b) and the strong regional unconformities bounding the sequences suggests correlation to the 5.5 and 10 MA low stands (Haq, et al., 1987).

The chronostratigraphic discrimination of the individual sequences is well beyond the resolution of biostratigraphic or most absolute dating techniques (Peck, et al., 1979b; Snyder et al., in press). Late Neogene glacio-eustatic sea-level cyclicity is relatively well established (Peck et al., 1979a; Matthews, 1984; Crowell, 1982; Snyder et al., in press).

Although Milankovitch period climatic fluctuations (40-100,000 KA cycles) and associated eustatic fluctuations are the best explanation for the late Miocene depositional sequences, this study contains no specific or absolute chronostratigraphic evidence that glacio-eustatic fluctuations have driven the Pine Island sequences.

Paleoenvironmental Interpretation

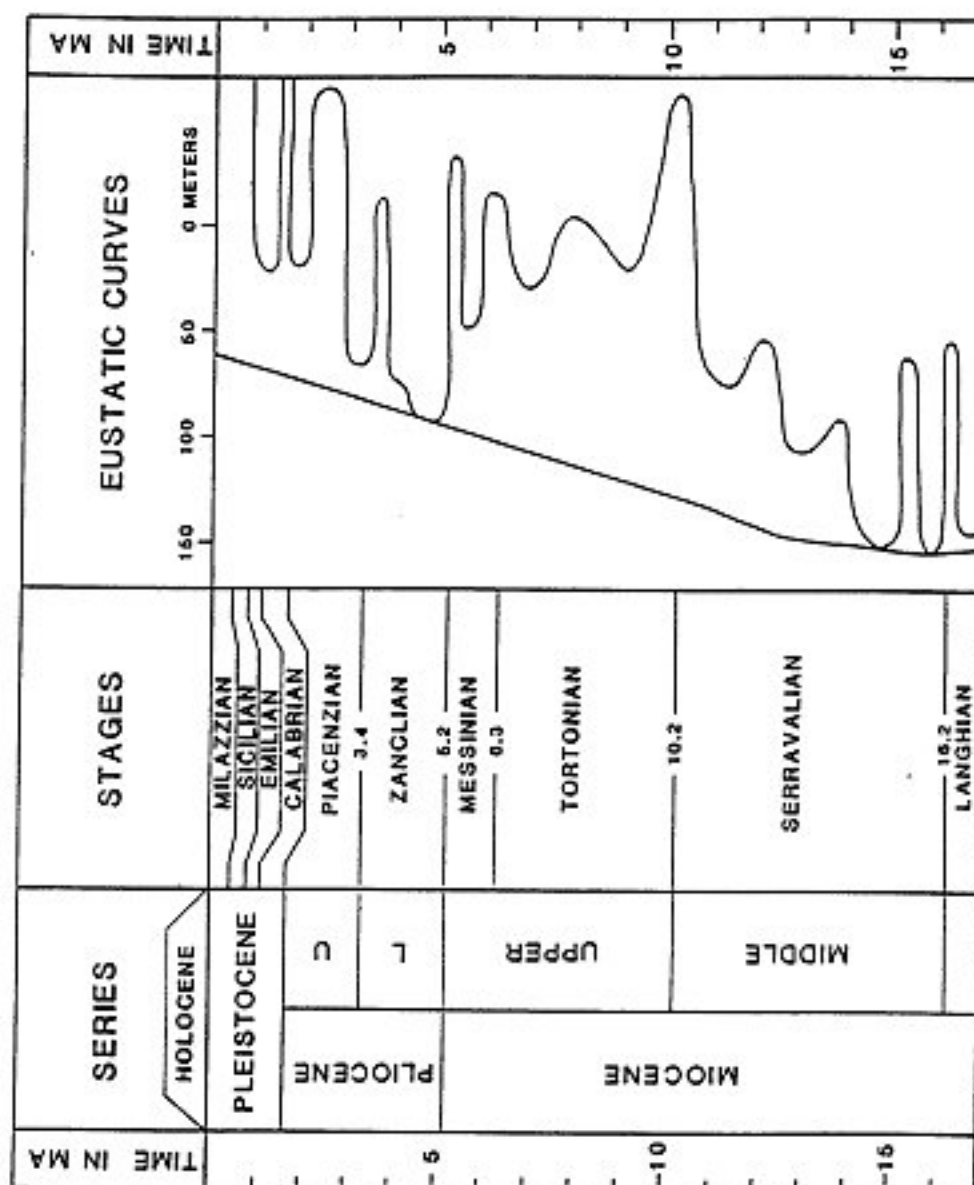
Six seismic sequences, herein termed the Pine Island sequences, have been identified between the R-3 and R-2 regional unconformities (Figure 22). These sequences prograde to the south-southeast and each contain a distinct parallel to oblique clinoform geometry (i.e., Figures 16, 17 and 18). Well-cuttings and three continuous cores indicate that there are three end-member lithologies within these sequences consisting of: 1) a usually thin, quartz dominated, phosphatic gravelly sand, which is gradationally overlain by, 2) dolomitic, calcite/aragonite clay interbedded with thin laminae of sand or silt, with an upward transition to, 3) variably indurated, molluscan/foram calcarenite with variable amounts of quartz sand which is often characterized by a strong erosional upper surface.

Figure 44 presents a conceptual model of transgressive depositional environments which may have produced the lateral facies relationships and vertical sequences. This block diagram shows a low-energy coastal transition from fluvially

derived sand/gravel channel lags to estuarine or inner shelf laminated muds (calcilutite), to an open marine shelf dominated by the accumulation of molluscan/foram hash. The horizontal and vertical scales are derived from the thickness and extent of the seismic and lithologic sequences.

This model (Figure 44) necessarily implies that each sequence represents a transgression of shallow or intertidal environments by deeper marine environments. Similar, transgressive paralic to inner shelf sequences have been described by Riggs, et al. (1982), Riggs (1984) from seismic and lithologic data. The transgressive nature of the middle Miocene Pungo River deposits of North Carolina is provided by benthic and planktic foram analyses (Katrosh and Snyder, 1982) and post-depositional boring by marine bivalves (Riggs, 1984).

Kidwell (1988) describes both transgressive and regressive depositional sequences in the middle and late Miocene of coastal Maryland from semi-continuous outcrop exposure (5-20 km exposure with 3-5 km gaps). Transgressive to regressive couplets were defined within the Burdigalian to Serravallian age Calvert and Choptank Formations as sandy shell/bone beds representing condensed sections overlain by fining, then coarsening upward clay, silt or sand. The Tortonian cycles were interpreted as transgressive units based on the presence of freshwater and estuarine fossils and trace fossils which were transitionally overlain by marine facies. In both North Carolina and Maryland, the transgressive depositional cycles were contained within an overall basinward

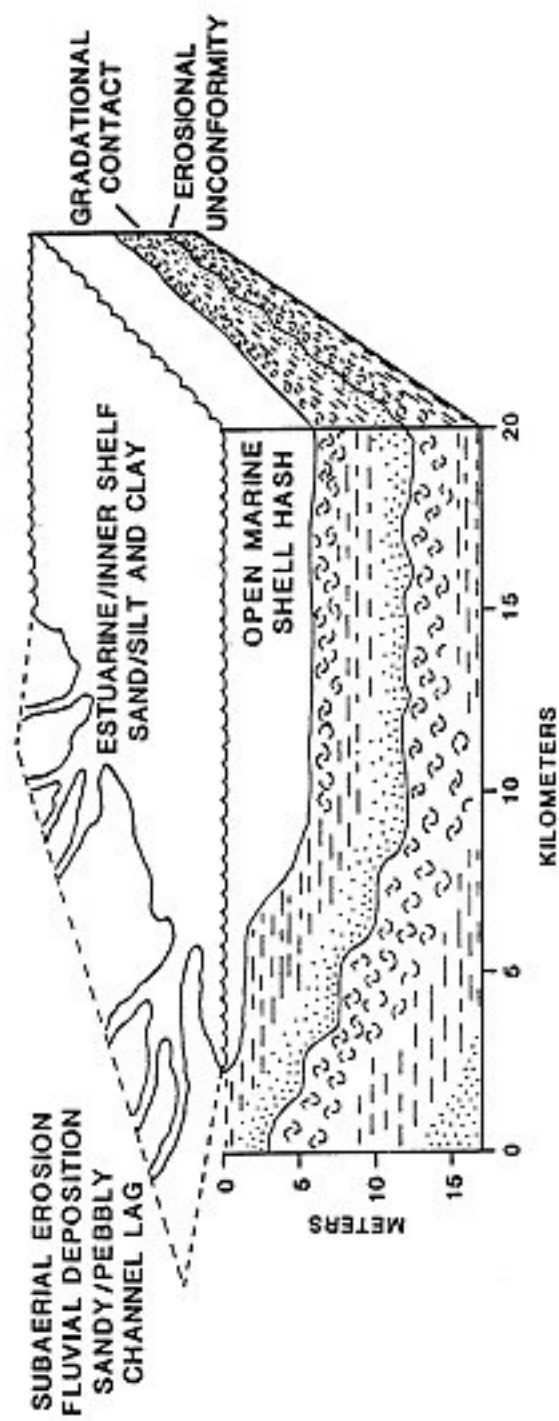


progradation of coastal environments.

An alternative depositional model based on Holocene prograding chenier/delta plains of the northern Australian coast (Rhodes, 1982) is presented in Figure 45. This model necessarily includes the same lithologic end-members and differs from the transgressive model primarily by identifying the sandy molluscan hash as subaerial chenier ridges (as opposed to shallow marine shelf). Rhodes (1982) documents the depositional environments and facies geometry of the prograding chenier coast of the Gulf of Carpentaria which are presented as Table 8.

The Gulf of Carpentaria is a semi-enclosed epicontinental sea with generally short steep waves generated locally and aperiodic high energy cyclones (Rhodes, 1982). Drainage and sediment supply into the Gulf is provided by a number of small short rivers with rainfall and runoff highly seasonal. The tidal range in the Gulf of Carpentaria is 2-3 times that of southwest Florida, however, Rhodes states that wave activity is important in modifying the coastal deposits and directly controls the deposition of the chenier ridges. Although the specific paleoenvironmental conditions of southwest Florida during late Neogene time are unknown, the modern conditions of southwest Florida are similar in several respects to the Gulf of Carpentaria and it may be instructive to examine the depositional environments of Table 8 as modern analogues to the Charlotte Harbor depositional sequences.

The fluvial environments consist of medium to coarse



quartz sands with pisolitic or lateritic gravel that occur as variable lags to deposits several meters thick (Table 8). Prograding blue-green subtidal muds that overlie the fluvial unit are continuous along the coast and contain up to 50 % carbonate. The subtidal muds are gradationally overlain by gray-green to brown low tide muds that are interbedded with sand layers.

Discontinuous organic muds may overlie the low tide muds or the unit grades directly into the shelly, chenier ridges (up to 75 % marine bioclastic detritus). Although these ridges are laterally discontinuous, Rhodes states that they are often flattened and transported landward during high energy events such that the preserved sandy calcarenites are more widespread than the discrete shell ridges. The shallowing upwards, regressive deposition described by Rhodes (1982) has occurred during the late Holocene sea-level transgression and consequently requires a large sediment supply (Brown and Fisher, 1980).

Several aspects of the Pine Island sequences indicate that the Gulf of Carpentaria chenier plain model may be applicable to the Pine Island sequences. First, Peck, et al. (1979b) described the sandy calcarenite units to be of hyposaline rather than normal marine origin. Secondly, the presence of minor but ubiquitous quartz sand in these units is difficult to explain in a carbonate shelf setting which is basinward of fine-grained terrigenous deposition. The limestone or semi-indurated mollusc beds are commonly overlain

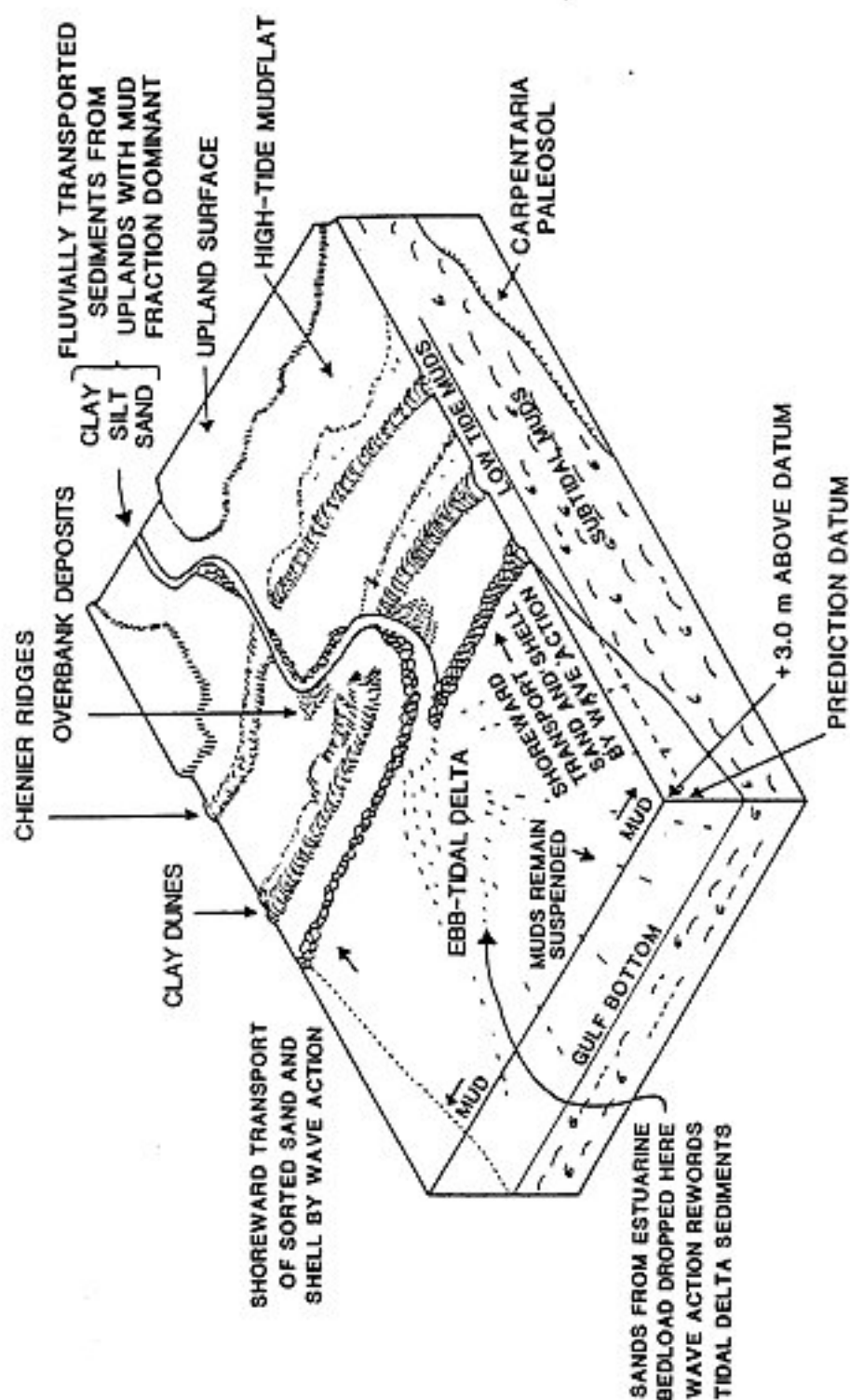


Table 8. Depositional Environments along a prograding chenier coast, northern Australia (from Rhodes, 1982).

ENVIRONMENT	FACIES	GEOMETRY	LITHOLOGY	SEDIMENTARY STRUCTURES
Fluvial	Upland channel sands	Variable from a thin gravel bed to a thick sand deposit several m thick	Medium to coarse quartzitic sandstone, laterite gravel, 2-3 cm diameter	Poorly laminated, occasional sand waves and scour mega-ripples
Mudflat	High-tide muds	Approx. 1-2 m thick, continuous lobe up to 50 km wide, graded downward to underlying low-tide muds	Deeply cracked and oxidized brown muds of silty clay, 2-3% insoluble salts, 2-3% insoluble organics, less than 1-0% CaCO ₃	Surfaces heavily scoured and fine worked, laminar, disapparing downward due to salt prescence, wetting drying
	Intertidal organic muds	Vegetation zone 100-300 m wide, continuous up to 20 km	Fine silt and clay with 2-3% organic matter, textural similarity to low-tide muds	Gastropod tracks
	Low-tide muds	1-2 m thick, but increasing in thickness seaward, continuous up to 20 km	Light to dark brown muds, mixed with fine silty sands, up to 40% clay in size	Flaser and wavy bedding with fine sand and clay, symmetrical and asymmetrical ripple marks
Strandline	Chenier ridge sands	150 m thick, 50-150 m wide, continuous up to 20 km	Up to 75% up to 75% molluscan shell with variable silt, sand and pisolitic gravel	Horizontal bedded, steep cross-bedding near back of ridge, gentle forward dip at front of ridge
Nearshore	Ebb-tidal delta sands	Lobe complex, 2000 m thick, 1000 m wide, 1000 m long	Medium shelly quartzic sands with CaCO ₃ variable (1-80%)	Massive composite bedding of silt and sand, layers, occasional scour mega-ripples
	Sub-tidal muds	Prism-shaped lobe up to 8 m thick, variable width to 10 km, continuous along coast	Blue-green muds with variable sand, silt and clay, median size 9-10 phi, CaCO ₃ varies, 1-50%	Not observed
Dune	Aeolian silts and clays	2-3 m thick, dune field may be 10 km wide and 10 km long	Light brown silt and clay with some sand, some soluble salts up to 25% organics less than 1-0%	Usually homogenous with some inherited micro-structure

by dark brown to gray organic-rich peatlike deposits indicative of intertidal conditions (i.e., wells 3, 5, 9, 10, 19 and 21).

The progradational nature of these sequences is readily apparent in the seismic profiles (i.e., Figures 19 and 22) and if subsequent transgression by open marine facies had occurred the stratal reflectors should indicate a transition from offlap to onlapping relationships. Reflector onlaps are only apparent at the updip limit of each sequence (except sequence E) and are rapidly transitional to the prograding offlap configuration. The parallel to oblique reflectors, consistent toplap configurations and dip-strike variability (Figures 15, 16, 19 and 22 and cross sections A-H, Appendix I) identify these sequences as prodelta deposits (Mitchum, et al., 1977; Brown and Fisher, 1980) which are conceptually similar to the prograding subtidal muds of the chenier plain (Rhodes, 1982).

The source of the and mineralogy of the clays in the prodelta facies are presently unanswered questions. Peck (1976) indicates that the clayey facies are primarily composed of silt to clay sized calcite and aragonite; Scott (1988) also states that the clays are calcareous to dolomitic, but concludes that montmorillonite and palygorskite are the primary clay minerals. The ultimate source of the clay minerals and dolomite may be in situ precipitation (Reik, 1982; and Baker and Burns, 1985; respectively) or weathering (Altschuler et al., 1964). Whatever the source, the external facies morphology and interbedded nature of the clayey facies indicates sub-aqueous, clastic deposition (Blatt et al., 1980).

The interpretation of the late Miocene environments as sea-level controlled, depositional sequences is well supported by micropaleontologic studies of Peck (1976) and Peck et al. (1979b). Those studies used benthic forams, calcareous nannofossils and diatoms to define fluvial, estuarine and open marine depositional environments and specifically interpreted the sandy calcarenite as a hyposaline deposit. Peck, et al. (1979a) used the geometry and cyclic nature of the sequences to show that the depositional sequences were created by glacio-eustatic sea-level fluctuations.

Examination of the lithologic cross sections shows that at any specific well site only two to four of the six depositional sequences are vertically juxtaposed due to the lateral progradation of successive sequences. Peck, et al. (1979a/b) recognized three transgressive-regressive cycles from well data and Missimer and Banks (1982) have tentatively added a fourth. The seismic data conclusively demonstrate that at least six sequences are present. Acknowledging the difficulty in correlating widely space borehole data, it is more important that eustatically controlled cycles were recognized rather than the determination of the specific number of such cycles.

Missimer and Banks (1982) used closely spaced well logs from Sanibel Island to interpret the cyclic nature of these units as a series of upward shoaling, regressive deposits. They state that the limestone or molluscan calcarenite is the base of each sequence with subsequent encroachment of estuarine/deltaic deposits. If the limestones are interpreted

as the basal facies of a regressive depositional sequence, unconformities or transgressive deposits should occur at the top of the sequence, i.e., overlying the fluvial or intertidal deposits. The common occurrence of indurated, subaerially exposed limestones overlain by a sand/gravel lag indicates that the limestones are uppermost facies of the depositional sequences and not the basal unit as described by Missimer and Banks (1982).

The individual sequences probably represent transgressive and regressive deposits similar to Kidwell (1988). The sand/gravel lag in northern Australia was interpreted by Rhodes (1982) to represent a transgressed fluvial/subaerial surface. As such, the Pine Island sequences represent a parasequence set as described by Vail (1987). Parasequences are "relatively conformable succession(s) of genetically related beds or bedsets bounded by marine flooding surfaces...Parasequences are progradational" (Van Wagoner *et al.*, 1987).

There is also a distinct morphologic difference between the short, steep, parallel clinoforms of sequences A-D and the thicker, oblique clinoforms of sequence E (Figure 25 and cross sections E-F). Whether this difference is due solely to the influence of the karstified topography or changes in sediment supply and the rate of relative sea-level change is unknown. It is possible that sequence E represents the regressive portion of a transgressive-regressive couplet with the underlying sequence D. Such a couplet would produce a partial unconformity separating the units (Nummedal and Swift, 1987;

Vail, 1987).

Biostratigraphic analyses which should clarify the interpretations of the Pine Island sequences are underway (V. Waters, personal communication). However, based on the strong progradational morphology and the lithologic composition of the Pine Island sequences, these units are interpreted as analogues to the prograding chenier plains of northern Australia. This interpretation reconciles the vertical lithologic sequences with the lateral distribution of seismic and litho-facies, and because of the lateral and vertical stacking of subtidal muds and subaerial cheniers requires depositional control by eustatic sea-level fluctuations.

Quaternary Sequences

Antecedent Topography

The R-2 structure contour (Figure 18) indicates that the karst topography has also played a significant role in the distribution of the Quaternary sequences. The late Neogene to Quaternary transition from a carbonate platform/ramp to a siliciclastic depositional environment is characterized by the localized development of fluvial systems within topographic lows (Figures 18 and 19). Only a few small downcutting channels were observed seaward of the barriers and the paleo-rivers apparently terminated in the Pine Island Sound trough (1985). The infilled lows contained an abundant erodable sediment supply which was locally reworked into the Quaternary

barrier/lagoon system.

The modern barrier system overlies a topographic high of unknown age with 2-4 m of vertical relief. Pine Island appears grossly similar in size and orientation to the high under the barriers. Using the terminology of Oertel (1985) these features could be considered barrier platforms. Both of these barrier platforms straddle the antecedent lows and the infilling paleo-fluvial deposits.

Numerous amino acid dates along the margins of Pine Island (P.I. Sound and Matlacha Pass cores) indicate that the Pine Island platform dates from the 220-250 KA high stand (isotopic stage 7). Although the platform underlying the modern barriers has not been dated, its location seaward of Pine Island and estimates of late Pleistocene sea-levels (Belknap, 1979; Cronin, *et al.*, 1981) suggest that it post-dates the Pine Island platform.

The persistence of these platforms as topographic features is due to two factors, 1) is the declining maximum transgression of late Pleistocene sea-levels (Demarest *et al.*, 1981; Cronin, *et al.*, 1981), and 2) is the concentration and lithification of carbonate-rich platform sediments. The cores show that inlet and overwash processes concentrate abundant gravel to sand-sized shells and shell fragments (Figures 30 and 32). Borings seaward of the barriers (U.S. Department of the Army, 1969) and on Pine Island (Wedderburn, *et al.*, 1982) indicate that earlier deposits are lithified. Numerous outcrops of beachrock have also been observed along these

barriers (Missimer, 1973; Hine, 1987).

Comparison of the post R-2 isopach map (Figure 35) and the total depth of dated Quaternary deposits in Table 6 indicates that there is apparently very little record of Pliocene deposition and limited preservation of Pleistocene deposition in the Charlotte Harbor area. The post R-2 sediment thickness is between 3 and 6 m for most of the study area and has depths > 6 m (1.7 m/msec acoustic velocity) only in the channelized areas of Pine Island Sound and the Harbor proper while Holocene sediment thicknesses are 3 m or more in 21 of 40 vibracores.

The relative lack of preservation of Plio/Pleistocene deposits is attributed to two factors: 1) the localization of estuarine and fluvial environments by antecedent topography has resulted in extensive channelization in a limited area, and 2) postulated high Pliocene sea-levels (Figure 43) subjected the Charlotte Harbor area to significant leveling and erosion by shoreface processes (Figures 20 and 21; Nummedal and Swift, 1987).

The initial creation of fluvial drainage on the carbonate platform was controlled by karstic relief which, in turn, controlled the development of the barrier/lagoon system through antecedent relief and sediment supply. There is no difference in the role of antecedent topography between this and the barrier/lagoon systems of the northern Gulf and middle Atlantic coasts. However, localization of the initial barrier system on antecedent lows (due to increased sediment supply) suggests

that there is a difference in the process by which antecedent control was manifested.

Paleoenvironmental Interpretation

The interpretation of the Quaternary depositional environments as recovered from the vibracores is well supported by comparison with modern environments. However, the specific interpretation of the shelly, FUS as tidal channel and/or tidal-delta deposits has been alternatively identified by Knowles and Davis (1983) and Bland (1985) as the result of storm overwash processes. They attempted to correlate basal "overwash" lags across distances up to 6 km using only subsurface vibracore data. Although overwash deposits are relatively common along the southwest Florida coast, they have rarely been quantified and are usually limited to the topographically low areas adjacent to the inlets (Andronaco, 1987; Hine, 1987).

The basal lags and FUS of overwash deposits undoubtedly contribute to deposition of barrier island sediments and may be an important factor in determining the size and frequency of shelly FUS. However, the presence of both Pleistocene and Holocene sequences and the discontinuous, channelized seismic reflectors indicate that correlation of individual sequences may not be possible. Boothroyd (1986) has shown that storm overwash and flood tidal delta processes exist as a continuum and it is more significant that, whatever the depositional

process, the sequences represent backbarrier lagoonal environments.

The distribution of the Holocene fluvial (point bar), estuarine and tidal channel sequences have a distinct transgressive geometry, i.e., tidal channel deposits overlie estuarine deposits which overlie point bar deposits (Table 6; Figure 31). Within the basic transgressive geometry of these sequences, the barrier islands have prograded seaward via beach ridge accumulation (Missimer, 1973). Lithologically, the prograded sequences consist primarily of quartz sand with variable molluscan carbonate (Missimer, 1973).

Examples of barrier island progradation during the Holocene transgression are common and usually explained as differential balances between relative sea-level rise and sediment supply (i.e., Wilkinson, 1975; Moslow, 1980; Evans, et al., 1985). Although the absolute progradation of southwest Florida barriers is small (1-5 km; Missimer, 1973) the barrier systems represent a significant accumulation and consolidation of quartz sand.

The gross lithology and infilling sequences of Charlotte Harbor are similar to those of northern Gulf and U.S. east coast estuaries and lagoons (Parker, 1959; Wilkinson, 1975; Kraft and John, 1979; Howard and Frey, 1985; and Boothroyd et al., 1985). The creation and lithification of barrier platforms should result in a different pattern of preservation from that described by Wilkinson (1975), Hine and Snyder (1985), and Belknap and Kraft (1985), for Texas, North Carolina and

Delaware, respectively. Infilled fluvial channels, which are below the depth of shoreface incision, are preserved on the inner shelf in all of those areas.

There are few downcut paleo-channels seaward of the Charlotte Harbor barriers because the fluvial systems terminated in the Pine Island Sound karstic depression. There are however, numerous, isolated, infilled sinkholes on the inner shelf. These structures are especially common along the trend of the NW-SE trough (Figure 14) and may represent the incremental enlargement of the trough system. While the ultimate preservation of Quaternary sequences depends on the maximum Holocene transgression, sediments deposited landward of the barrier platforms (up to the depth of the platform) should be protected from shoreface incision but subject to reworking by tidal and fluvial channels. Seaward of the barrier platform(s) sediments deposited in karstic sinkholes will be preserved in isolated conical depressions.

Comparison of Neogene and Quaternary Sequences

The Neogene to Quaternary depositional sequences defined and mapped in this study represent very different patterns of sediment accumulation. The late Neogene deposits contain at least six discrete depositional sequences which are up to 30 m thick, extend up to 55 km in a dip direction (Sequence E; Cross section G-G'') and are continuous in strike sections

throughout the study area (maximum distances of 40 km; Figure 22). The six dated Quaternary sequences (Figure 34) have a cumulative maximum thickness of 18 m, average between 3 and 6 m (cumulative thickness; Figure 35), and cannot be areally correlated within seismic profile grids of 3-5 km spacing (Figure 13). The lithologic end members of the Quaternary and Neogene sequences are similar, i.e., quartz sand, molluscan calcarenite and/or shelly FUS, and mud-sized clay or silt (both of questionable mineralogy). The duration of the Quaternary depositional cycles has been correlated to discrete glacio-eustatic sea-level fluctuations (Figure 34). Biostratigraphic dating of the Neogene sequences indicates that they are of Tortonian age (Peck, *et al.*, 1979b) with sequence durations possibly within Milankovitch periodicities.

The Neogene and Quaternary sequences represent different depositional environments, i.e., fluvially-dominated, prograding chenier plains versus tidal inlet dominated backbarrier lagoons. Quartz sand was introduced to the carbonate terrain of southwest Florida via southward flowing rivers which, during the late Miocene, terminated as the progradational Pine Island cheniers. Prior to the deposition of those sequences quartz sand is present only as a minor component of limestones and marls (Scott, 1988).

In addition to the transport of large quantities of quartz sand, the fluvial and deltaic systems of the late Neogene were responsible for the accumulation of relatively impermeable, widespread and locally thick carbonate clays which

prevented vertical drainage and reinforced the developing pattern of lateral runoff. With respect to the transition of the Florida Platform from a carbonate to a siliciclastic regime, the late Neogene prodelta and chenier environments represent a logical precursor to the Quaternary fluvial and estuarine environments.

The Quaternary environments are predominantly composed of quartz sand (although they contain significant quantities of carbonate as molluscan shell debris). Discrete channel structures which are not apparent in the Neogene deposits, are the primary preserved seismic facie in Quaternary deposits. Although the Neogene and Quaternary sequences are separated by 5 MY of non-deposition or erosion, they represent the gradual progradation of sand-rich fluvial systems into the study area and subsequent reworking by fluvial and coastal processes.

An important question raised by this scenario is that if the deltaic environments prograded during the late Neogene, then the fluvial environments should also have prograded across earlier deltaic deposits. This suggests that some of the downcutting channels ascribed to the Quaternary sequences may be contemporaneous with the Pine Island sequences. As these channel structures are isolated by subsequent channel erosion, there is no way to identify and correlate the channel structures with their synchronous depositional sequences short of absolute or biostratigraphic dating.

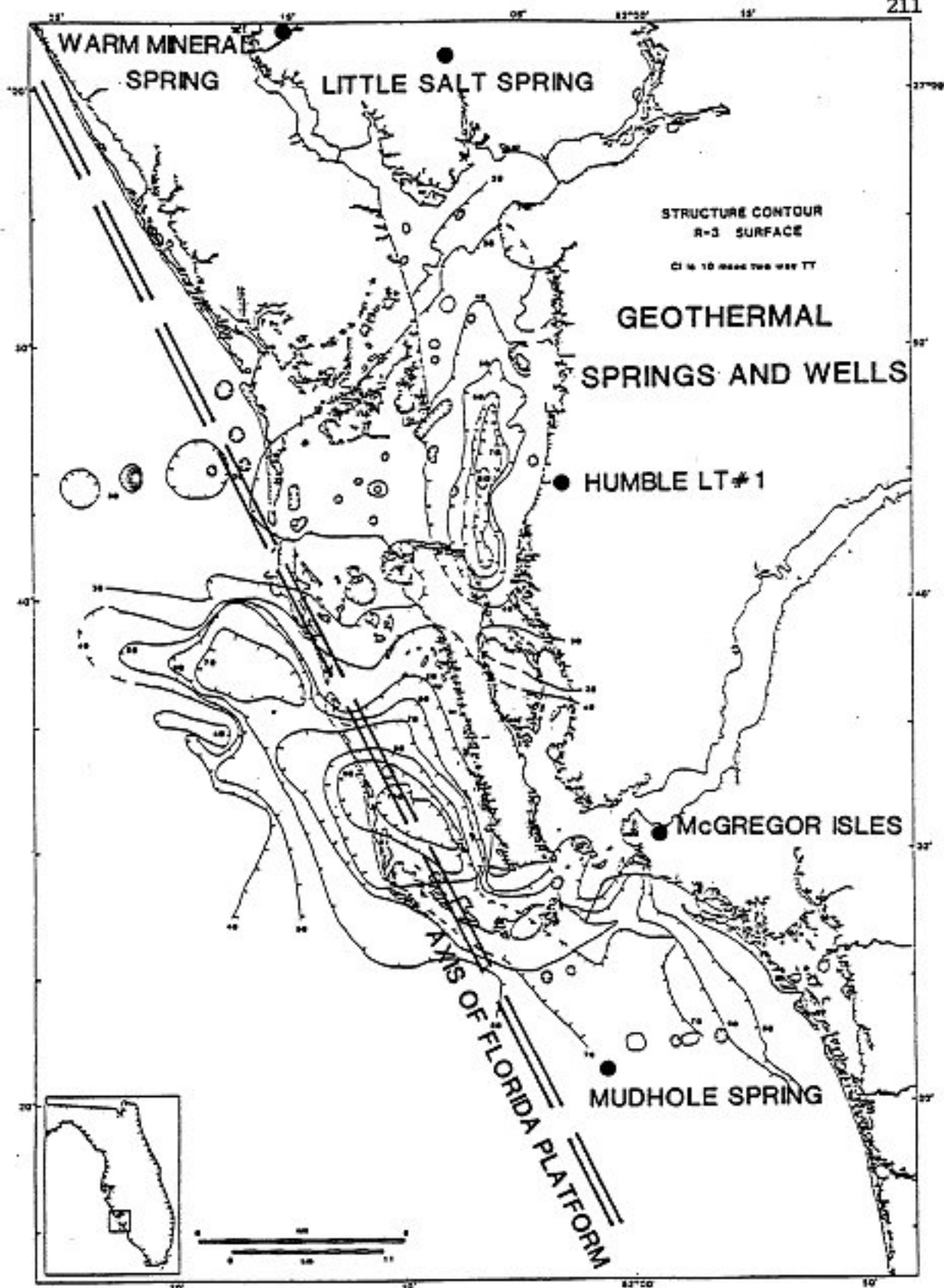
Antecedent topography has played a significant role in both the Neogene and Quaternary sequences by localizing

deposition and thus creating a relatively discrete source of sediment for subsequent reworking. The location of the karst system has thus created the southward and westward limit of quartz sand transport. Similar control by karstic topography at Tampa Bay and Lake Okeechobee may be responsible for limiting the westward and southward distribution of quartz sand and the subsequent physiographic development of peninsular Florida.

Hydrogeologic Significance

The antecedent topography and depositional sequences of southwest Florida as described by this study have several important implications for the distribution and integrity of hydrogeologic systems. The widespread distribution of karst means that vertical exchange between aquifer systems is probably the norm rather than the exception. The effect of such vertical exchange is illustrated by the presence of several saline, geothermal springs which discharge in the area (Figure 28). In addition to those springs, which occur at the extreme north and south margins of the study area, other sites of artesian, saline, geothermal aquifers are present along western Charlotte and Lee Counties (Figure 46; Sproul, et al., 1972; Roseneau, et al., 1977).

The large karst depressions which underlies Pine Island Sound and Charlotte Harbor are infilled with thick deposits of



impermeable clay. As indicated by this study, the deposition and distribution of each of the end member lithologies is controlled by the interaction of antecedent topography and sea-level cyclicity. Along much of the Florida peninsula, the hydrogeologic role of the middle to late Miocene deposits is to confine the underlying Floridan Aquifer System. In southwest Florida these deposits are still an important aquitard, but it functions to separate underlying upward-leaking saltwater from localized freshwater aquifers.

Aquifers within the late Miocene sequences, most notably the Sandstone aquifer of Wedderburn, et al. (1982), will be of limited areal distribution due to their origin from laterally restricted, high frequency, sea-level controlled depositional cycles. The additional limiting character of antecedent topography further constrains their distribution. Consequently, attempts to regionally correlate aquifers or aquitards developed within the high frequency sequences will be unsuccessful. However, the discrete pattern of karstified antecedent topography and the derived depositional model which reveals the distribution of permeable and impermeable facies provide a means of discovering and managing water resources.

CONCLUSIONS

The goals of this study are to document the late Miocene to Quaternary depositional sequences of southwestern Florida from continuous, high resolution seismic profiles and correlate the resulting sequence model with borehole lithologies and geophysical logs. The specific objectives to accomplish those goals are:

- 1) To define the pattern and role of karstification in creating small intra-platform basins;
- 2) Evaluate the pattern of karstification with respect to basement structure and regional fracture patterns;
- 3) Define and correlate the depositional environments and distribution of the seismic and lithologic sequences comprising the late Miocene strata of southwest Florida;
- 4) Document the role of karstified antecedent topography upon the development of the siliciclastic-dominated fluvial/estuarine environments of south Florida, and;
- 5) Examine the resulting depositional model with respect to the hydrogeologic system.

The carbonate deposits that comprise the Florida platform and underlie peninsular Florida periodically undergo subaerial and/or meteoric dissolution during sea-level low stands. In the Charlotte Harbor area that dissolution has taken the form of a rectilinear pattern of troughs that trend NW-SE and NE-SW. These troughs are 50-60 km in length, 5-8 km in width and have

vertical relief of up to 60 m below the adjacent topographic highs.

The troughs, or uvalas of karst terminology, are coincident with the regional fracture pattern and surficial geomorphic features, such as shoreline orientations, stream beds and aligned ponds or sloughs. The karst pattern is also coincident with deep structural features i.e., fracture zones and postulated hinge zones (Klitgord et al, 1984) as indicated by gravity and magnetic anomaly maps. No obvious displacement indicative of active faulting is present in the seismic profiles. It is suggested that the coincident patterns reflect vertical and horizontal propagation of structural weakness through fracture selective karstic dissolution.

The late Neogene depositional sequences which infill and overlie the karstified antecedent topography consist of six unconformity bounded units which prograde south-southeast with southerly dipping internal reflectors. These seismic sequences exhibit a distinct proximal-distal facies transition, have a strong offlap relationship with basal unconformities and minor onlap at their updip limits. Although there is significant variation in the external geometries depending on location relative to the karst troughs, each sequence is best described as a prograding oblique seismic facie.

The lithologies which comprise the seismic sequences consist of three end members: 1) quartz-dominated, phosphatic, pebbly sand, 2) a calcilutite interbedded with thin sand/silt layers and containing some molluscan shells and/or molds and,

3) variably indurated quartz sandy, molluscan/foram calcarenite. Unconformities are common and most frequently characterized by indurated, blackened limestone or cherts and sharp sonic and gamma peaks on associated geophysical logs.

Correlation of seismic and lithologic sequences indicates that the late Miocene depositional environments can be modeled as a series of regressive coastal lithosomes beginning with a subaerial unconformity overlain by fluvial pebbly sands which is transitional to the sand/silt interbedded, deltaic calcilutite which becomes increasingly shell rich and culminates in a sandy, chenier ridge calcarenite.

The lithic-seismic correlations establish an acoustic velocity of 2.0 m/msec which is supported by published velocity estimates and a nearby sonic log, but is too low by about 2x based on velocities from Well 21 and 22 sonic logs. Limited biostratigraphic data from Well 16 indicates that the sequences are late Miocene age and their seismic and lithologic character indicate that they are controlled by eustatic sea-level fluctuations.

The NW-SE, NE-SW trending rectilinear troughs that underlie the Charlotte Harbor estuary and lagoon system occur in middle Miocene limestones and are infilled with late middle Miocene calcareous, sandy, phosphatic clays. Residual relief and/or increased erodibility of the infilling sandy clays localized the developing fluvial systems within the infilled troughs and limited cross shelf sediment transport. Some small paleo-channels were observed in seismic profiles seaward of the

barriers. These channels are smaller than those landward of the barriers and may be due to rivers flowing east or southeasterly towards the karstic depression.

Three Quaternary lithologic sequences were differentiated from the vibracores and interpreted on the basis of modern depositional environments. These sequences are: 1) a fining upward, fluvial to upper estuarine point bar unit, 2) a transitional estuarine unit of highly bioturbated, relatively unfossiliferous, muddy to slightly muddy fine sand, and 3) shelly to muddy sand, fining upward tidal delta/tidal channel sequence. The point bar FUS is limited to upper estuarine cores except where it underlies tidal channel deposits. The estuarine sequence occurs primarily within the upper and lower Harbor and also underlies the tidal channel sequence in several lagoonal cores.

The tidal channel FUS occurs throughout the lower Harbor and is the dominant infilling sequence within the lagoons. The thickness and frequency of this unit changes from the Harbor to lagoonal sites and landward to seaward across the lagoons. Close to the islands the unit is thin (35 to 100 cm) and repetitive due to frequent inlet/flood tidal delta migration and possibly overwash events.

Dating of the sequences using amino acid racemization and determination of the Pleistocene/Holocene unconformity by indurated, leached shell beds and Fe/humic stained paleosols indicates that most of the interpreted sequences are Holocene but five Pleistocene depositional episodes are also present.

Pleistocene sequences are present (as each of the interpreted units), however, they are usually incomplete and difficult to correlate. AA dates of articulated shells show that most of the deposition in this system can be attributed to sea-level high stands at isotopic stages 7 and 1 with some units from stages 9, 5a and 5e.

The hydrogeologic development of the Charlotte Harbor area is affected by episodes of both low sea-levels as periods of major karstification and by transgressing sea-levels which have generated discrete distributions of permeable and impermeable deposits. Evidence of lateral karst progression as sinks and springs around the margins of the troughs suggest that vertical exchange is the rule rather than the exception. It is also apparent from the widespread variability of depositional and karstic processes that correlation of aquifer systems on a regional scale without continuous seismic profiles or very closely spaced borehole data will be unreliable.

In conclusion, the transition from a carbonate to a siliciclastic depositional system which began in the late Neogene in southwest Florida is ongoing. A regional pattern of karstic dissolution and collapse has controlled the transition by localizing deposition of impermeable prodelta clays and the developing paleo-fluvial systems. The introduction of quartz sand into the carbonate terrain was affected by the localized fluvial systems and, the location of modern and relict barrier systems has, in turn, been controlled by the residual relief and increased sediment supply derived from the paleo-fluvial

systems. Processes of carbonate accumulation, lithification and dissolution are ongoing and have resulted in the formation and potential preservation of modern and relict barrier platforms.

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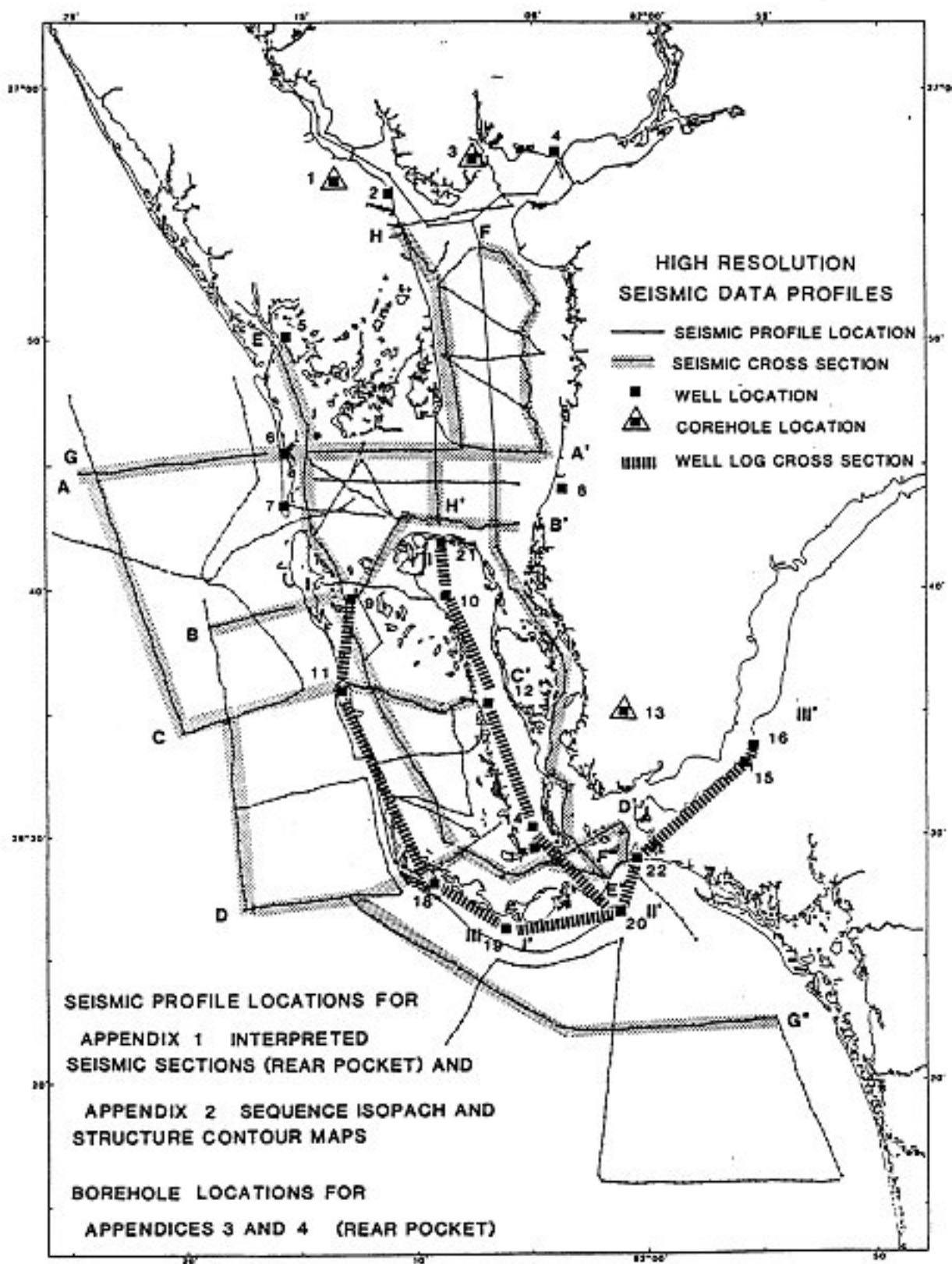
APPENDICES

APPENDIX 1

SEISMIC PROFILE LOCATIONS

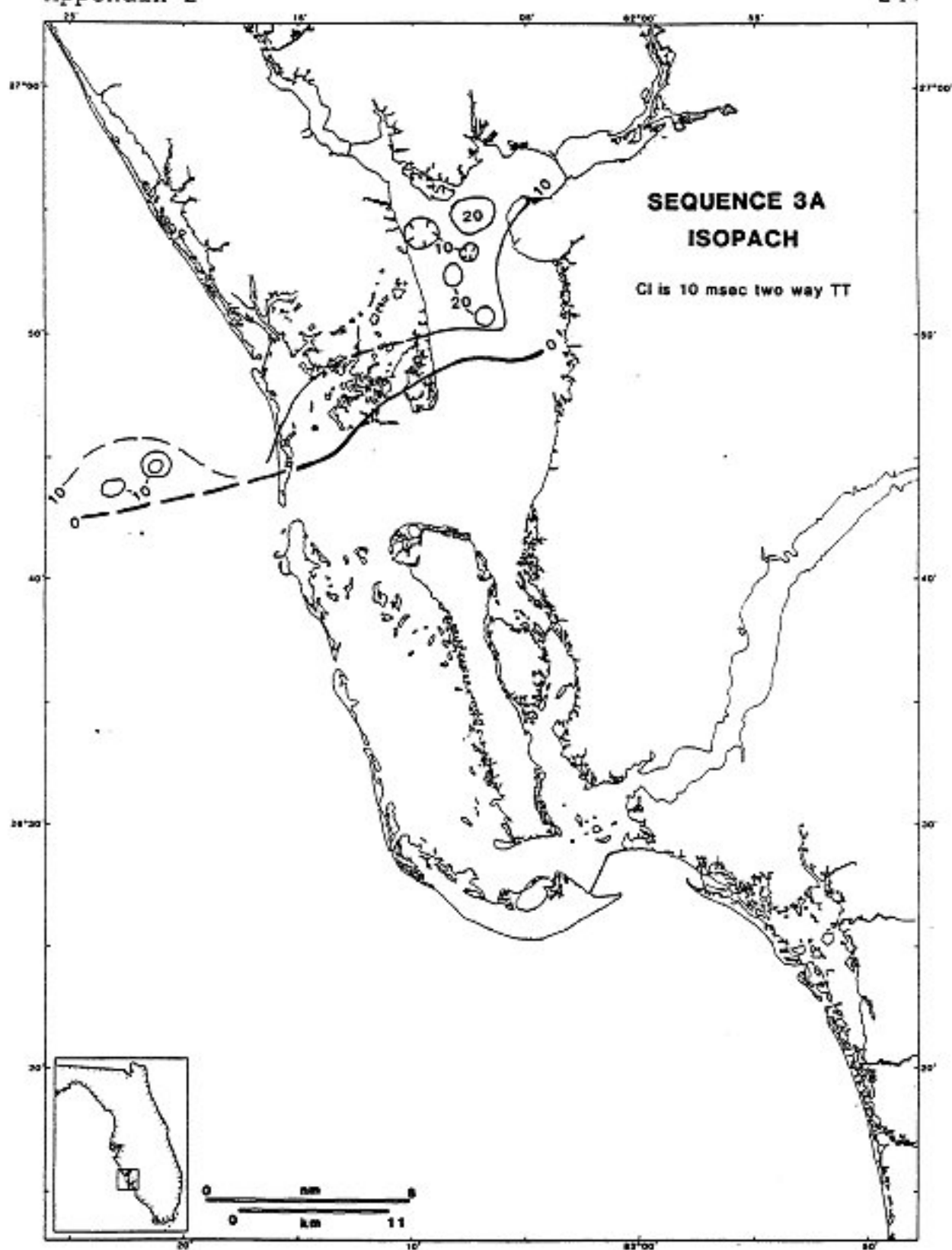
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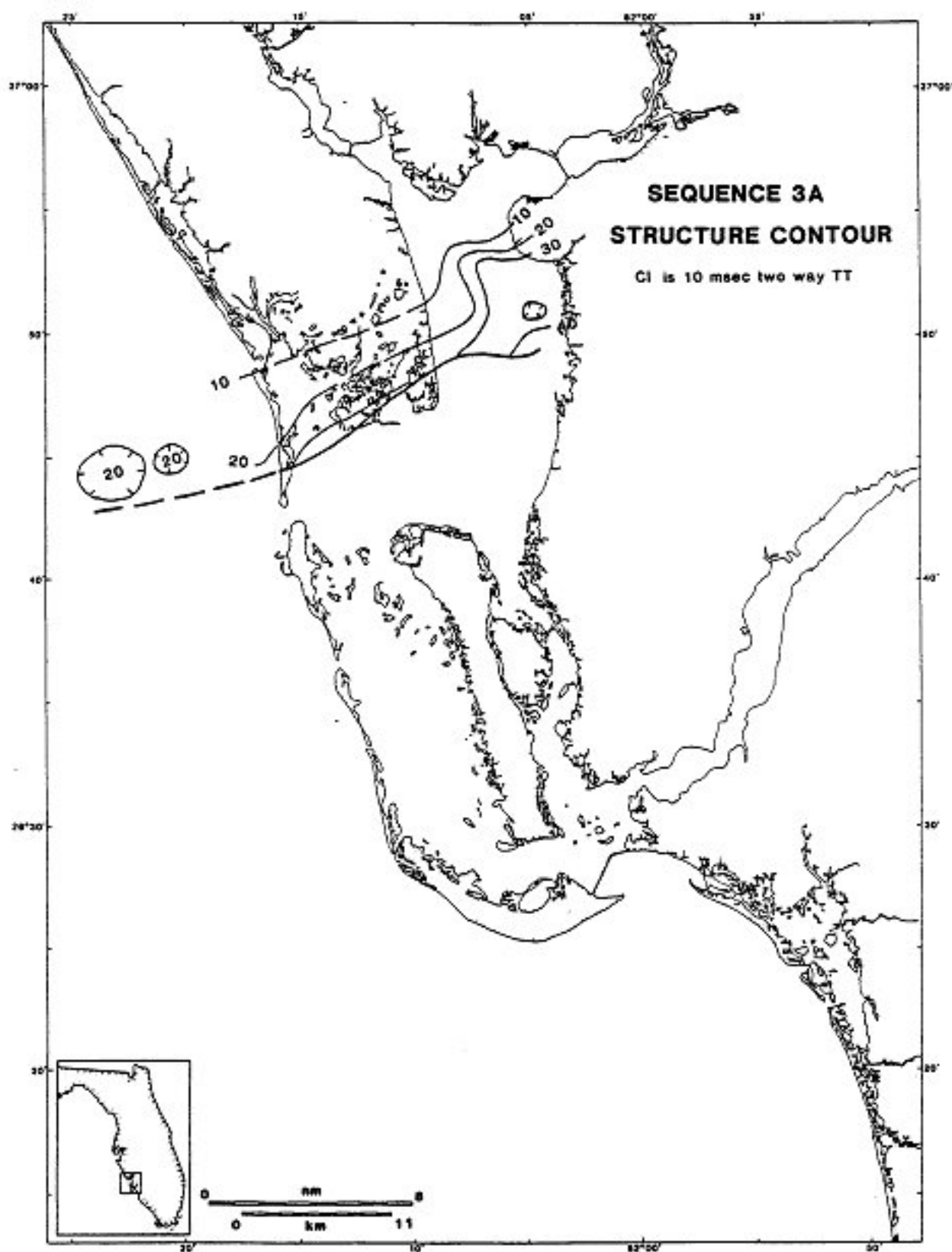
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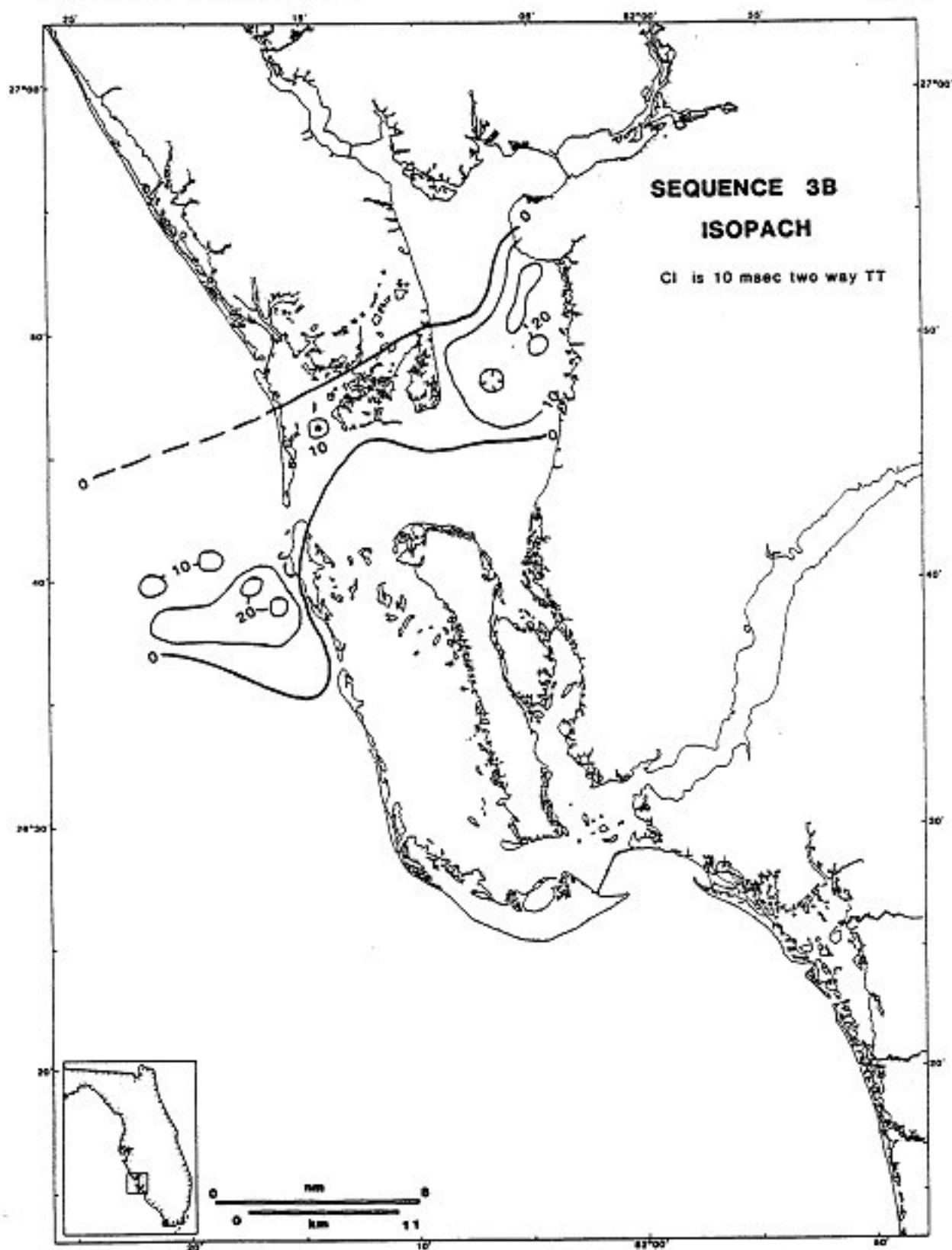


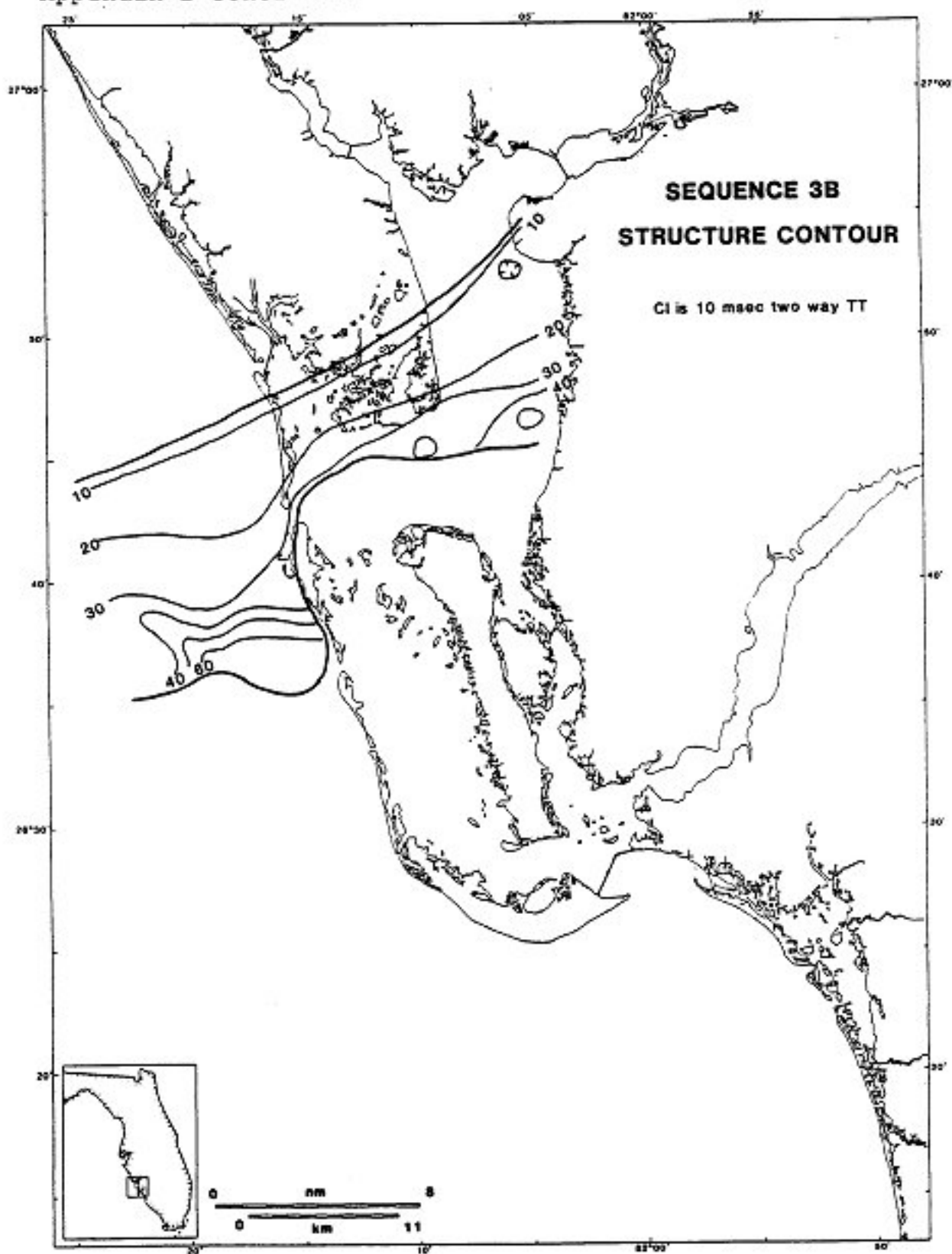
APPENDIX 2

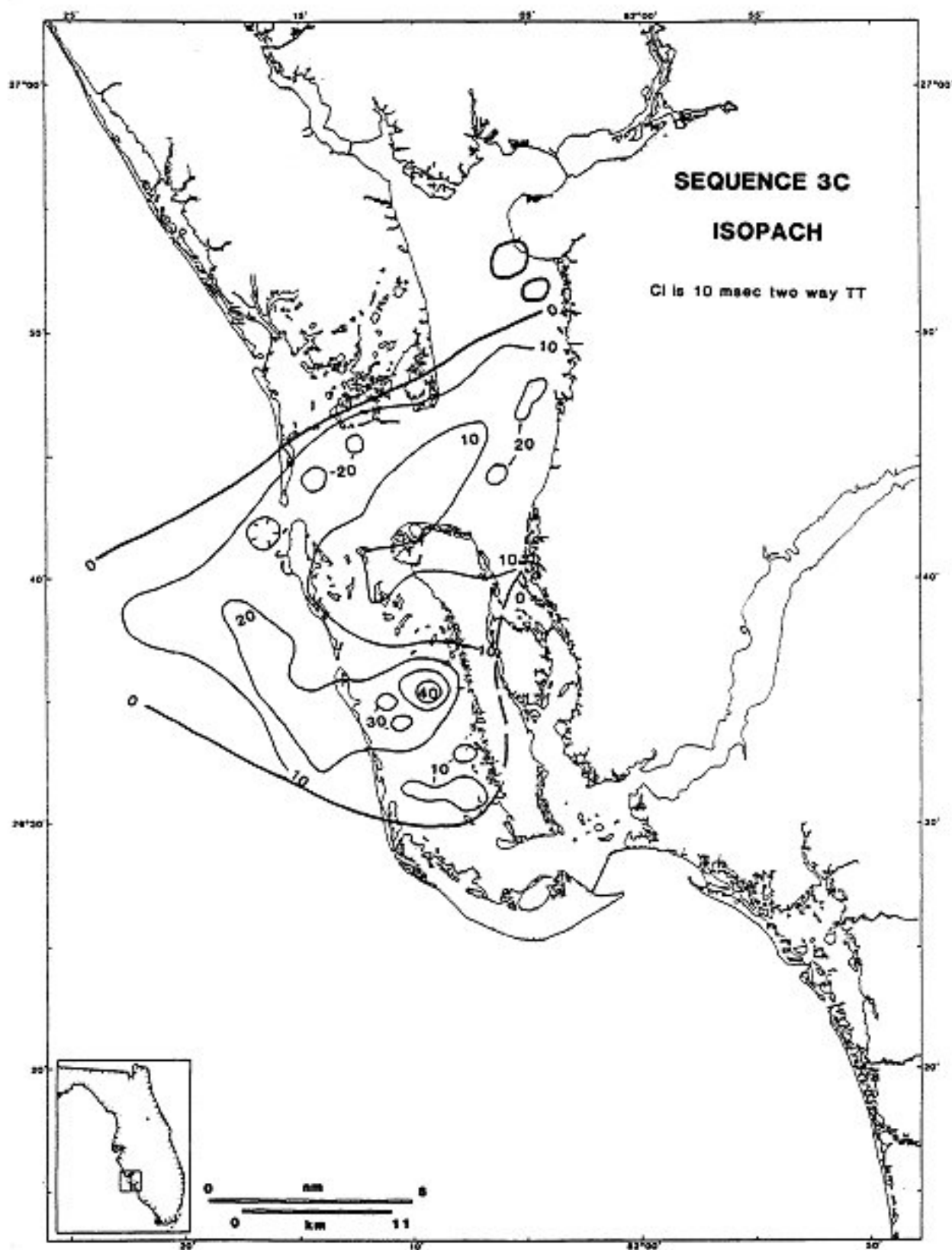
SEISMIC SEQUENCE ISOPACH AND STRUCTURE CONTOUR MAPS

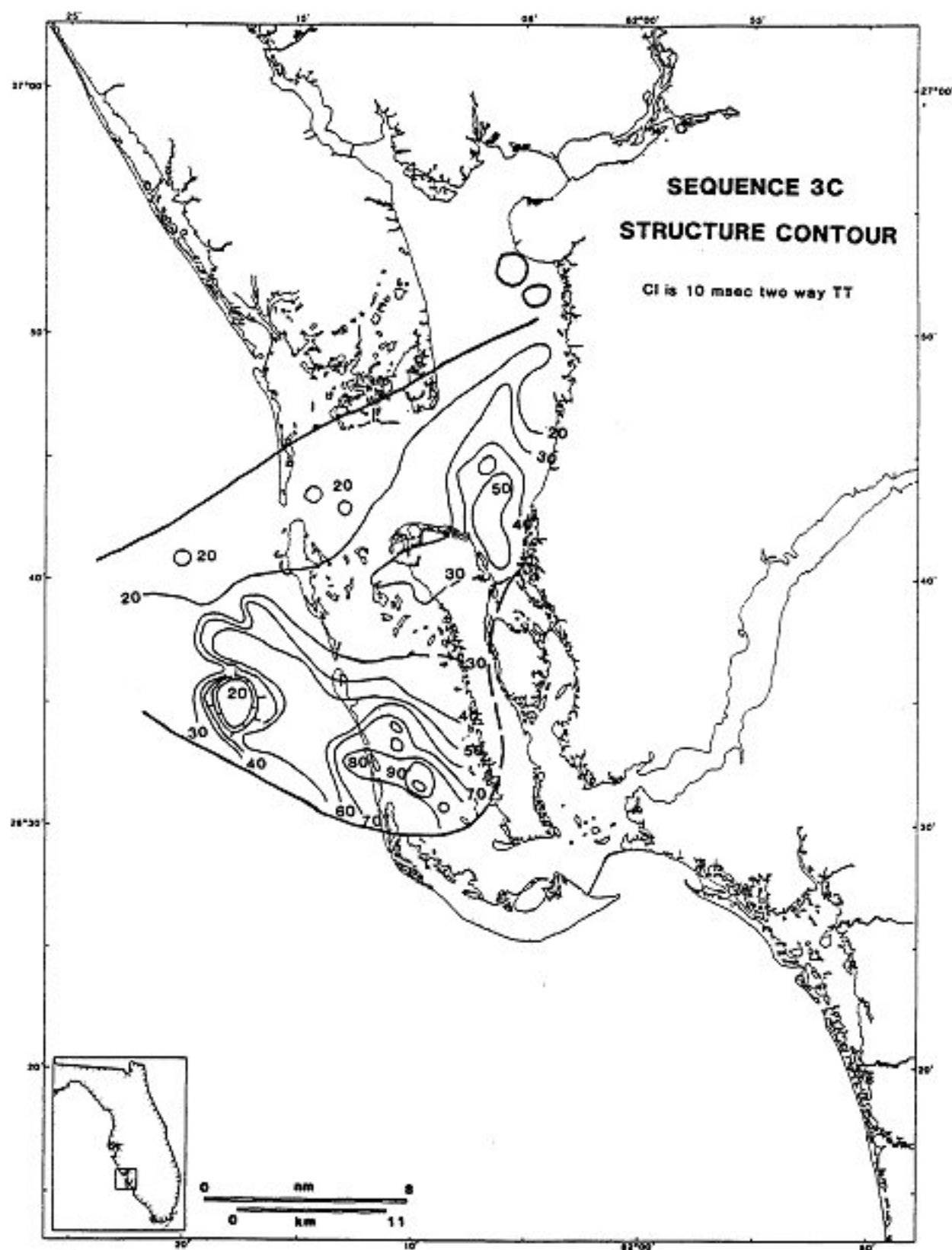


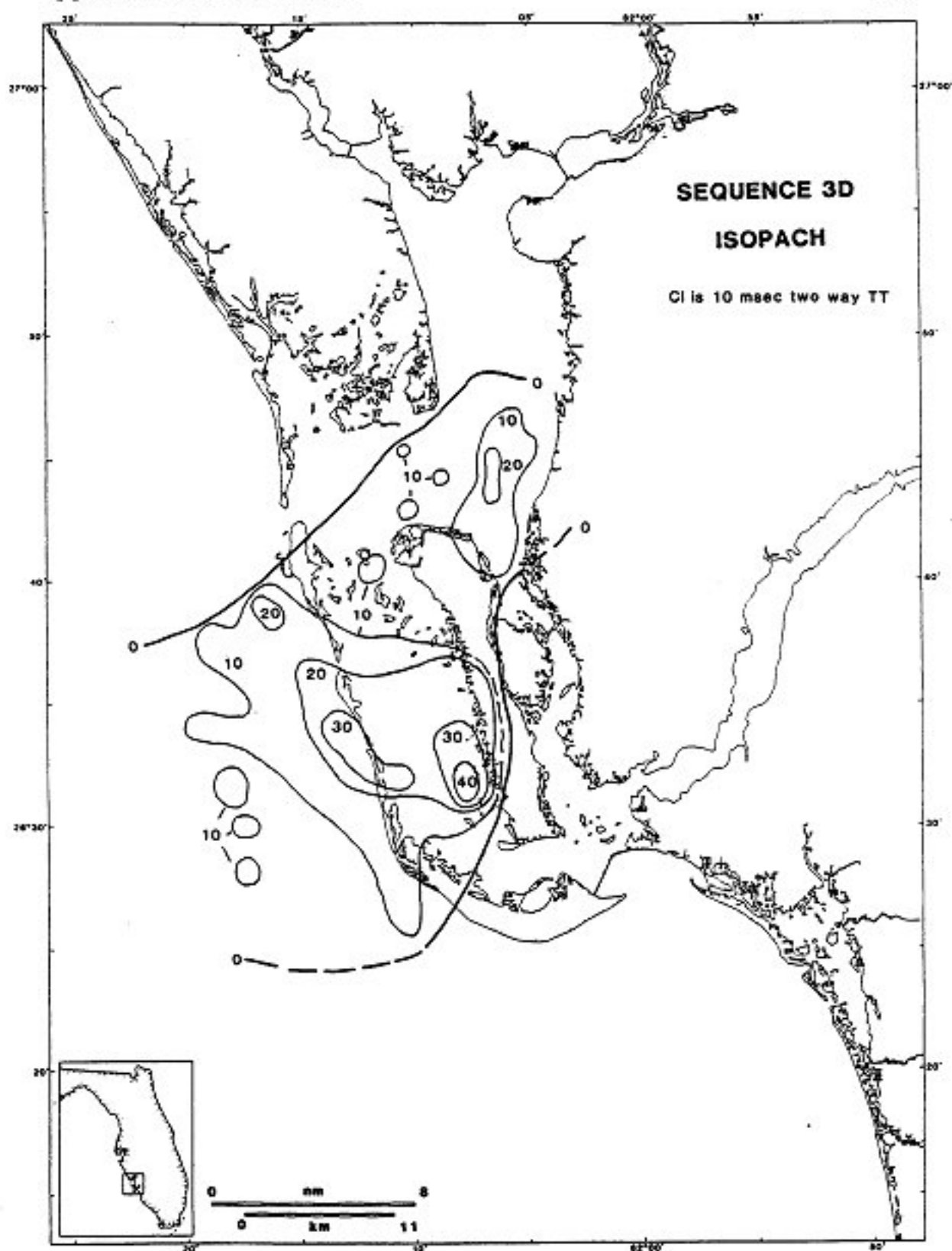


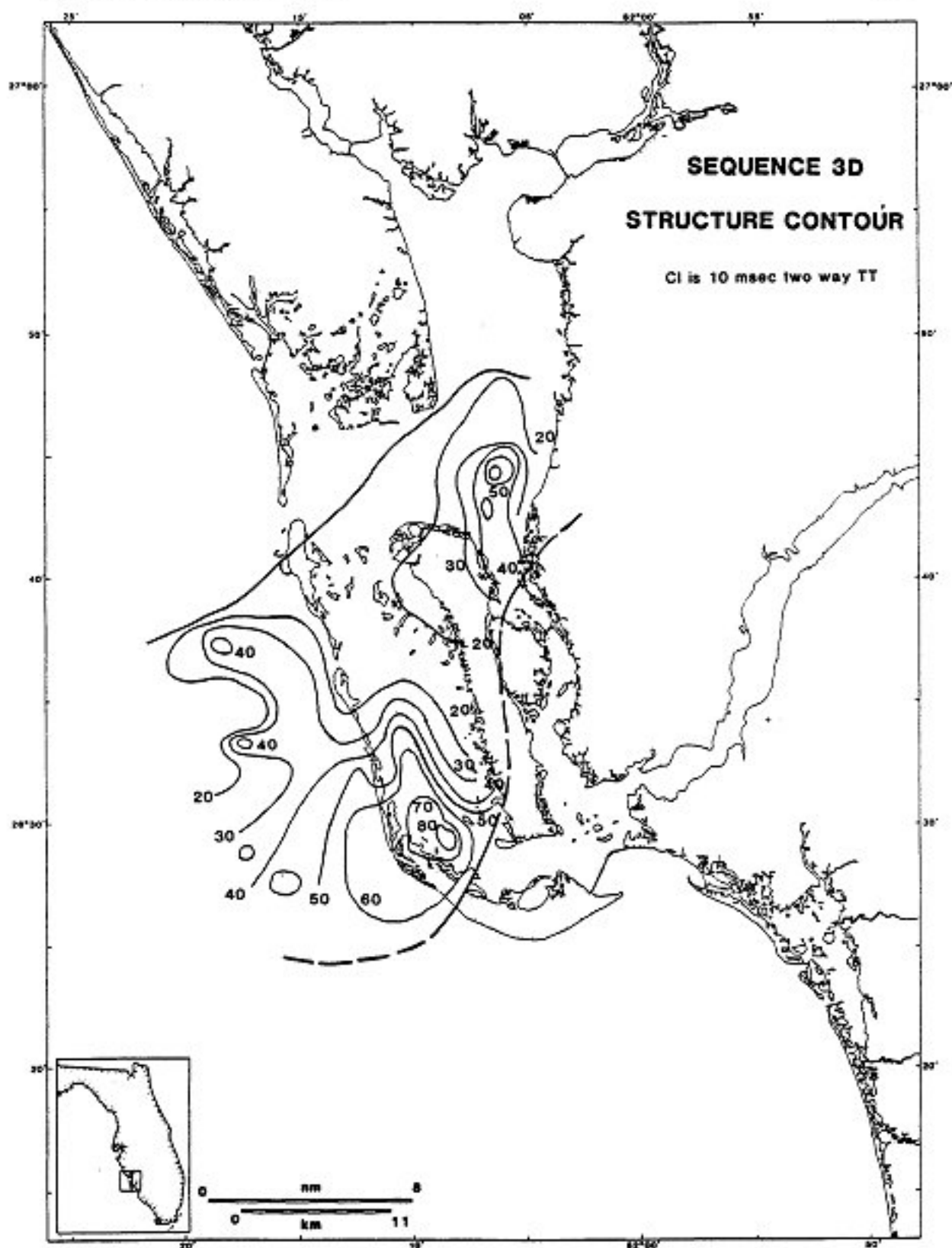


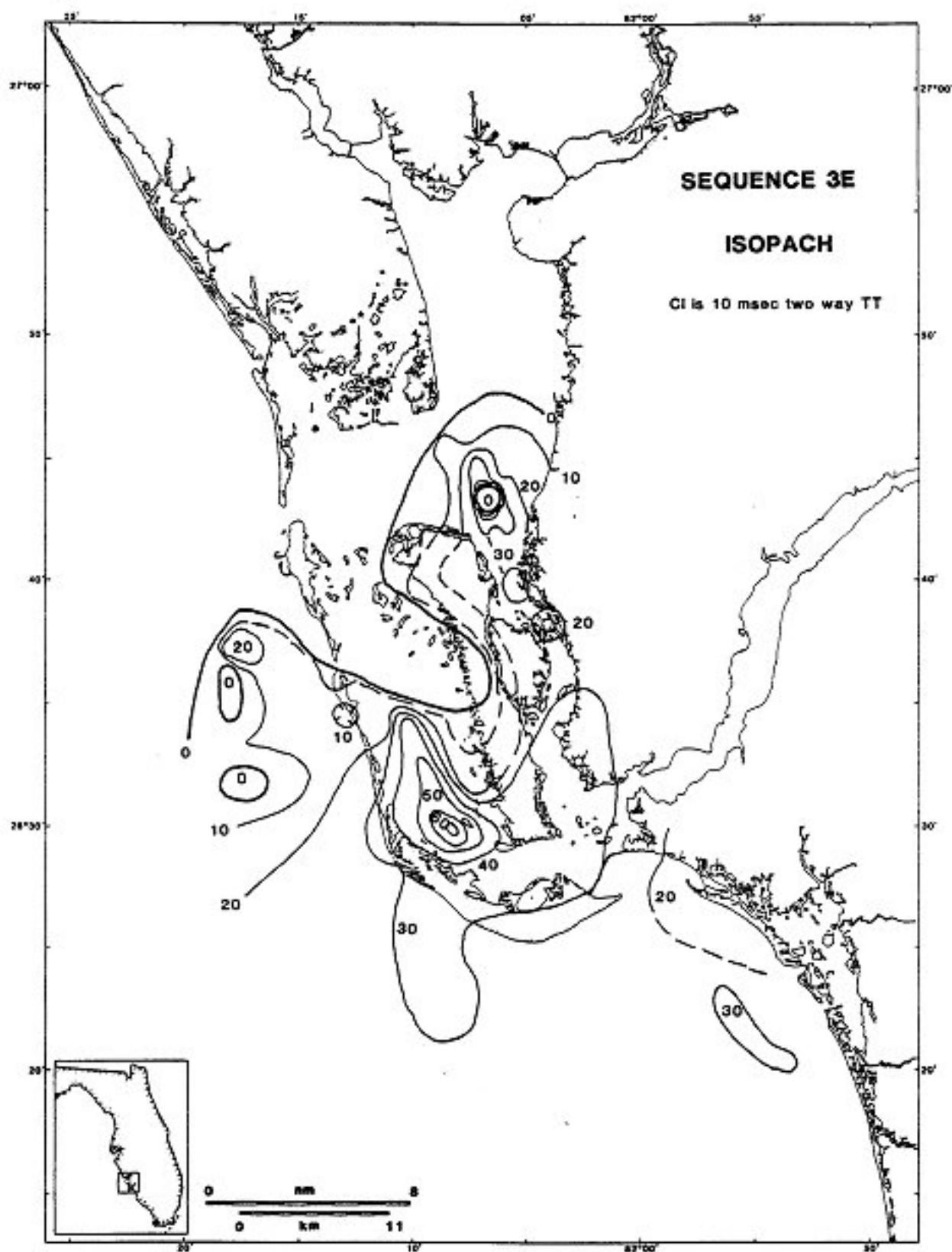


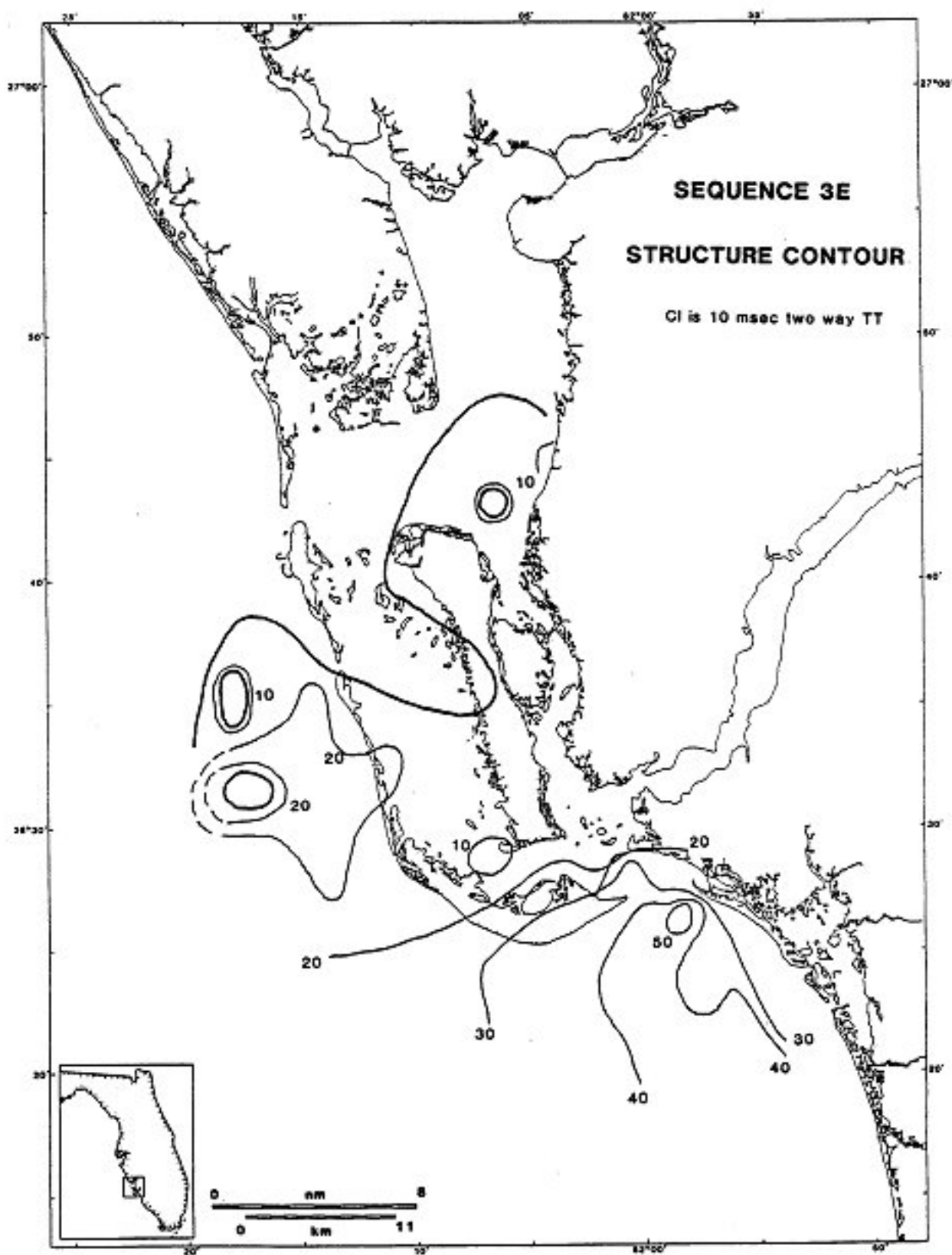


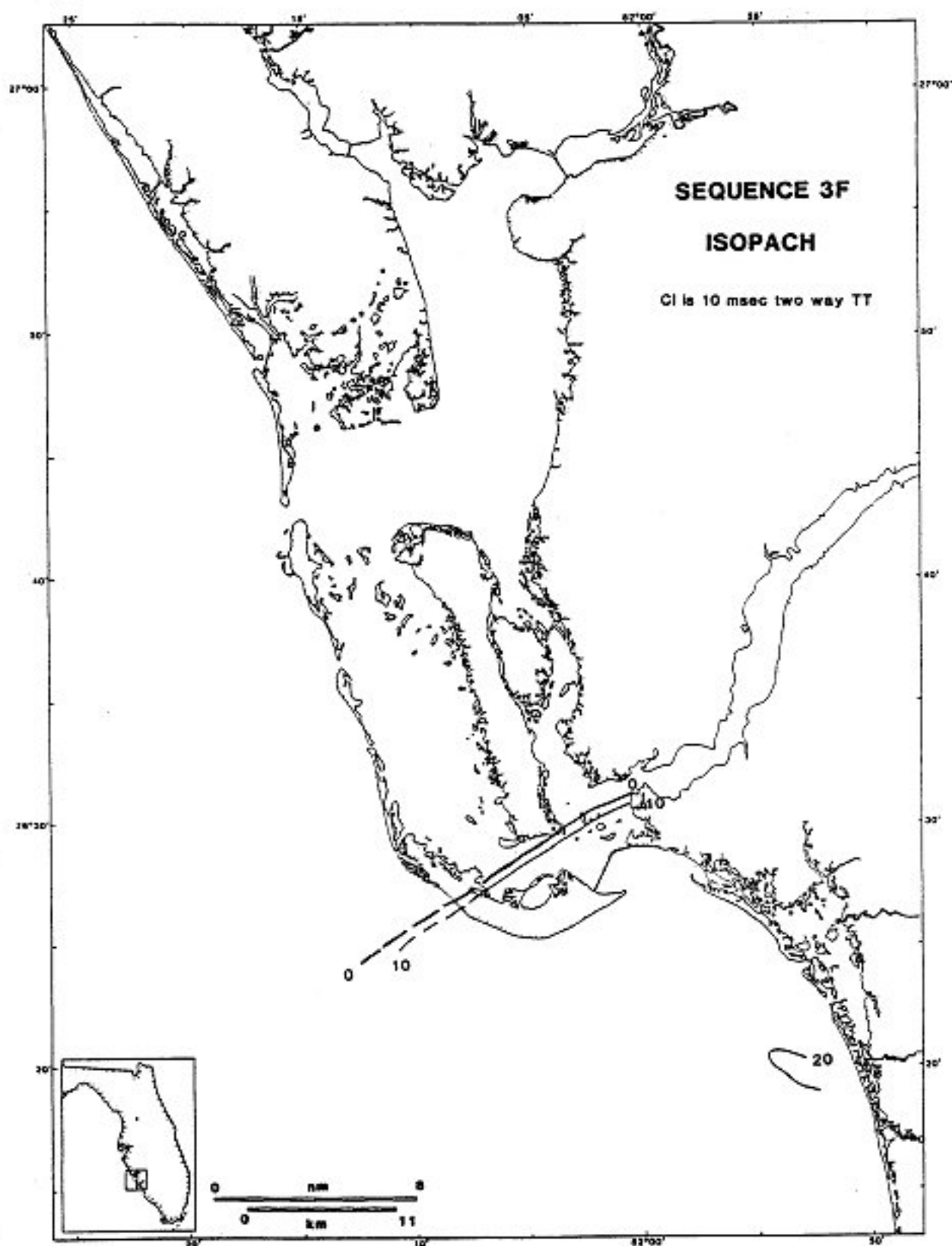


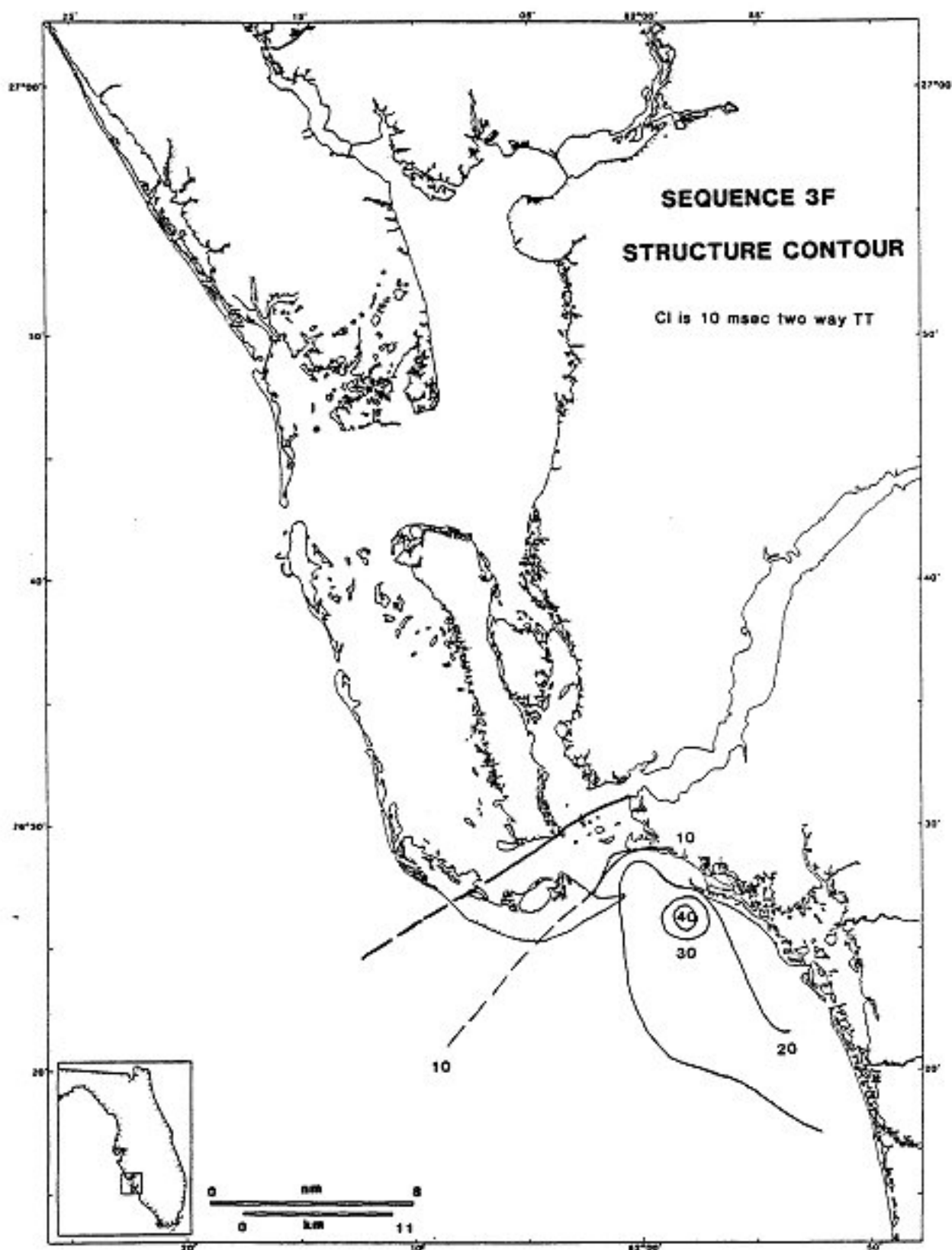








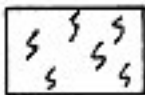
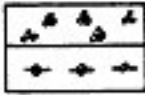
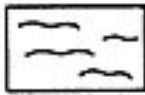
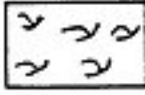
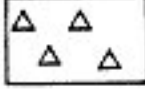
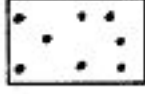
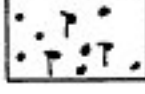
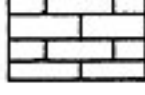
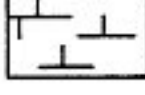
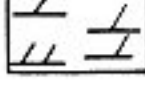




APPENDIX 3

BOREHOLE LITHOLOGIC AND GEOPHYSICAL LOGS

Lithologic Key

Clay		Microfossils Foraminifera	
Silt		Pelecypod Fossils	
Sand		Chert	
Pebbles		Phosphate Phosphatic Pebbles	
Limestone		Calcite/ Aragonite Cement	
Dolomite		Dolomite Cement	

#1 (TR3-1, W-15332)

meters feet

0 0
5 20
10 40
15 60
20 80
25 100
30 120
35 140
40 160
45 180
50 200
55 220
60 240
65 260
70 280
75 300
80 320
85 340
90 360
95 380
100
105
110



OFF WHITE TO YELLOWISH BRN.
W/ DARK BRN. AND ORANGE STAINS
LT. BLUISH GRAY/DARK GRAY

OFF WH. TO BLUISH TAN,
GRAY TO BL. GRAY DK GRAY
LT. TO BL. GRAY

LT. TO DARK GRAY GRN. TO BLUISH
GRAY AT BOTTOM

BL. TO LT. GRN. CLAY
LT. GRN. TO BLUISH GRAY

CREAM TO OFF WH.
GREENISH GRAY TO BLACK
CREAM TO WH. LT. TO GRN. GRAY
GREENISH GRAY CREAM TO OFF WH.
CREAM TO LT. GRAY

GRN. TO GRN. GRAY

LT. YELL. GRN.

GRN. GRAY TO LT. YEL. GRN.
CREAM TO OFF WHITE
TAN TO TANNISH GRAY

OFF WH. TO LT. GRAY

LT. GRAY

CREAM - OFF WH.

LT. GRAY TO DK. DARK SOFT
LT. GRAY

OFF WH. TO LT. TAN GRAY

LT. GRAY

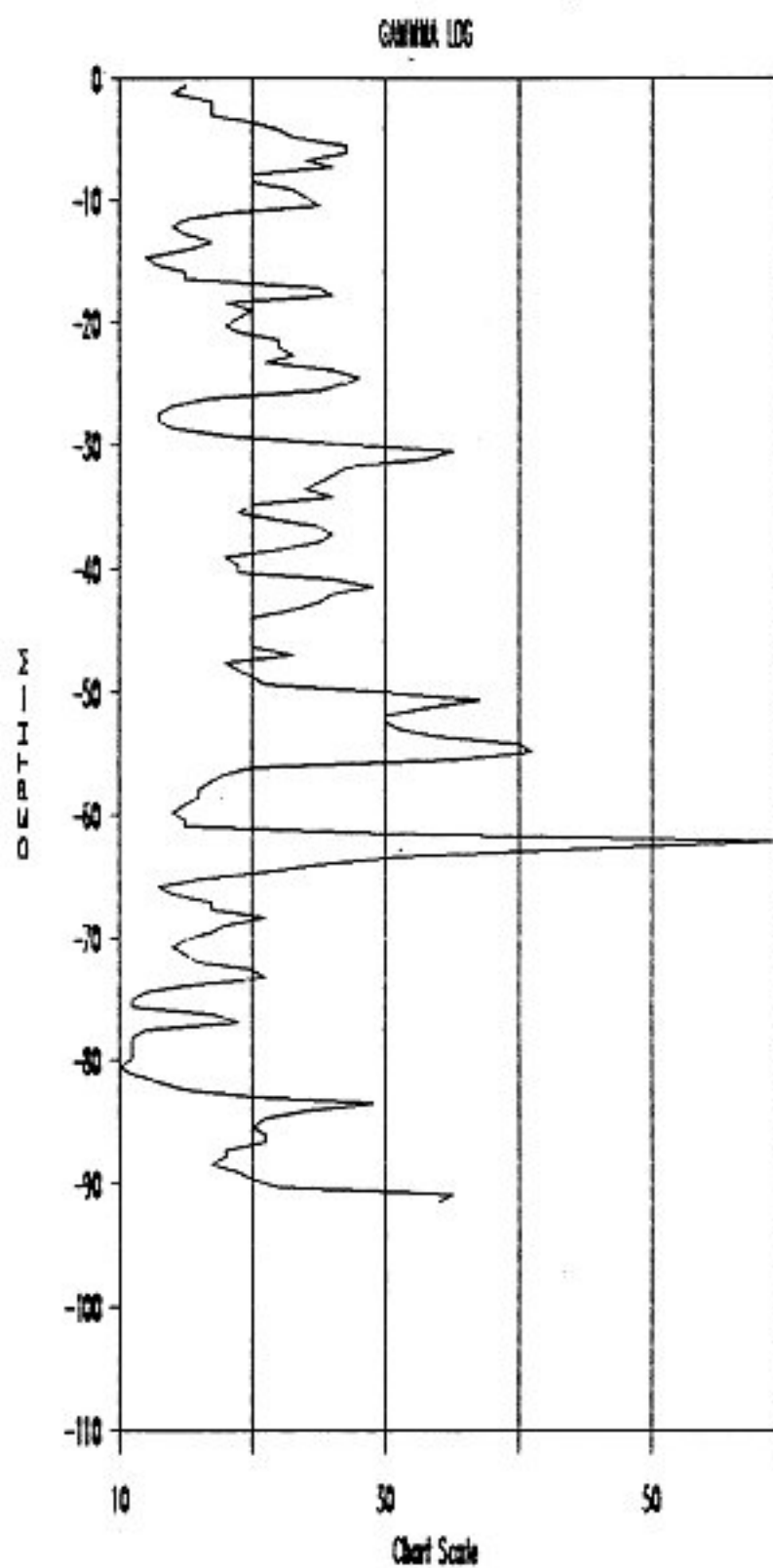
DK. GRAY

LT. GRAY TO TAN GRAY

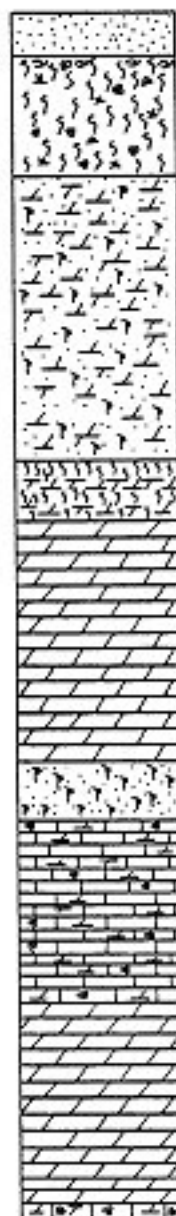
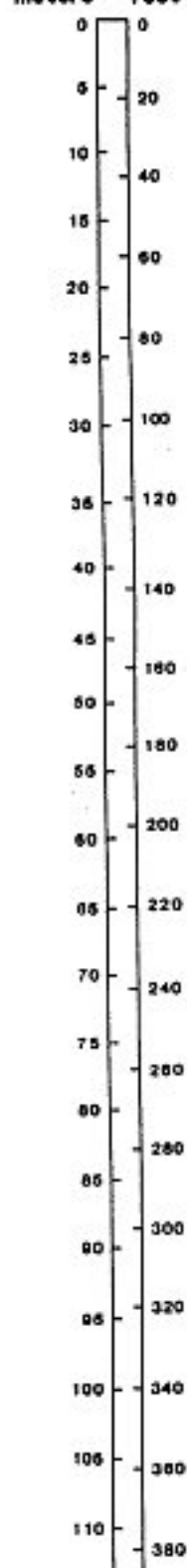
CREAM - LT. TAN. GRAY TAN

OFF WH. TO LT. GRAY

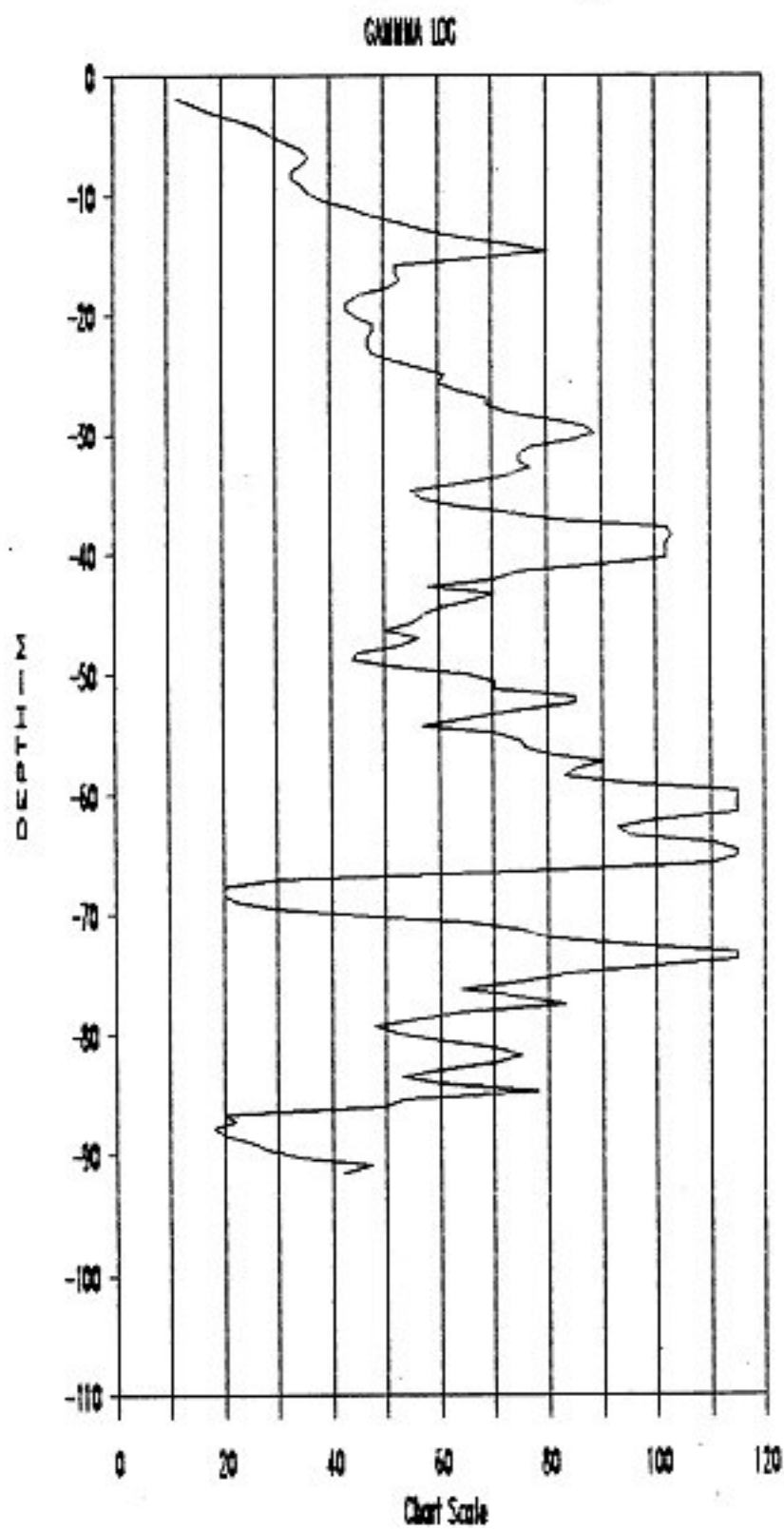
WELL 01 (EI JOBEAN)



meters feet



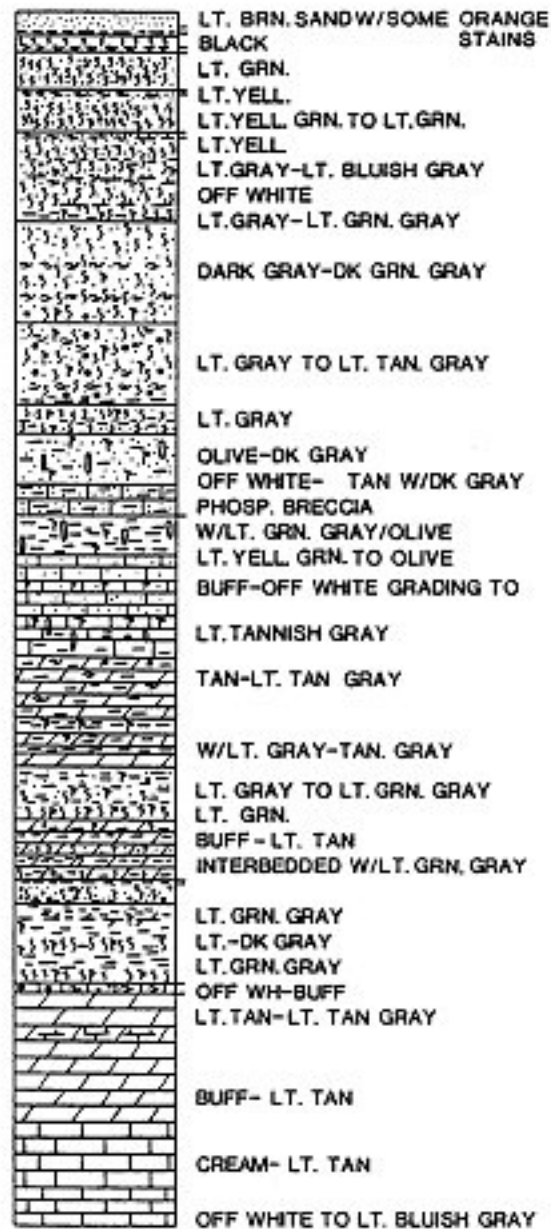
WELL 02 (Cattle Dock Pt.)



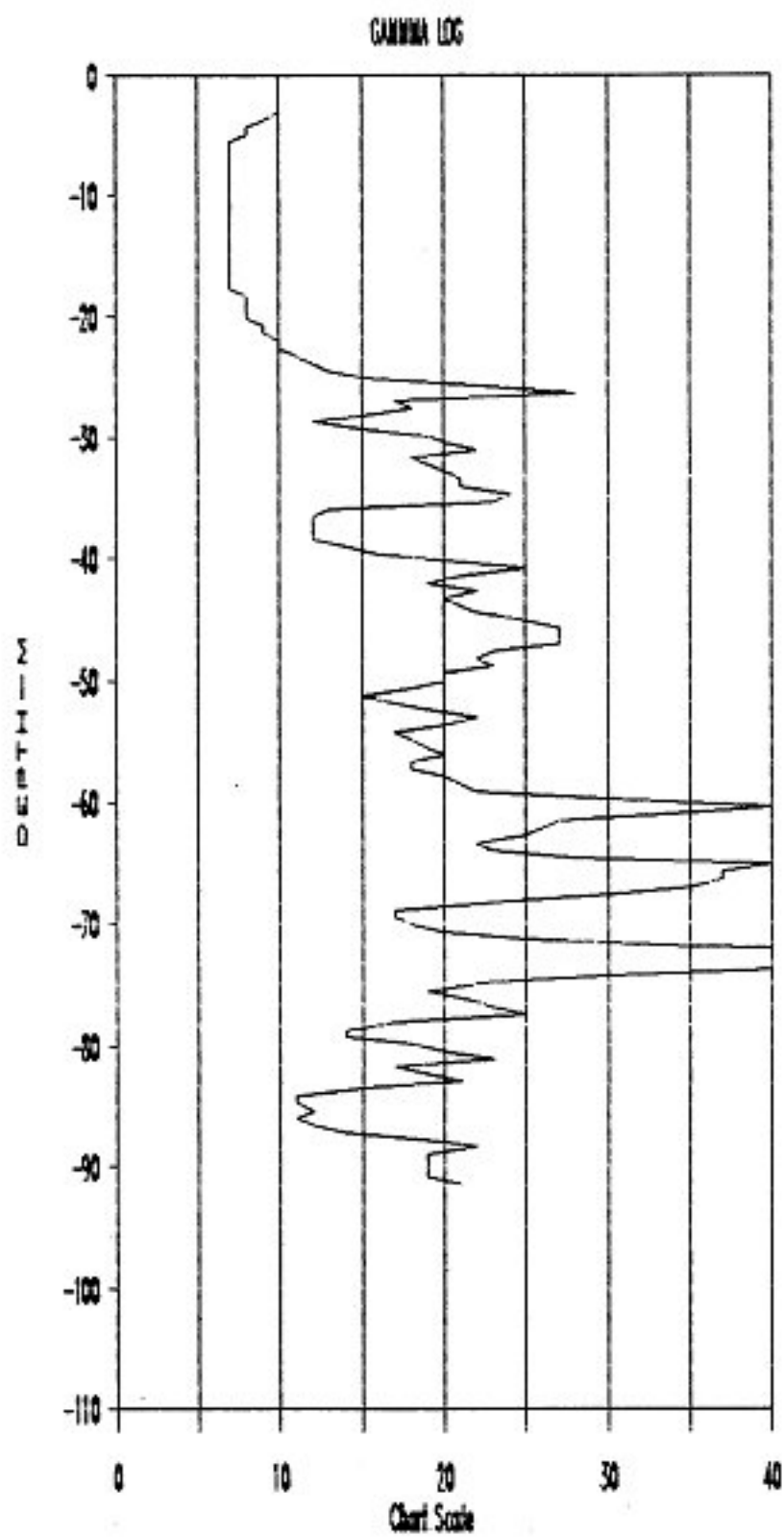
#3 ROMP TR 2-1

HOG ISLAND

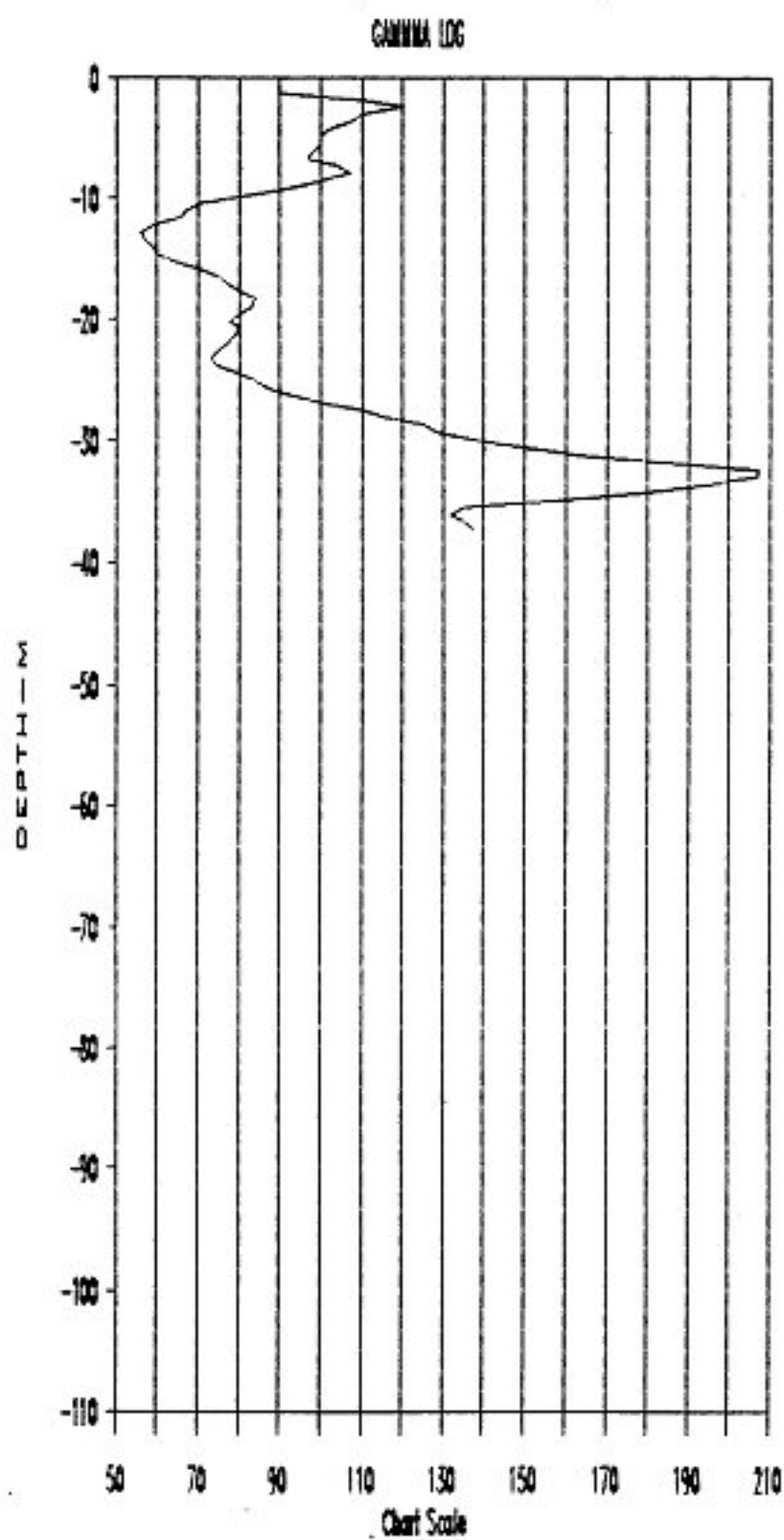
meters feet



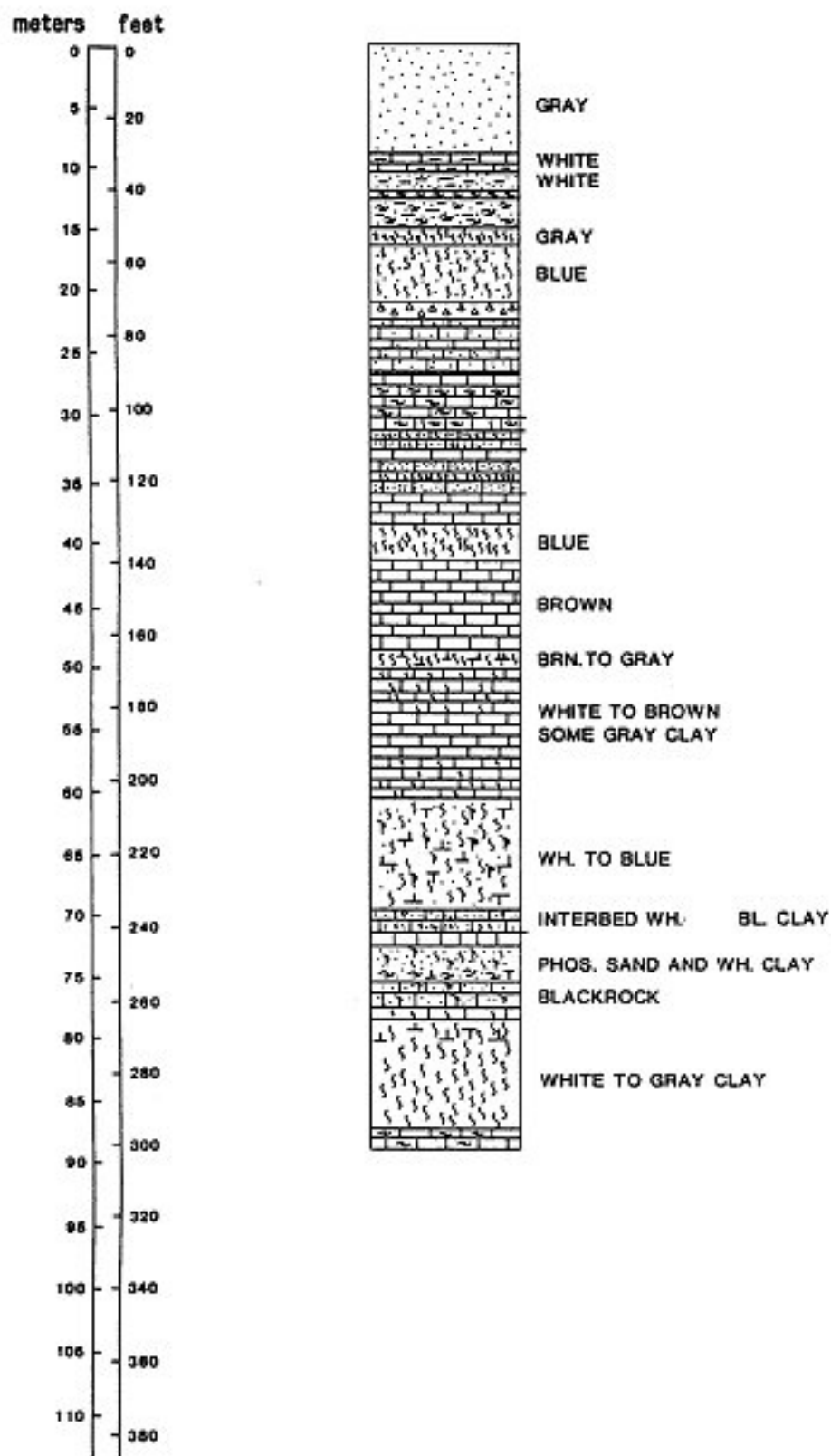
WELL 03 (Hog Island)



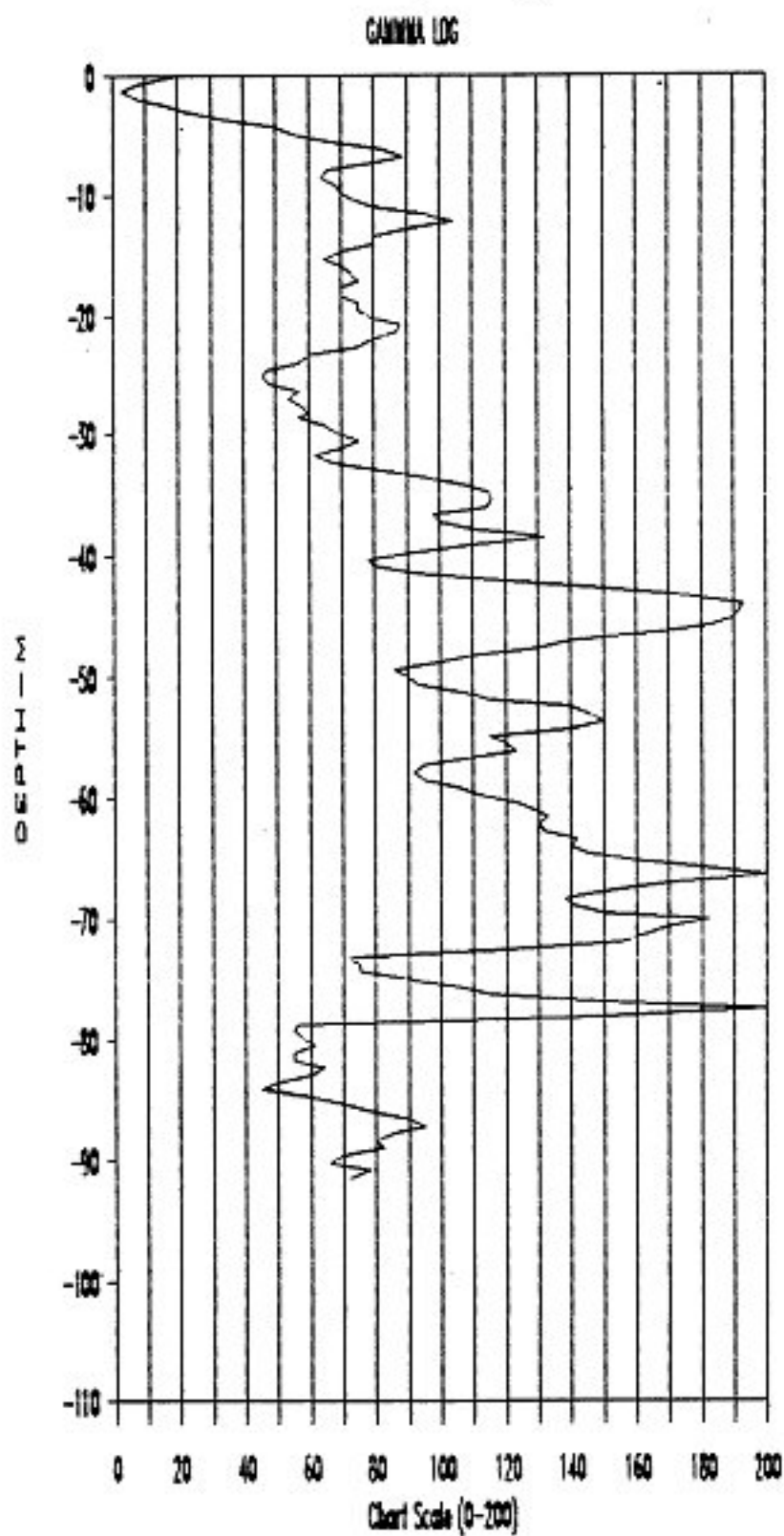
WELL 04 (PORT CHARLOTTE)



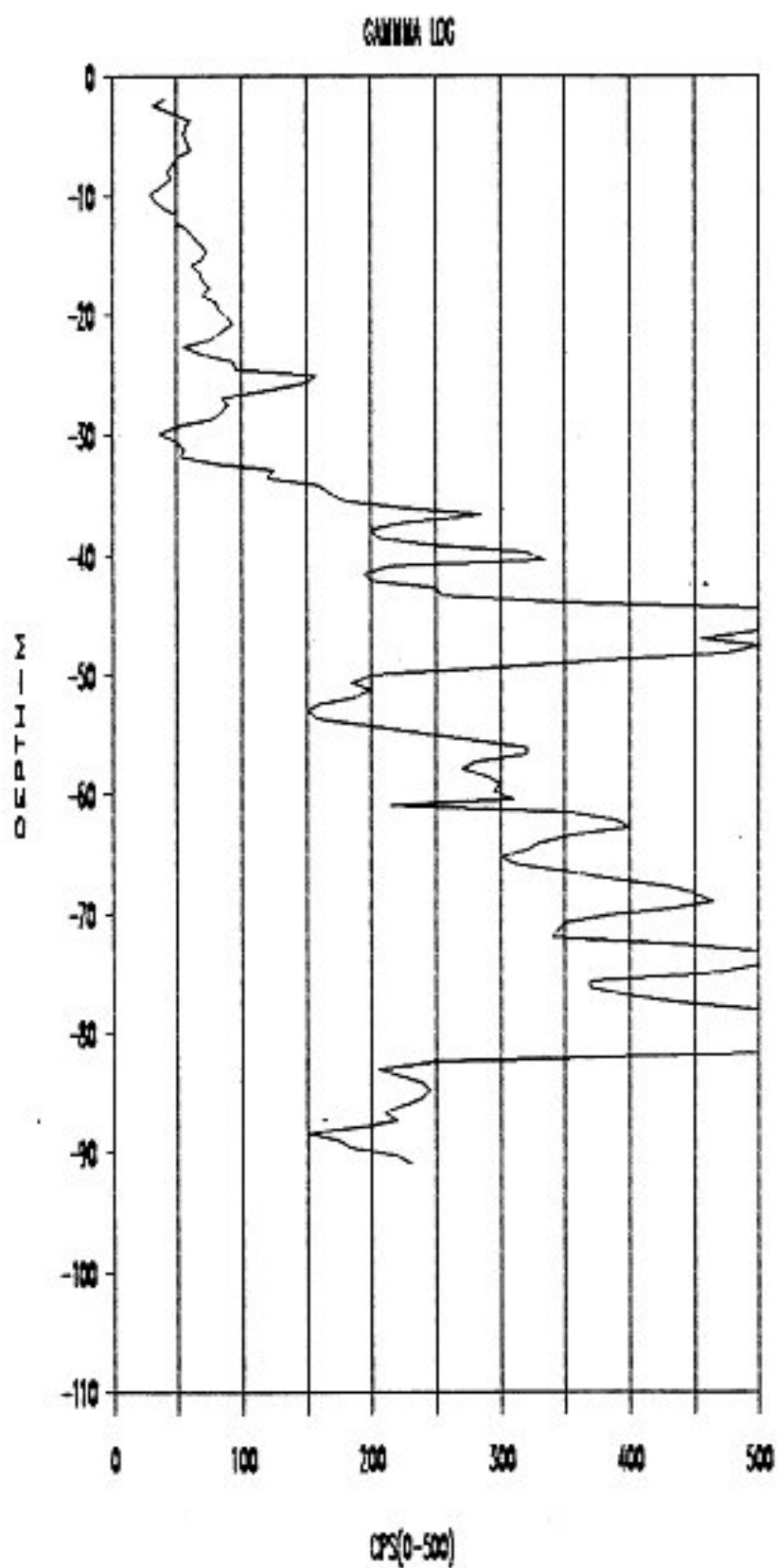
#5 (PLACIDA #8)



WELL 05 (Placido)



WELL 06 (GASPARILLA ISLAND)

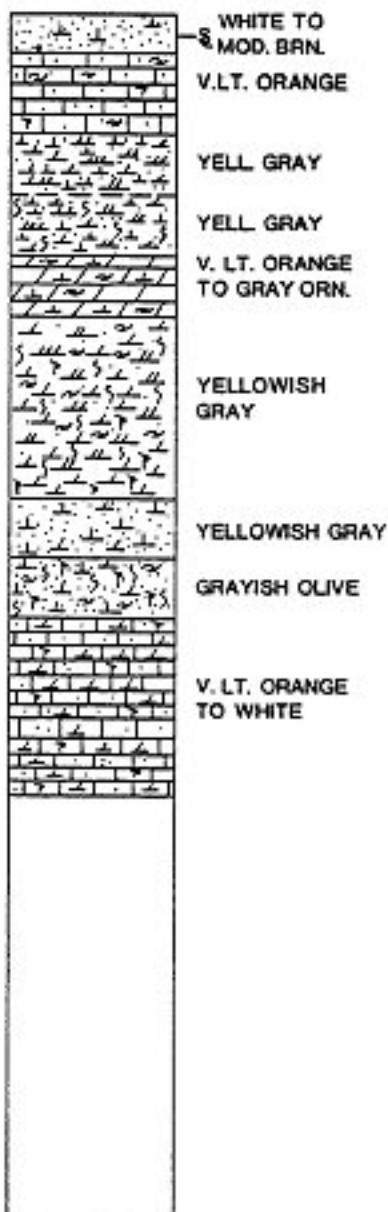


meters feet

0 0
 5 20
 10 40
 15 60
 20 80
 25 80
 30 100
 35 120
 40 140
 45 160
 50 180
 55 180
 60 200
 65 220
 70 240
 75 260
 80 280
 85 280
 90 300
 95 320
 100 340
 105 360
 110 380

W-LE005 #8

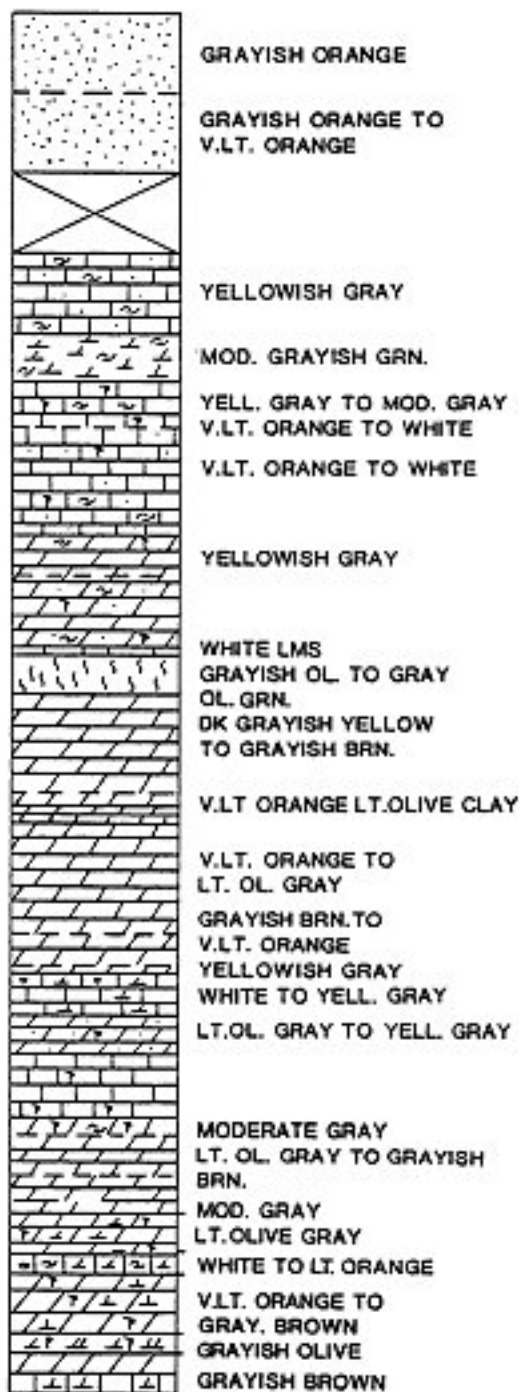
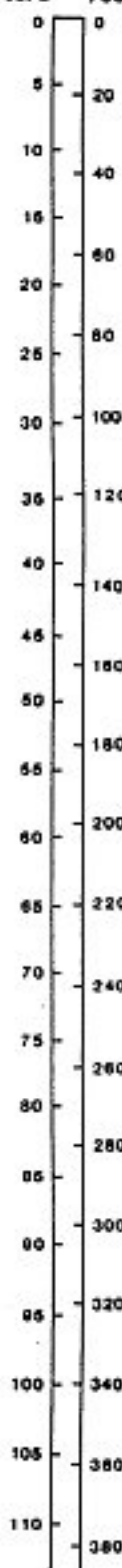
PEN CREEK



#9 W-LE015

USEPPA ISLAND

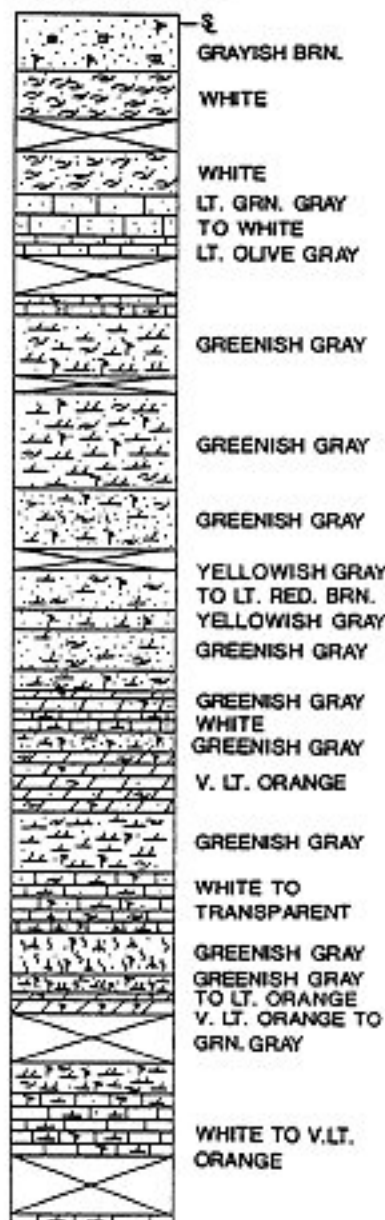
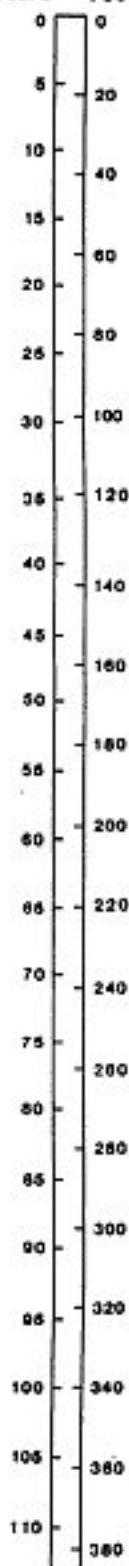
meters feet



W-LE014 10

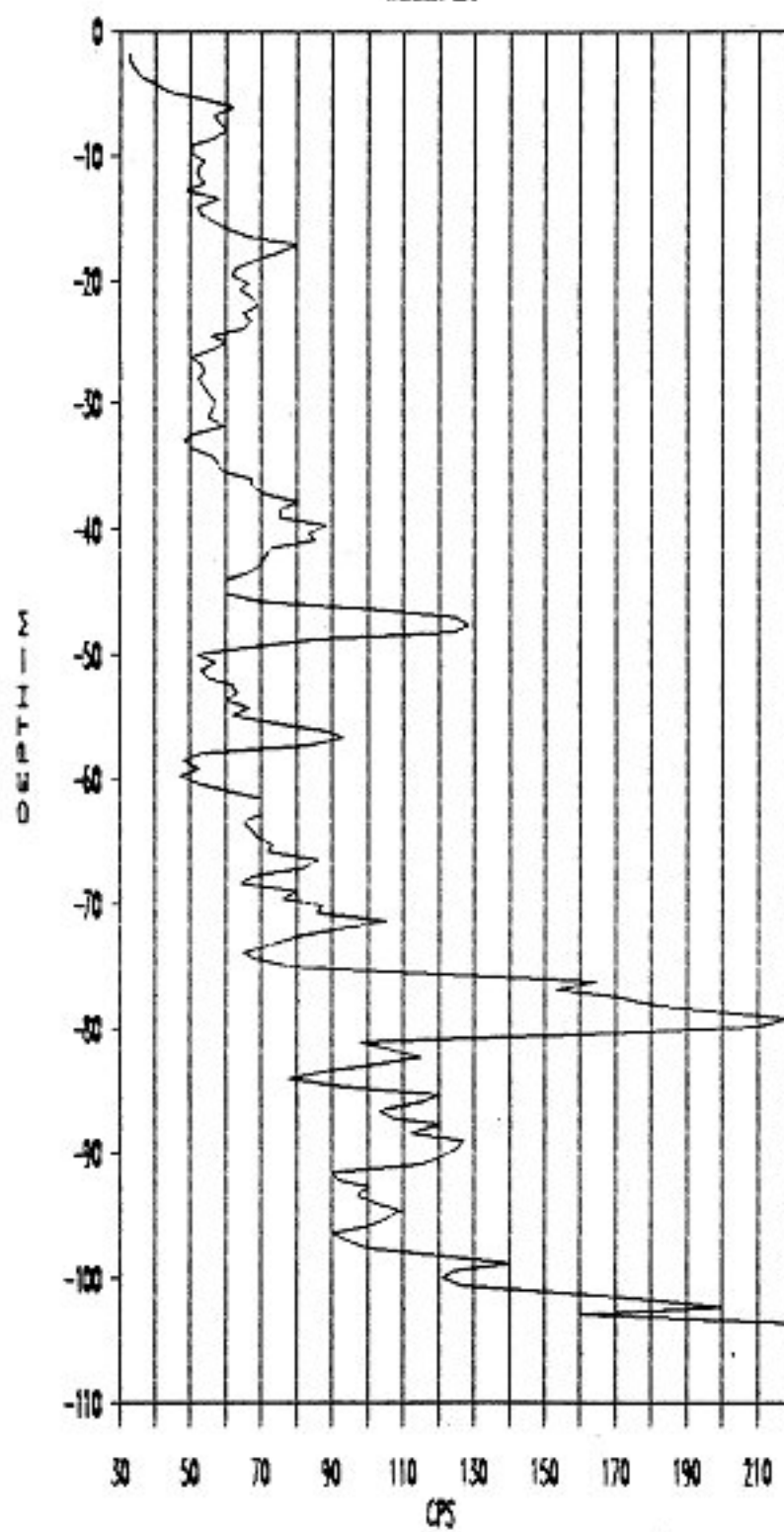
N. PINE ISLAND

meters feet

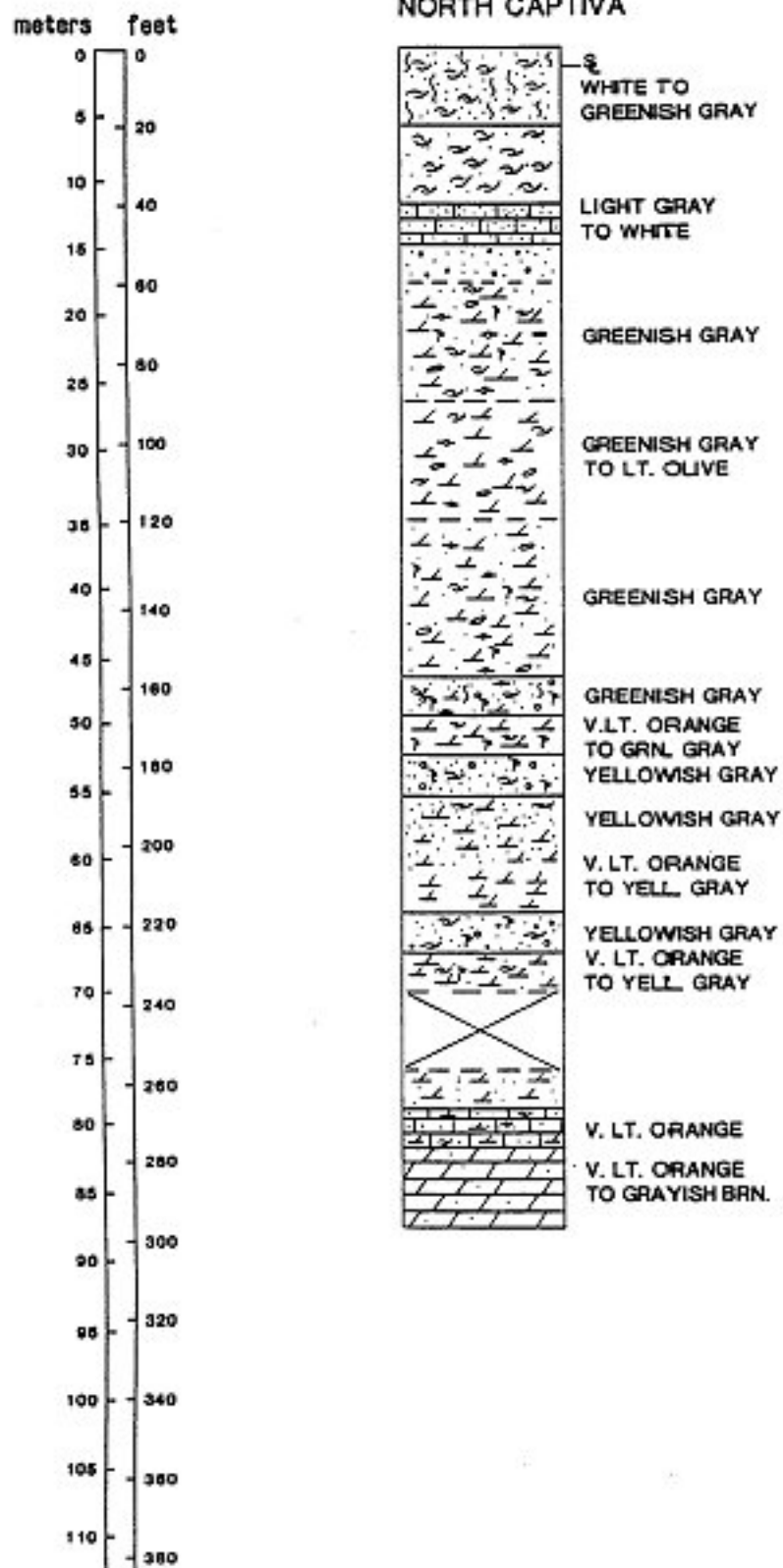


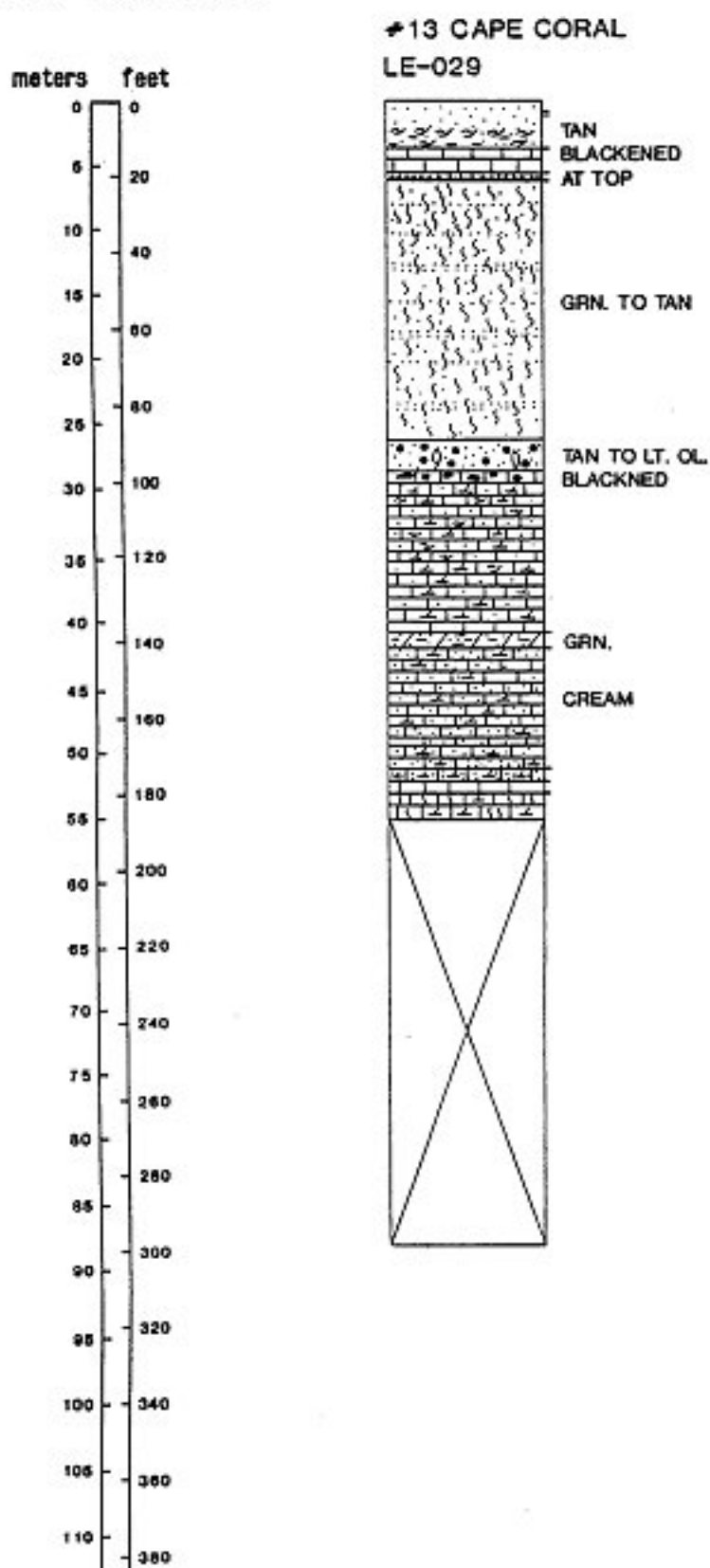
WELL 10

GAMMA LOG



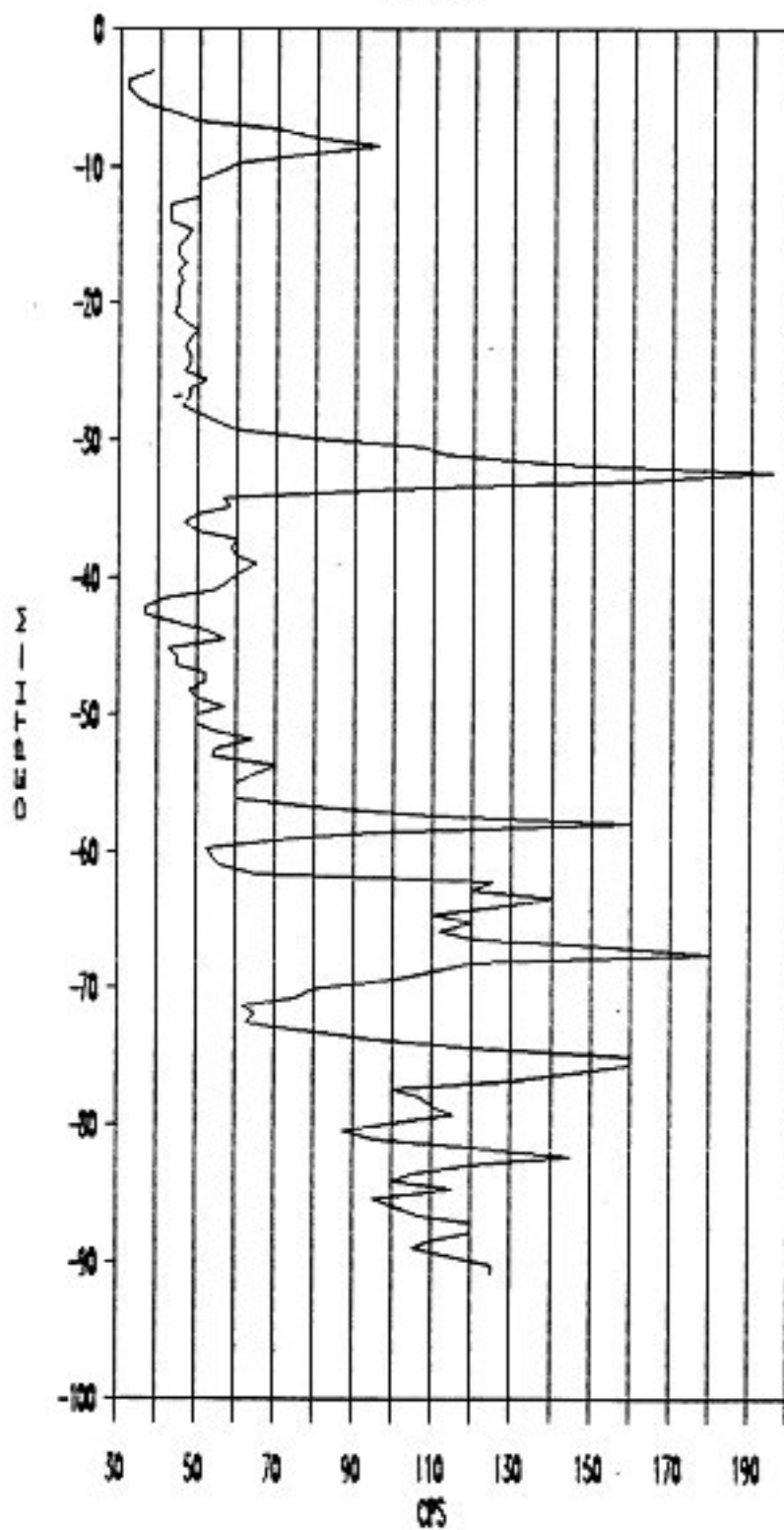
#11 W-LE021
NORTH CAPTIVA





WELL 13 (CAPE CORAL)

GAMMA LOG

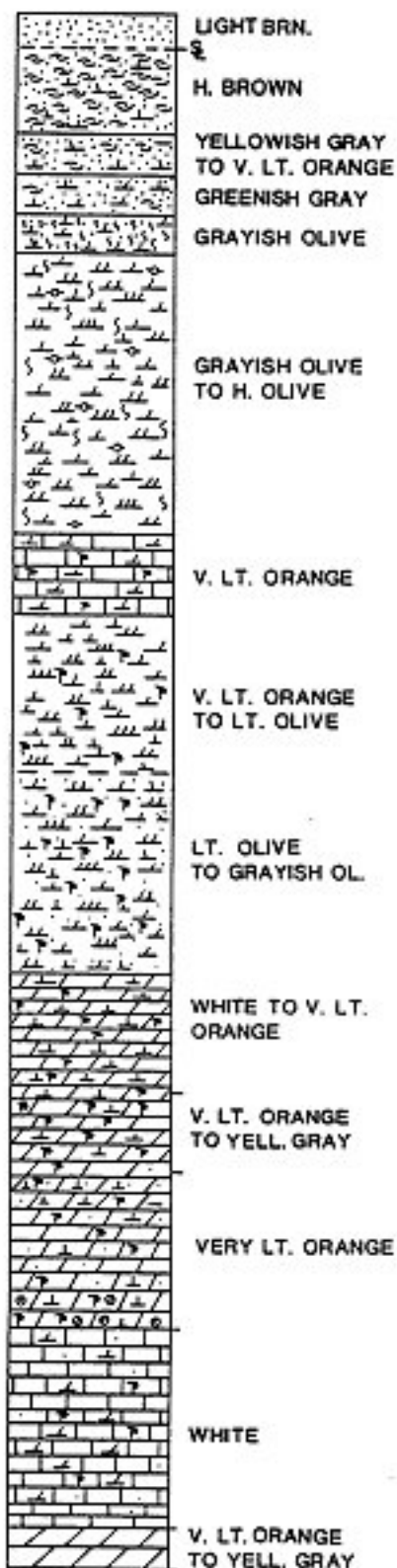


*14W-LE006

S. PINE ISLAND

meters feet

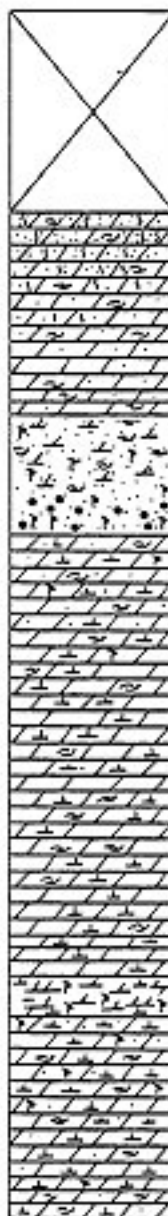
0 0
5 20
10 40
15 60
20 80
25 100
30 120
35 140
40 160
45 180
50 200
55 220
60 240
65 260
70 280
75 300
80 320
85 340
90 360
95 380
100 400
105 420
110 440



W-LE018 #15
THE LANDING

meters feet

0 0
5 20
10 40
15 60
20 80
25 100
30 120
35 140
40 160
45 180
50 200
55 220
60 240
65 260
70 280
75 300
80 320
85 340
90 360
95 380
100
105
110



V. LT. ORANGE
TO LT. OLIVE

LT. OLIVE

WH. TO V. LT. ORN.

GRN. OL GRN.

YELL. GRAY TO
WH.

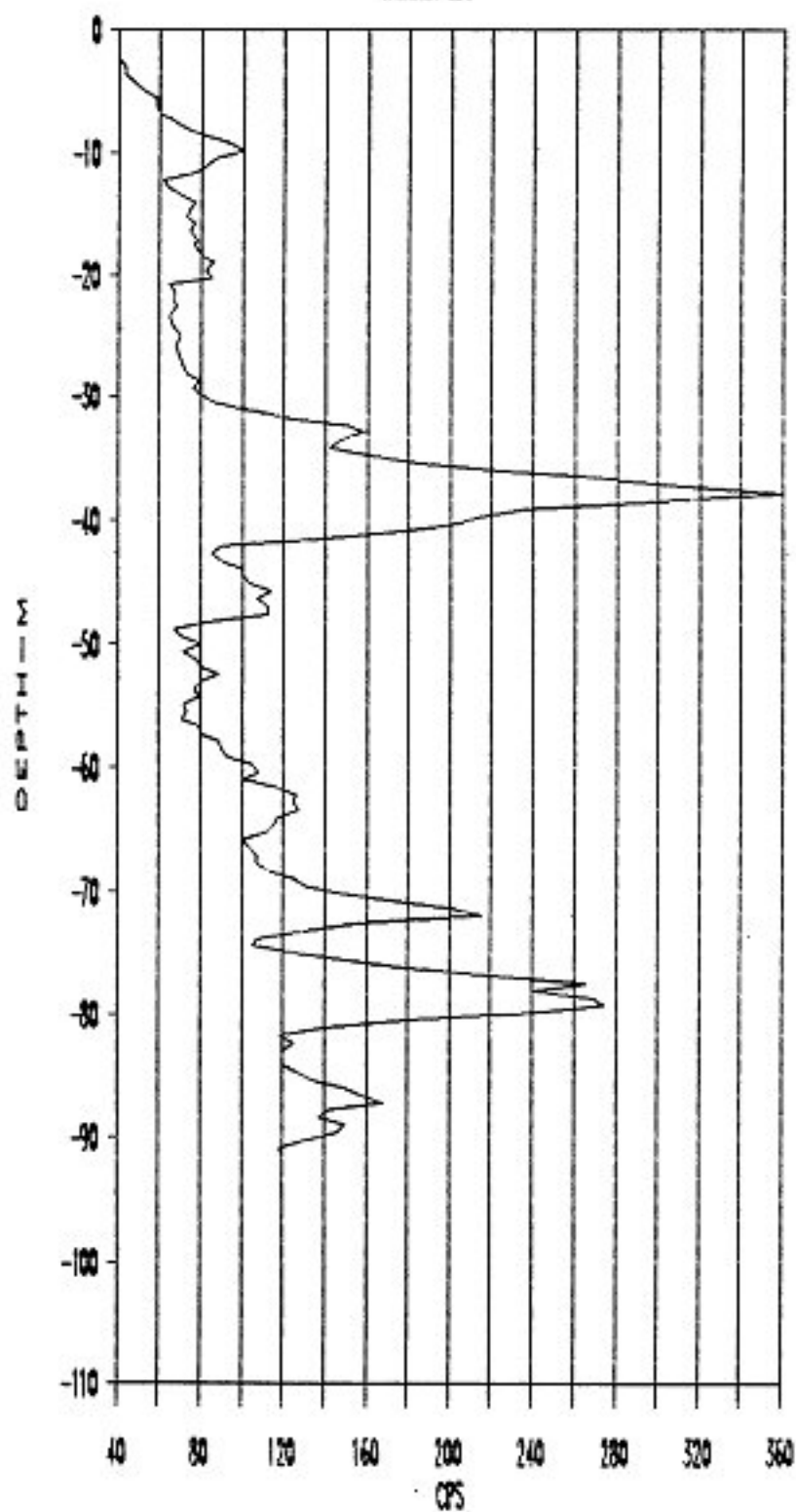
WHITE

YELLOWISH GRAY

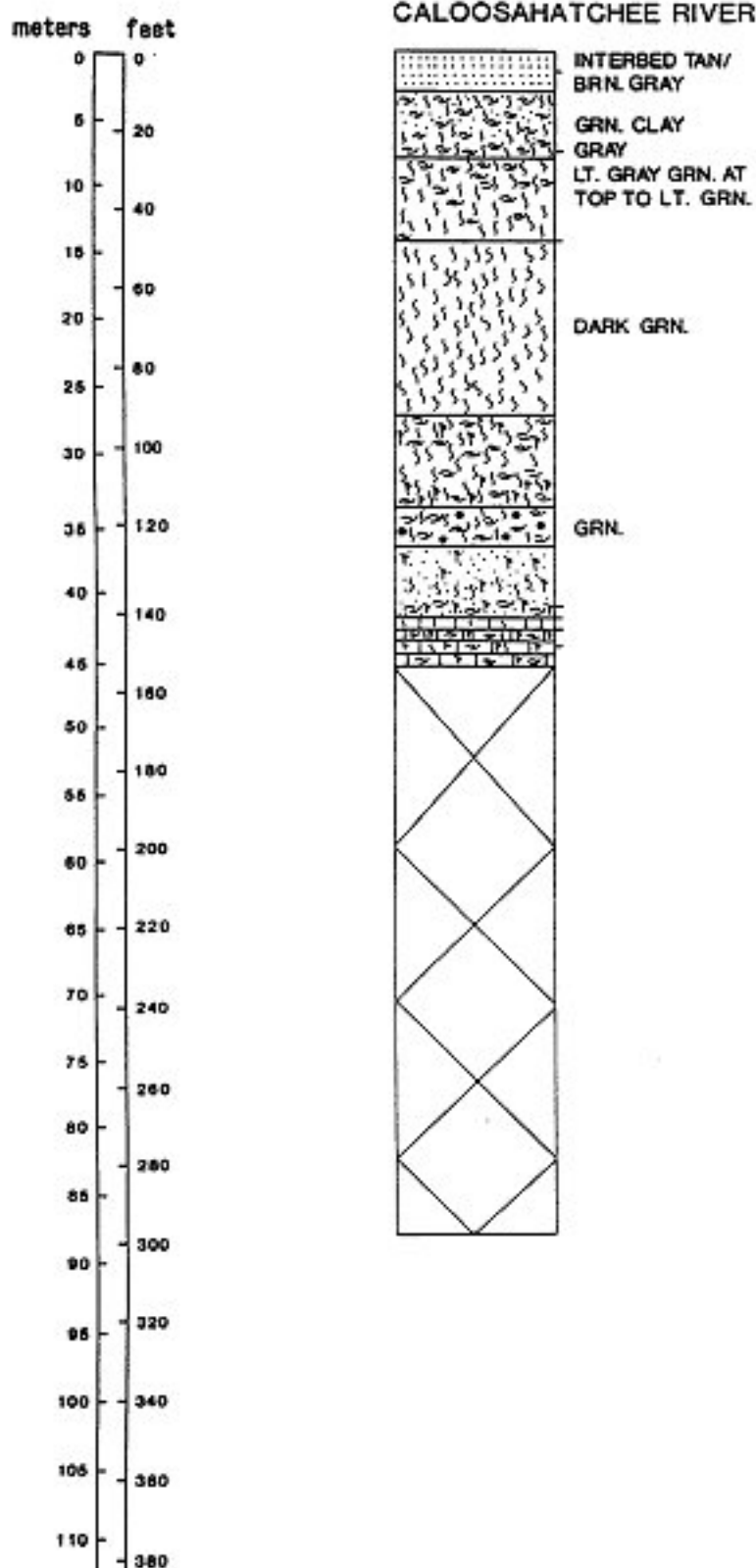
WHITE

WELL 15 (The Landings)

GAMMA LOG



CALOOSA HATCHEE RIVER

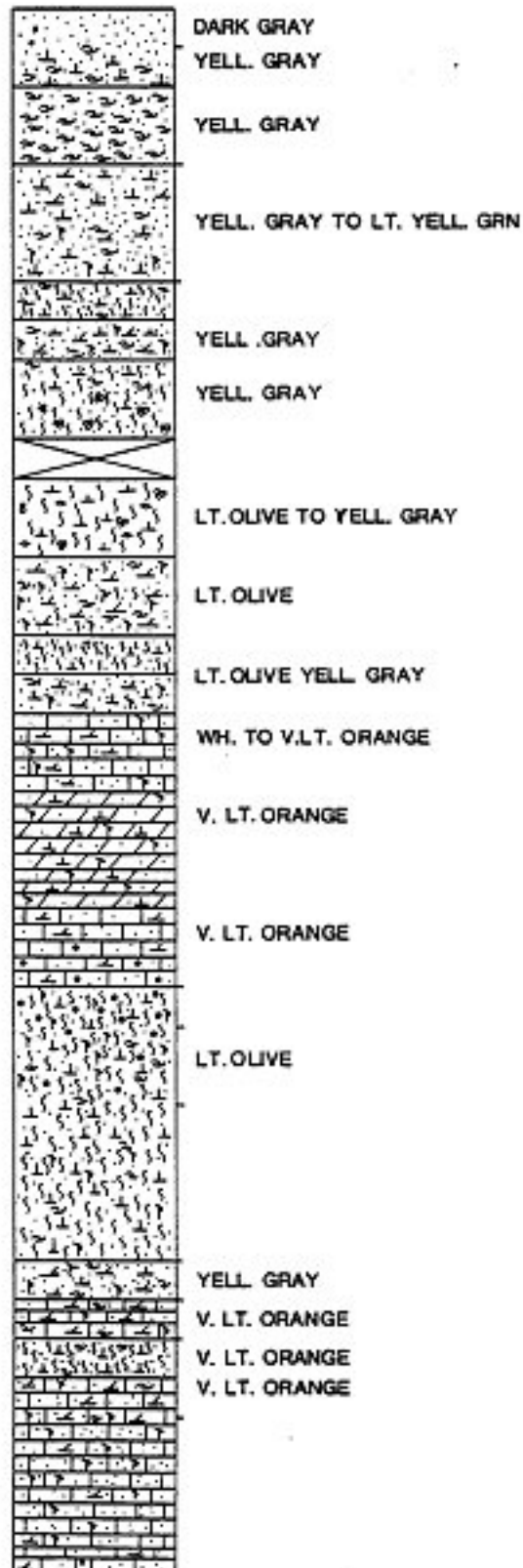


#17 W-035

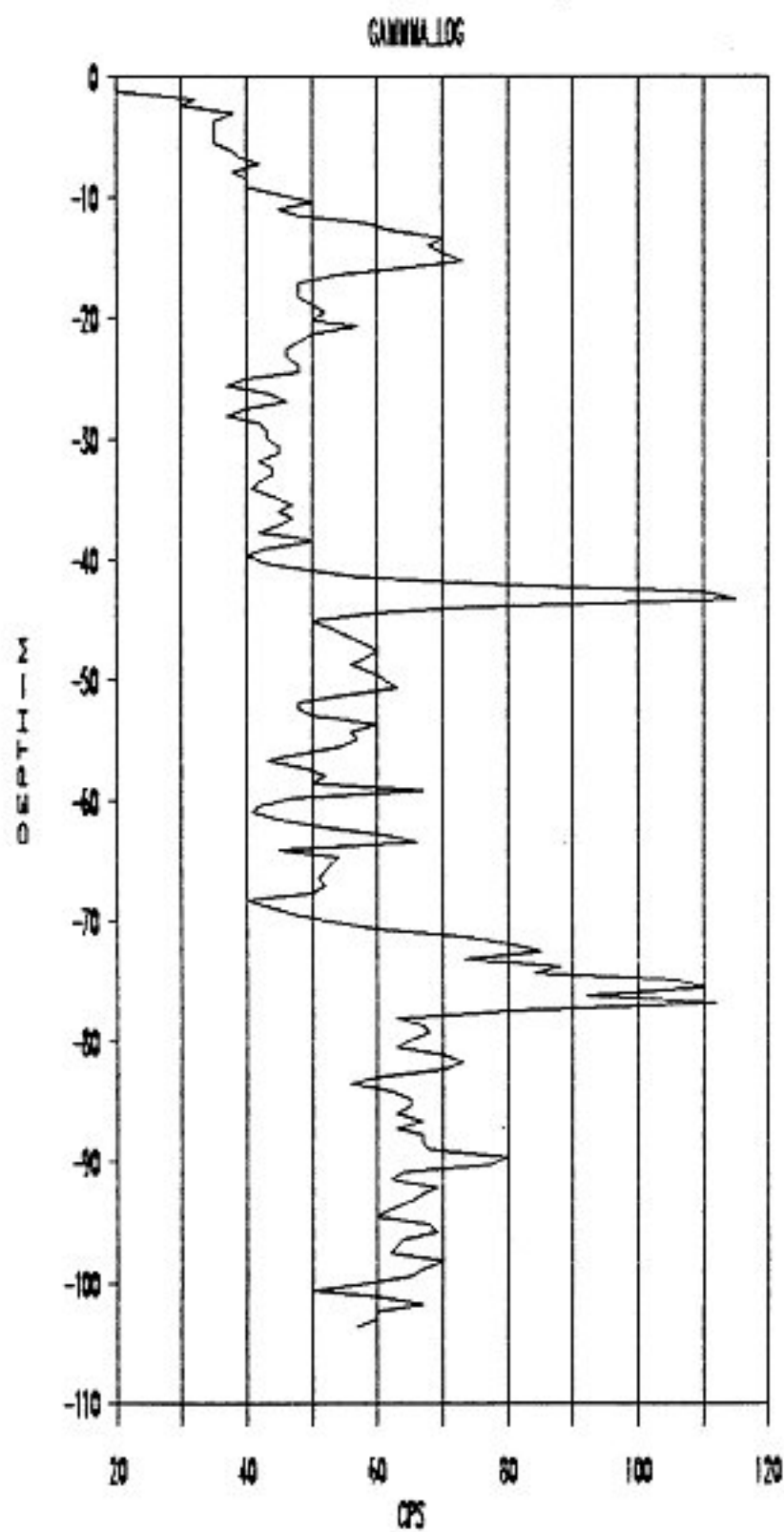
ST. JAMES CITY

meters feet

0 0
5 20
10 40
15 60
20 80
25 100
30 120
35 140
40 160
45 180
50 200
55 220
60 240
65 260
70 280
75 300
80 320
85 340
90 360
95 380
100
105
110

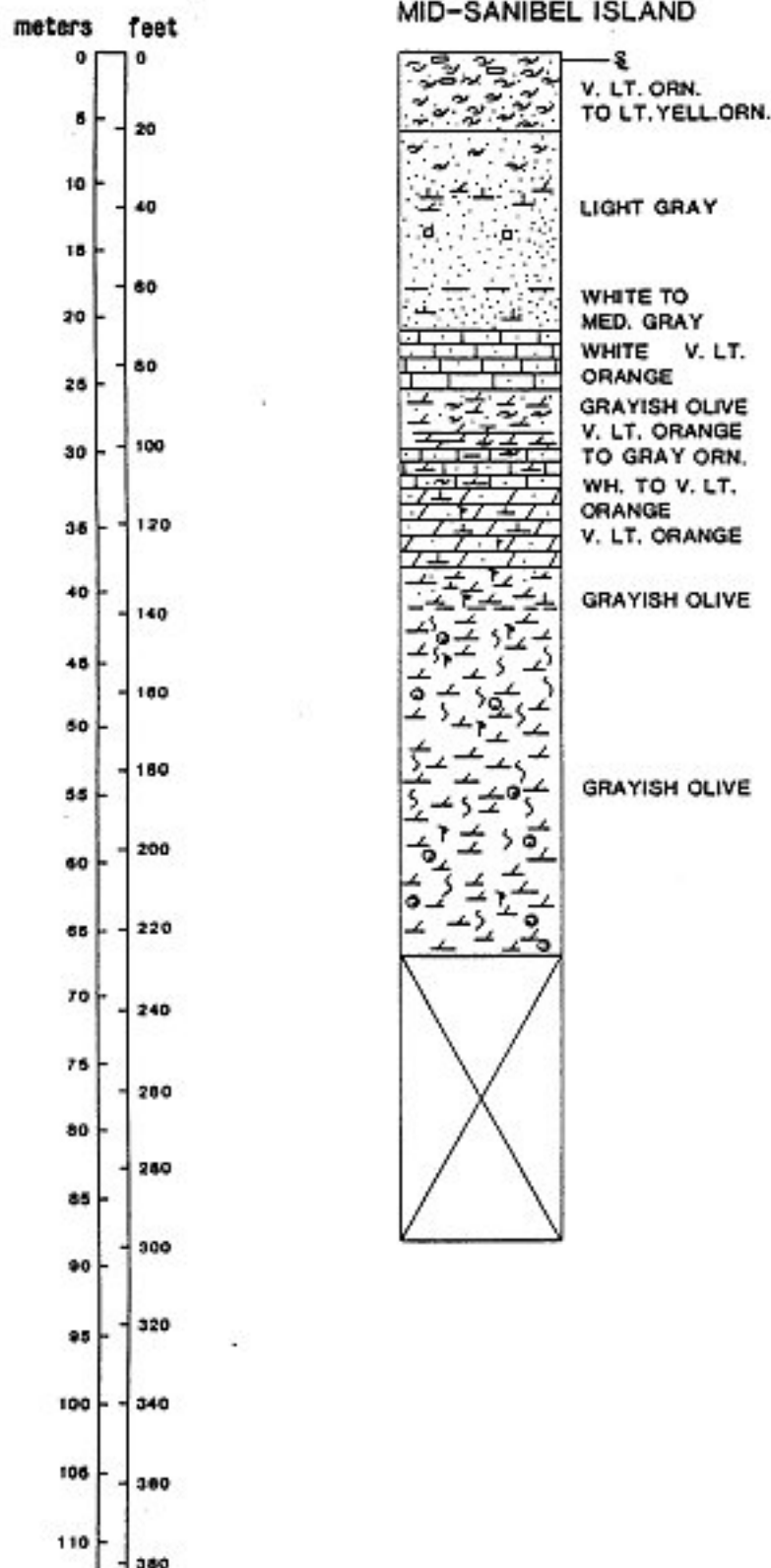


WELL 17 (ST. JAMES)



*18 W-LE026

MID-SANIBEL ISLAND

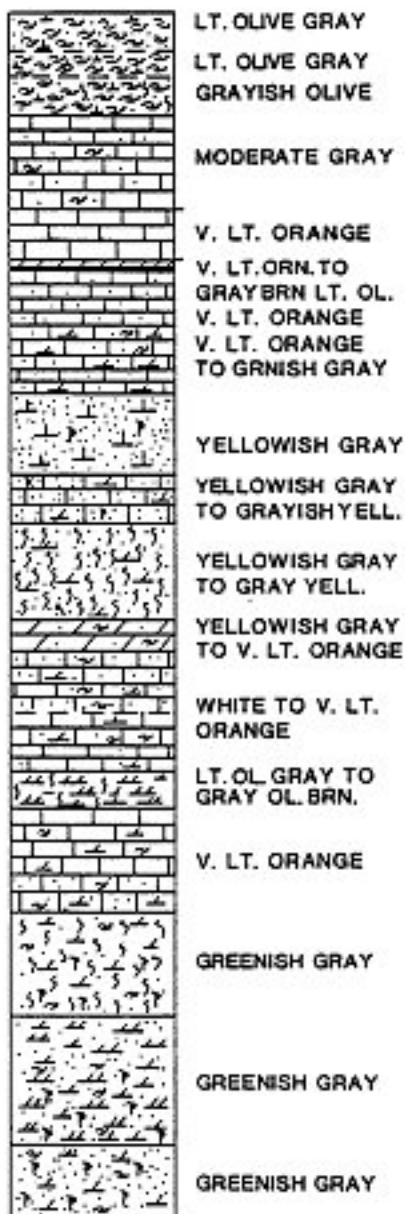


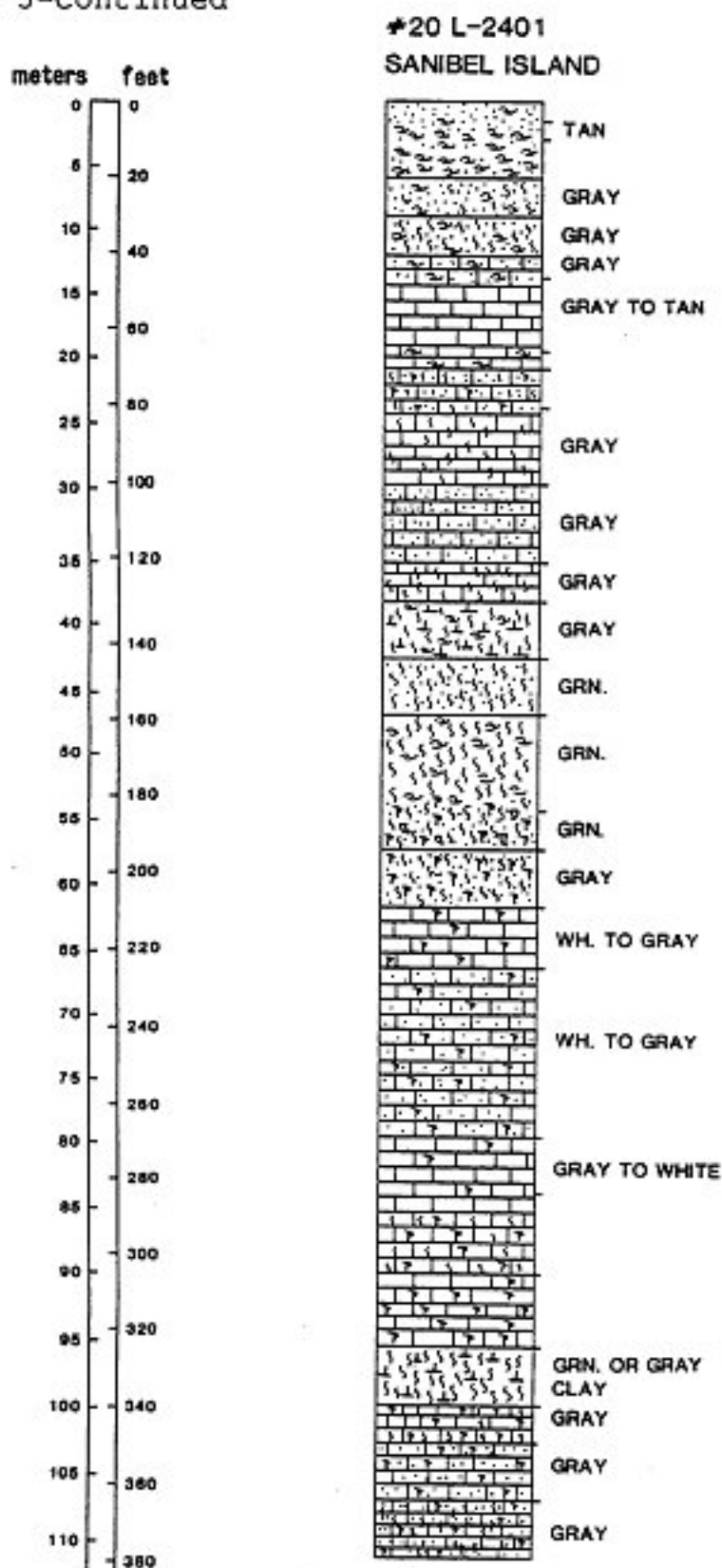
#19 W-LE004

SANIBEL ISLAND

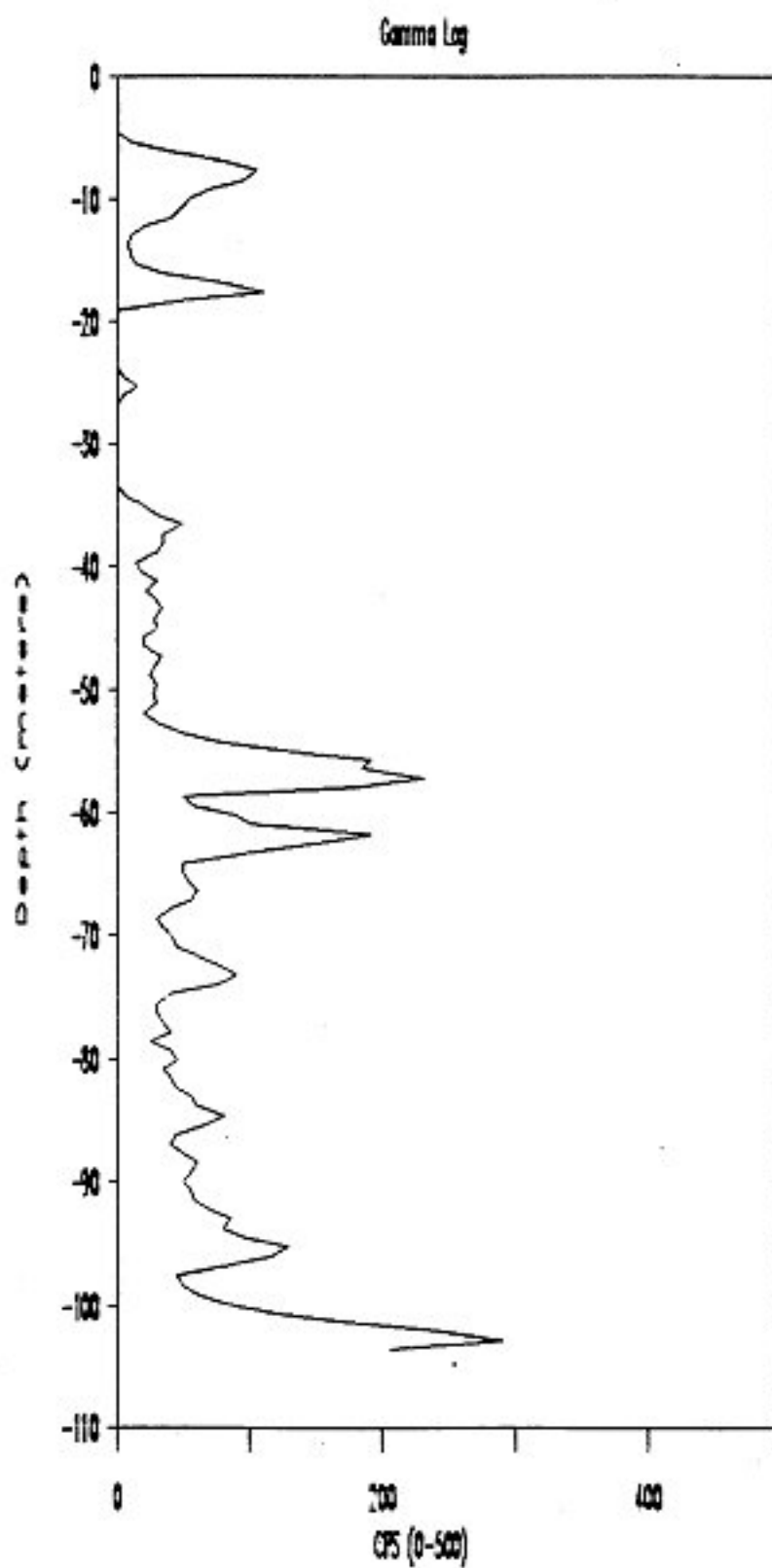
meters feet

0 0
5 20
10 40
15 60
20 80
25 100
30 120
35 140
40 160
45 180
50 200
55 220
60 240
65 260
70 280
75 300
80 320
85 340
90 360
95 380
100 400
105 420
110 440



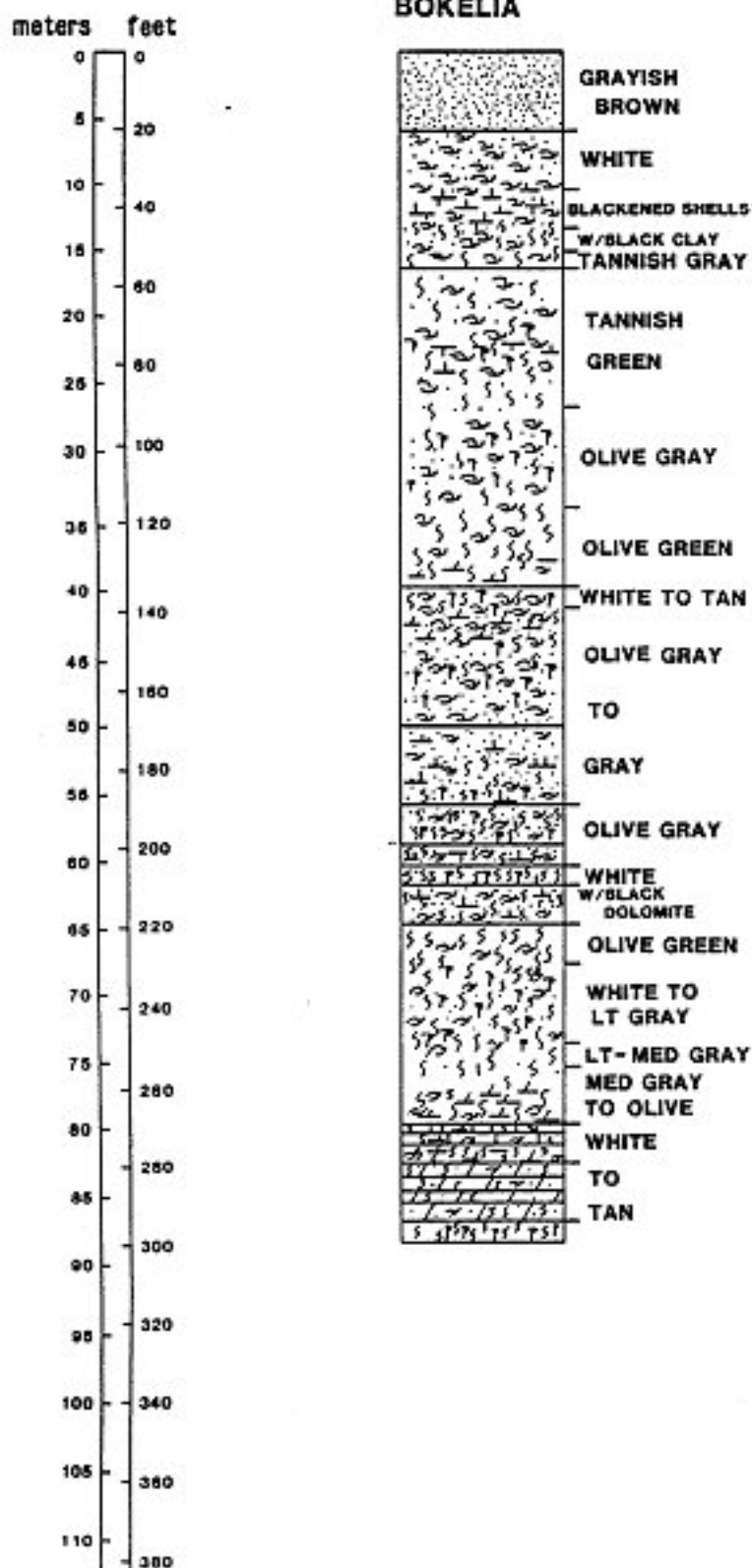


WELL 20 (SANIBEL ISLAND)

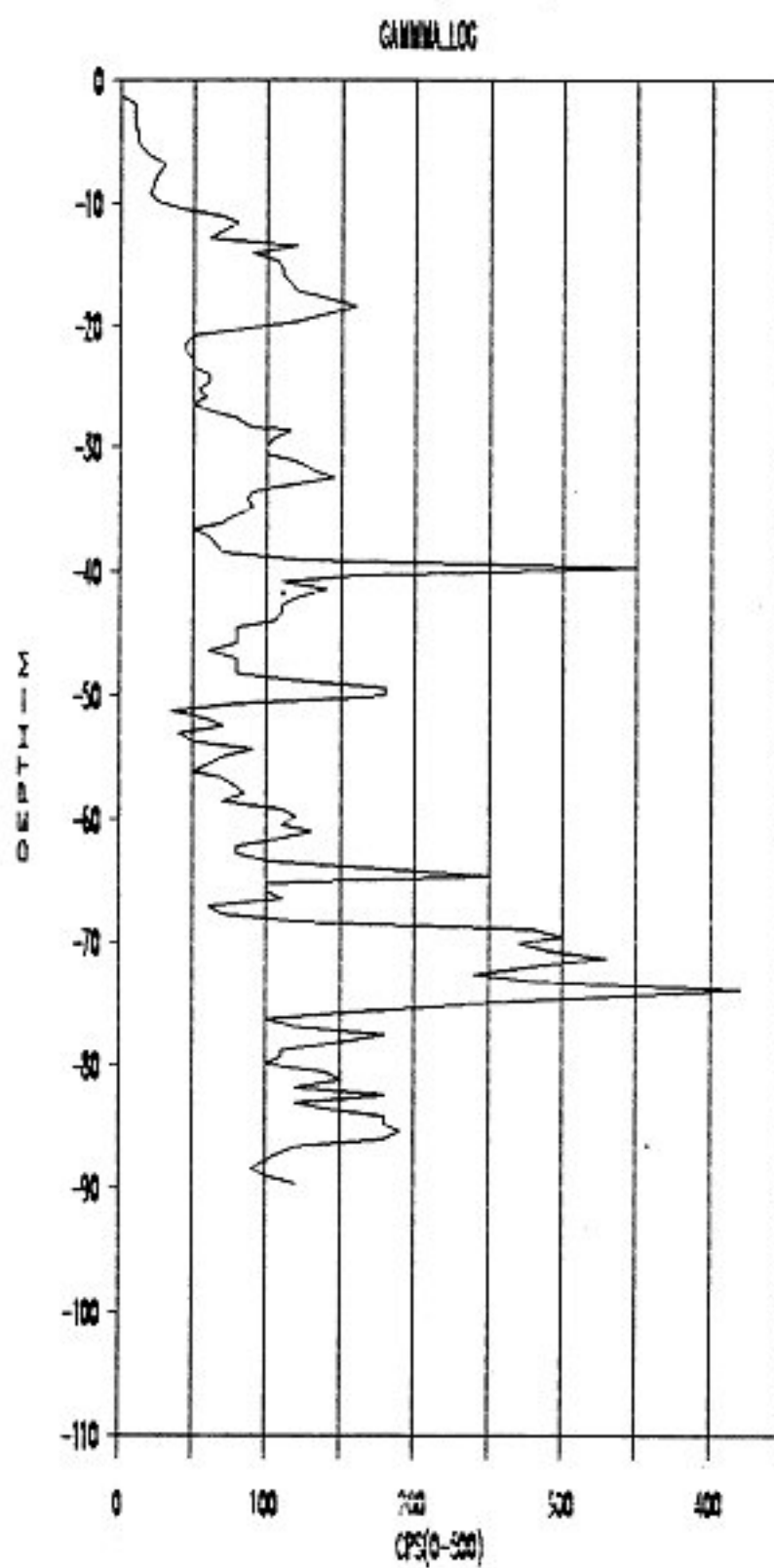


★ 21 (SEIS 1)

BOKELIA

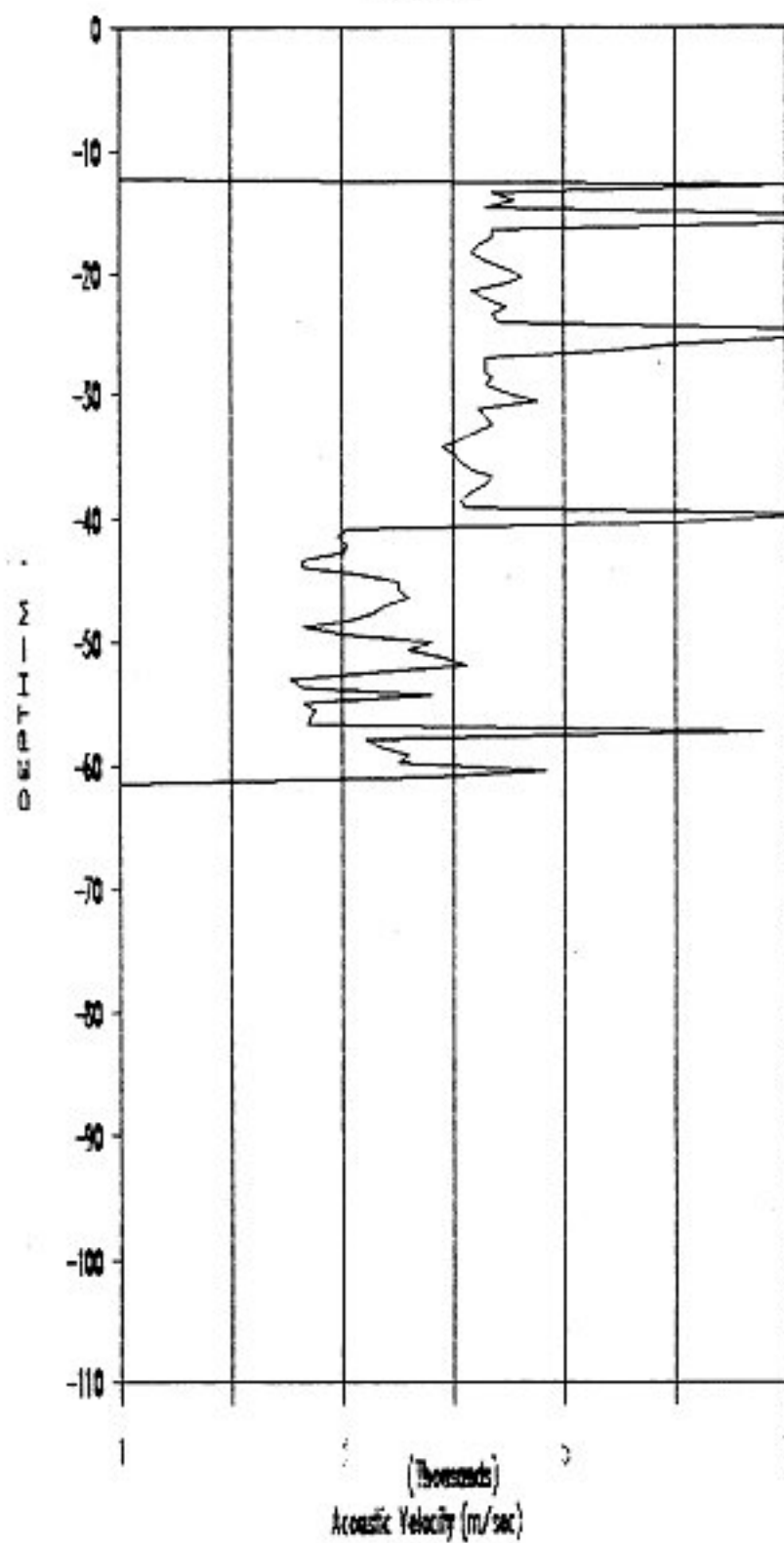


WELL 21 (BOKELIA)

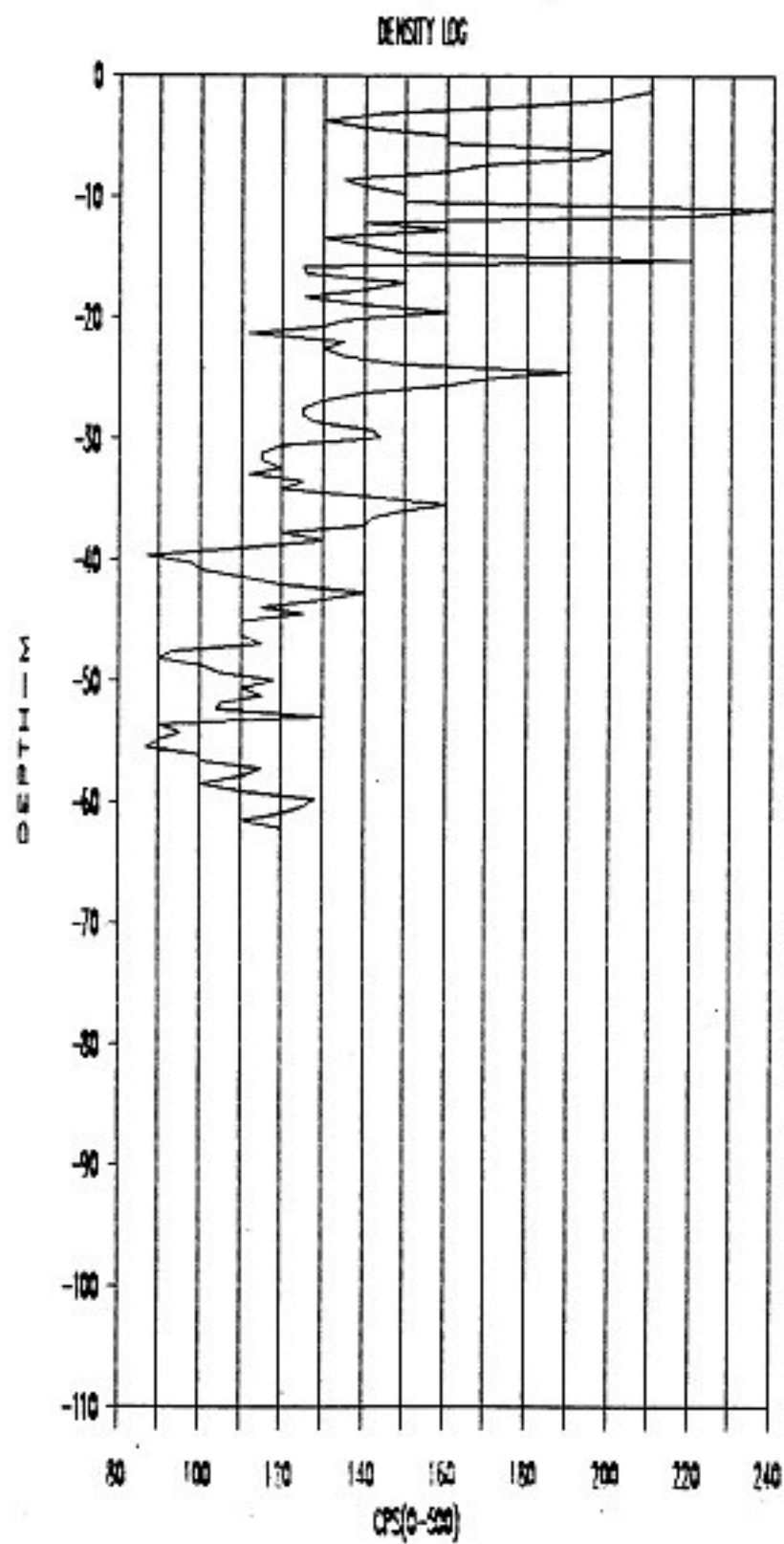


WELL 21 (BOKELIA)

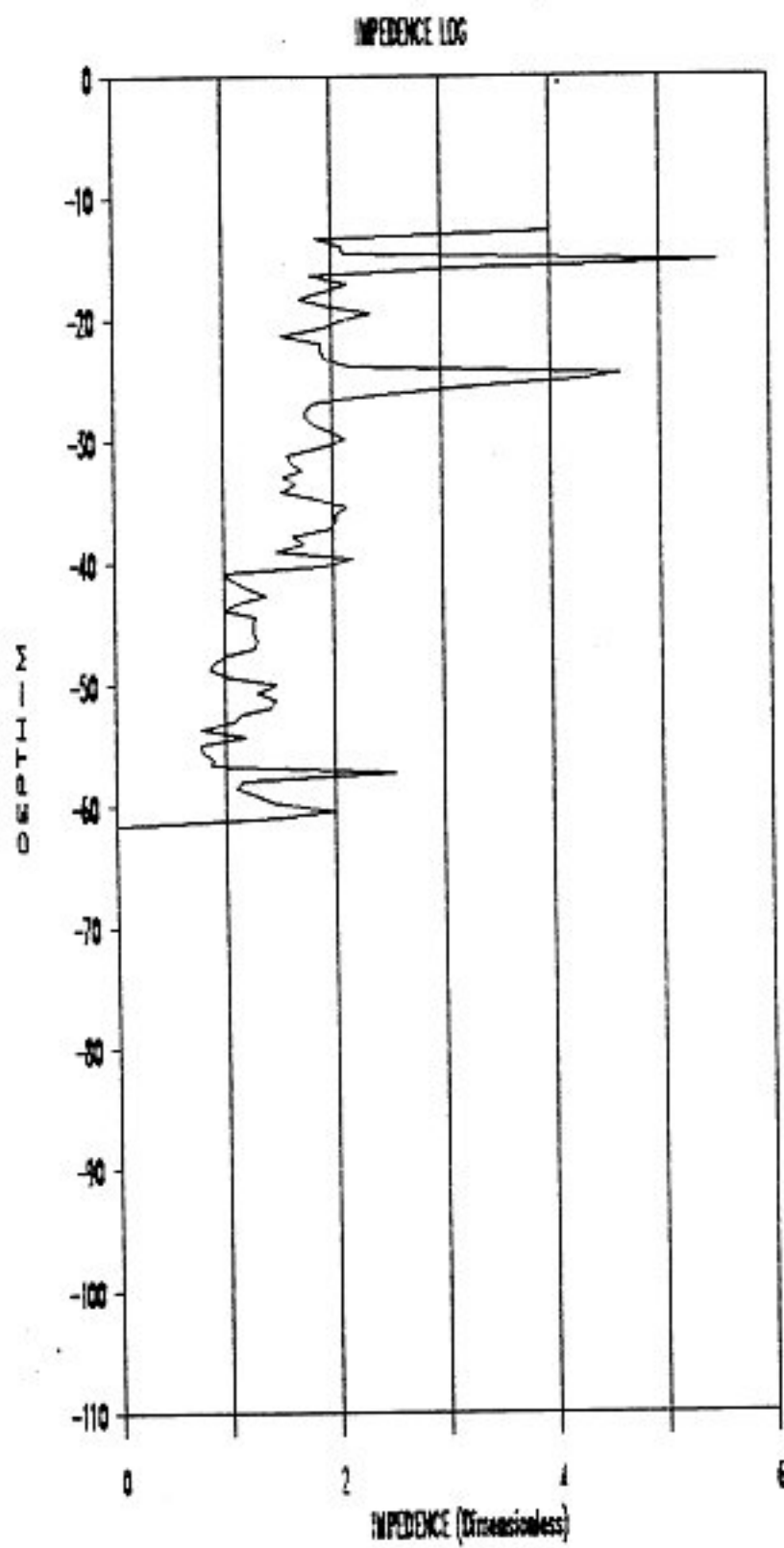
VELOCITY.LOG

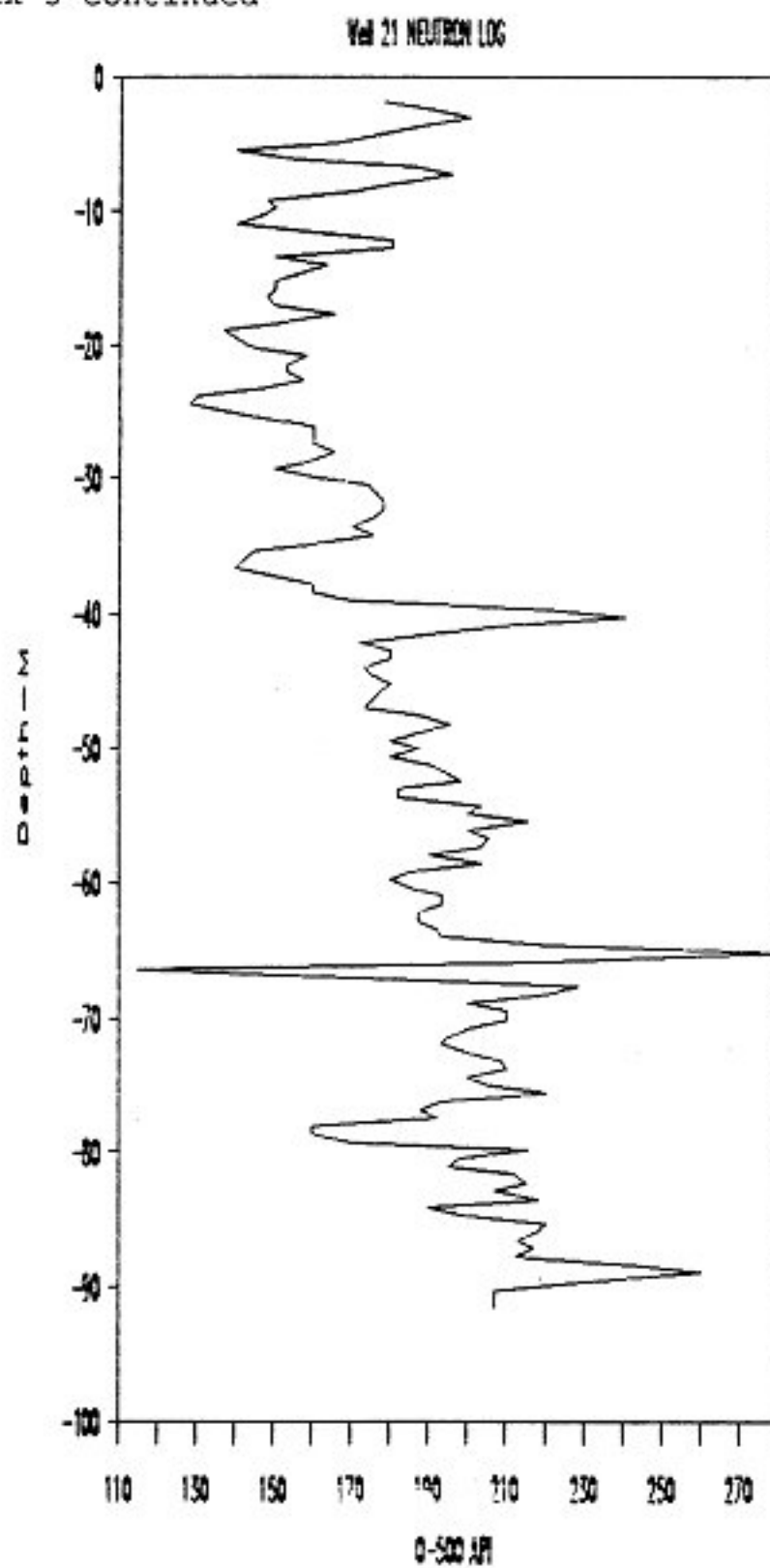


WELL 21 (BOKELIA)



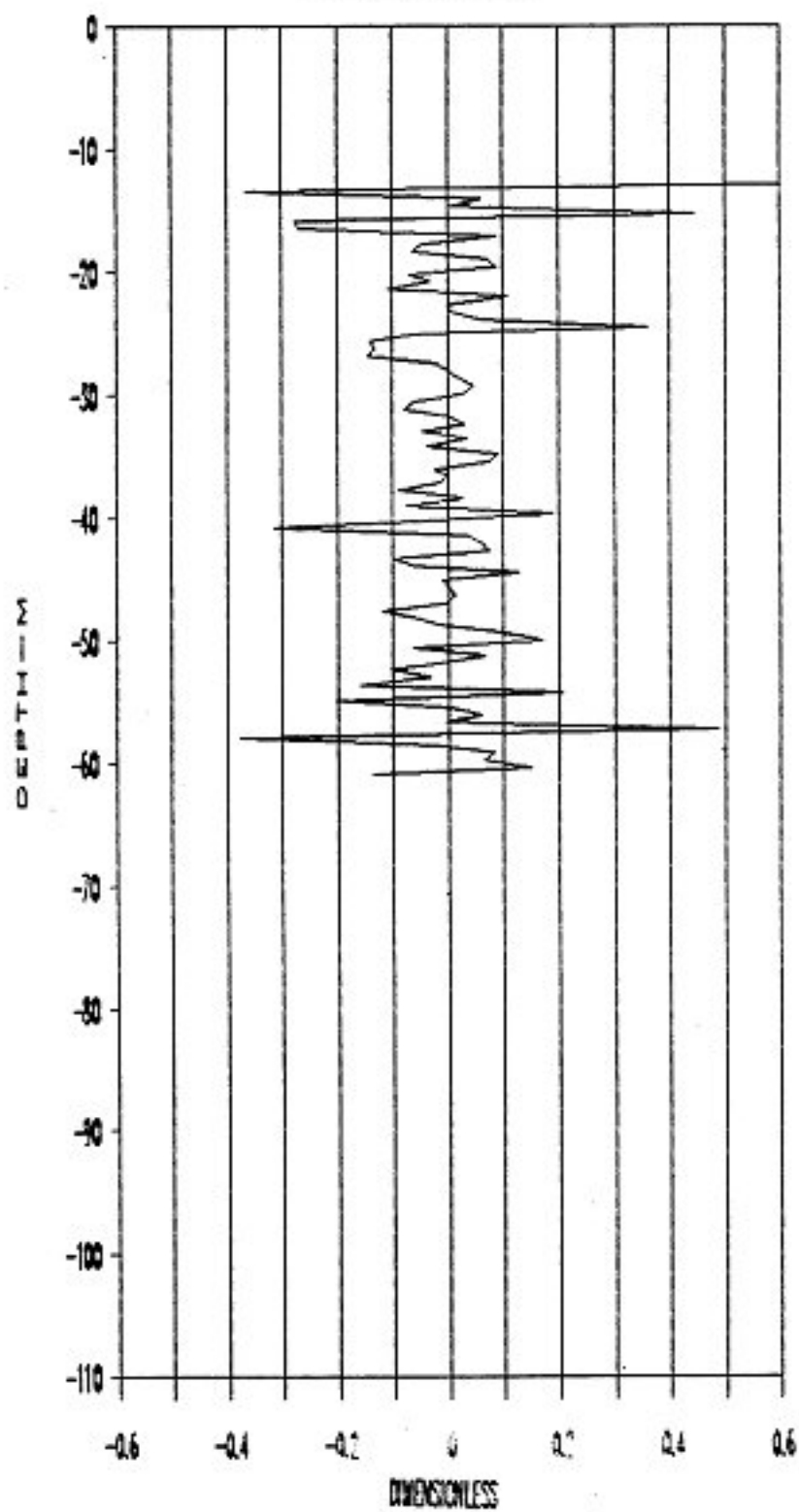
WELL 21 (BOLEKIA)





WELL 21 (BOKELIA)

REFLECTION COEFFICIENT LOG



meters feet

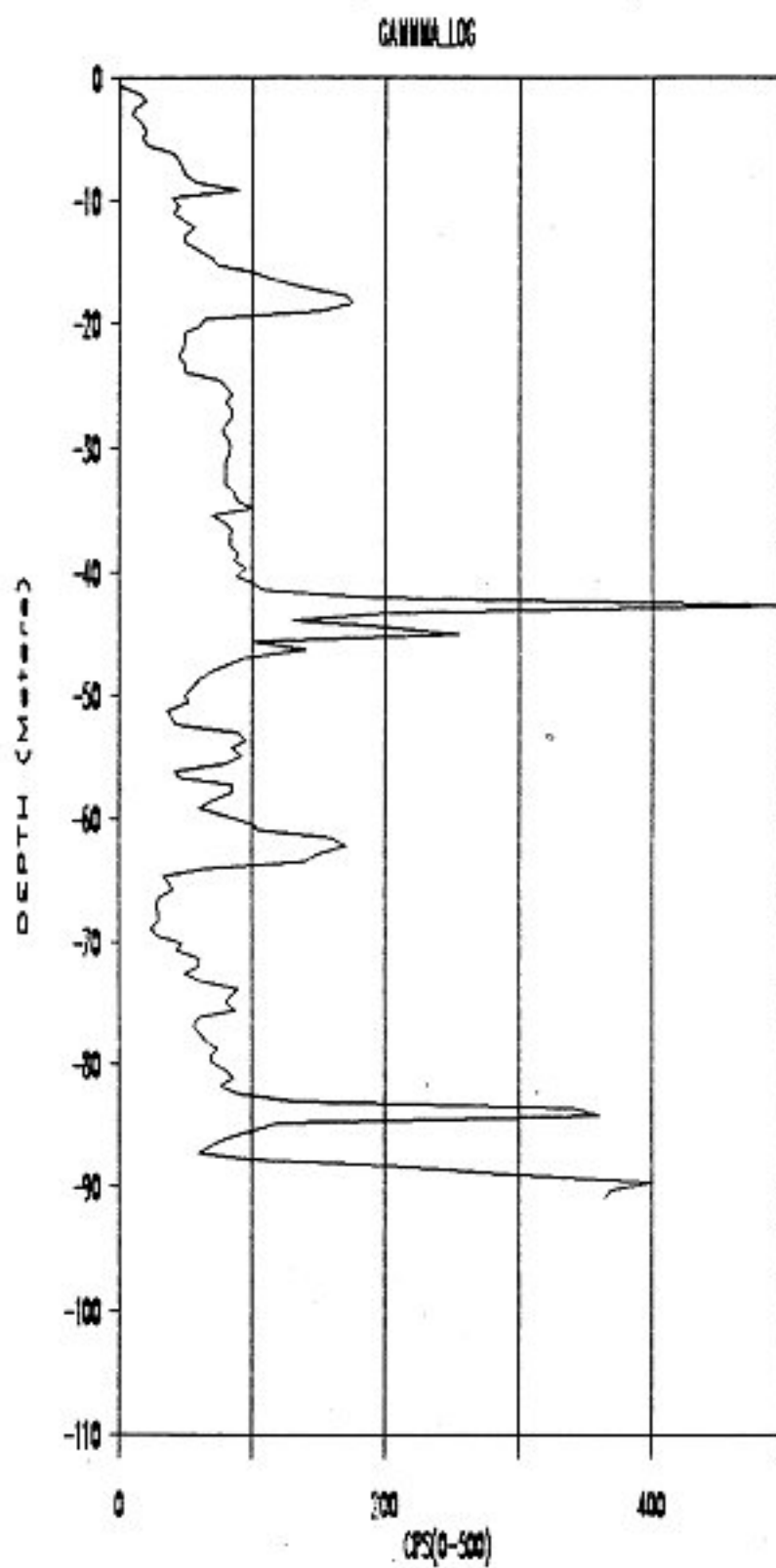
0 0
 5 20
 10 40
 15 60
 20 80
 25 100
 30 120
 35 140
 40 160
 45 180
 50 200
 55 220
 60 240
 65 260
 70 280
 75 300
 80 320
 85 340
 90 360
 95 380
 100
 105
 110

#22 (SEIS 2)

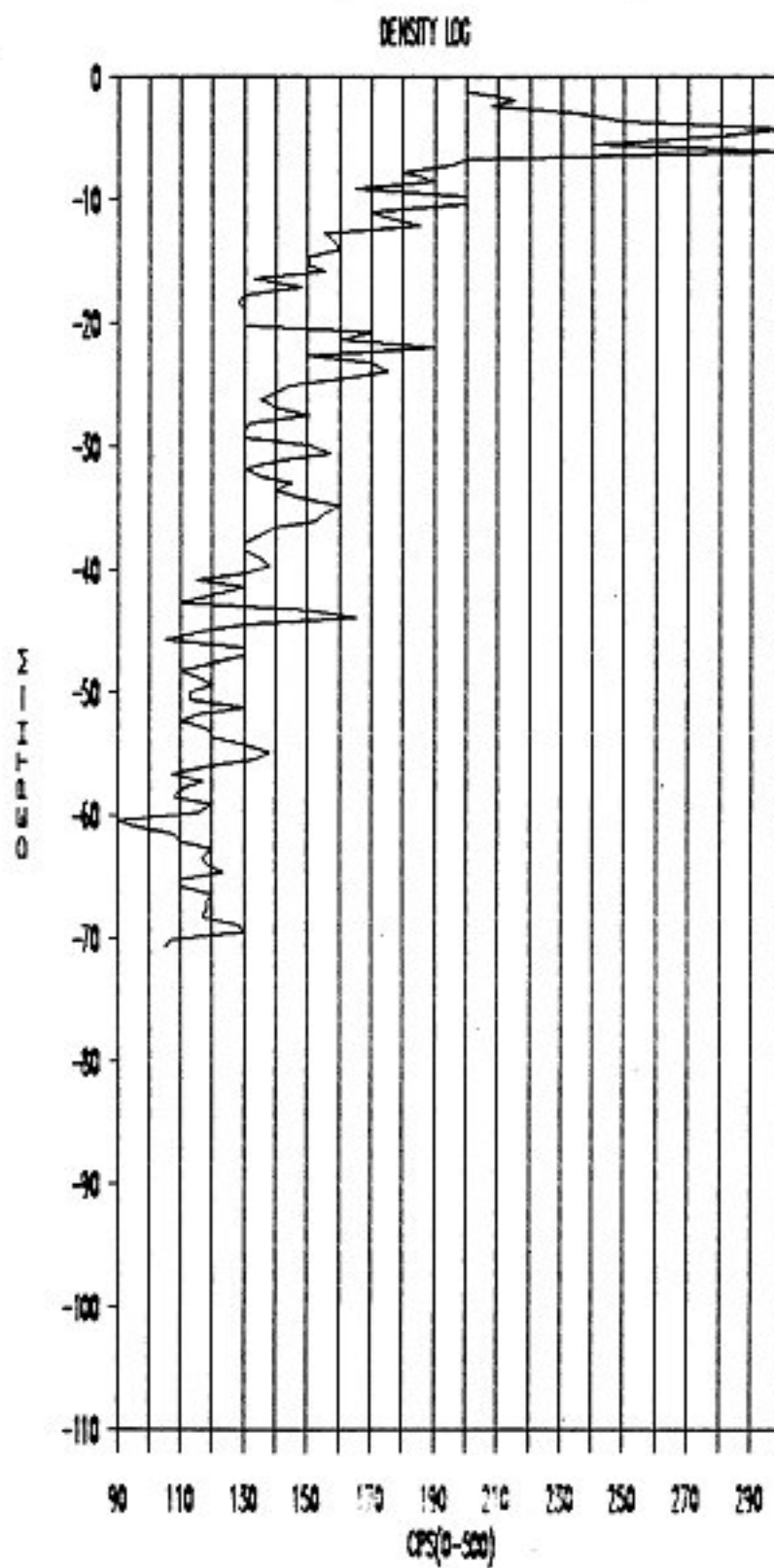
SANIBEL CAUSEWAY



WELL 22 (SANIBEL CAUSEWAY)

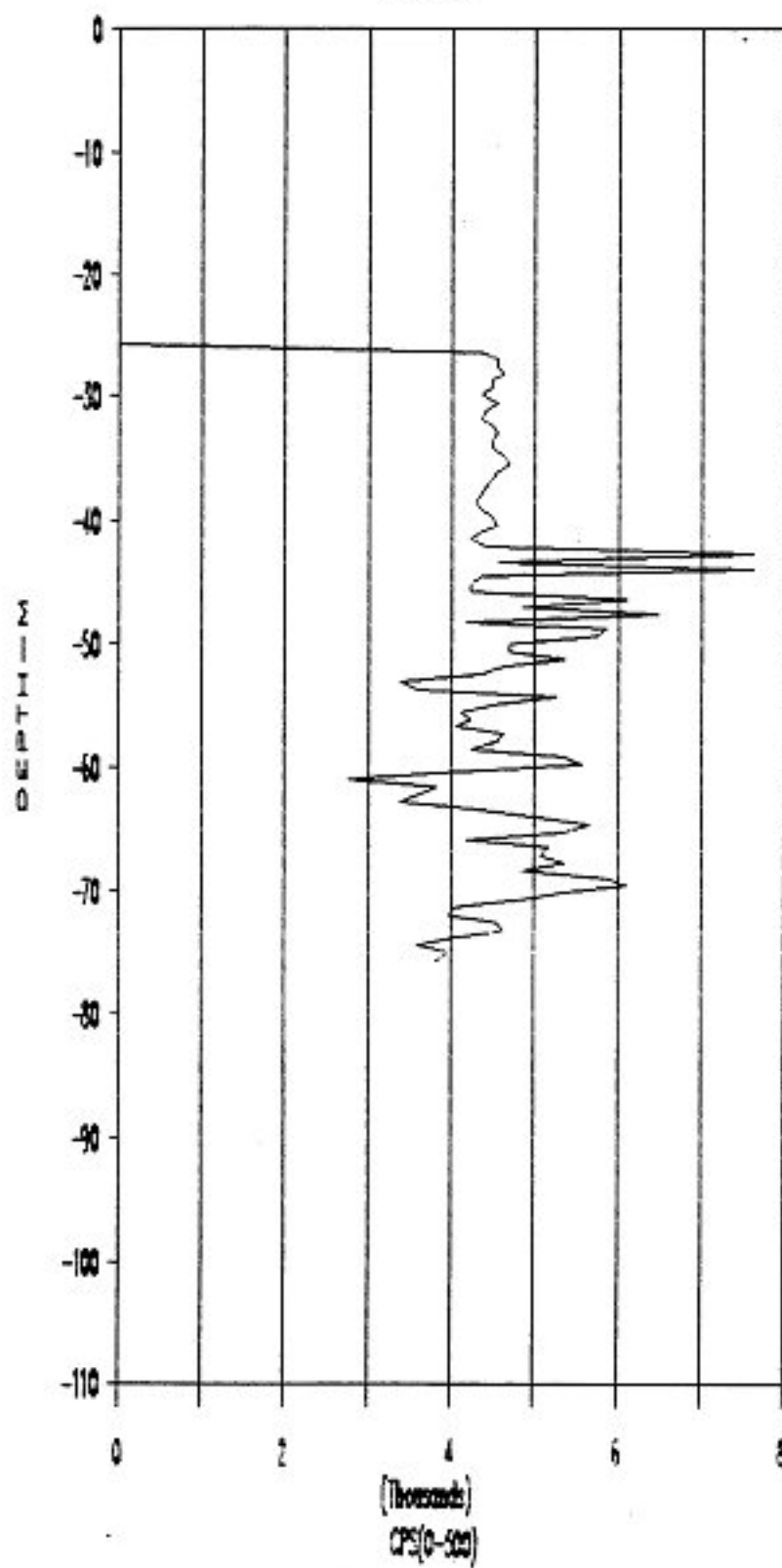


WELL 22 (SANIBEL CAUSEWAY)

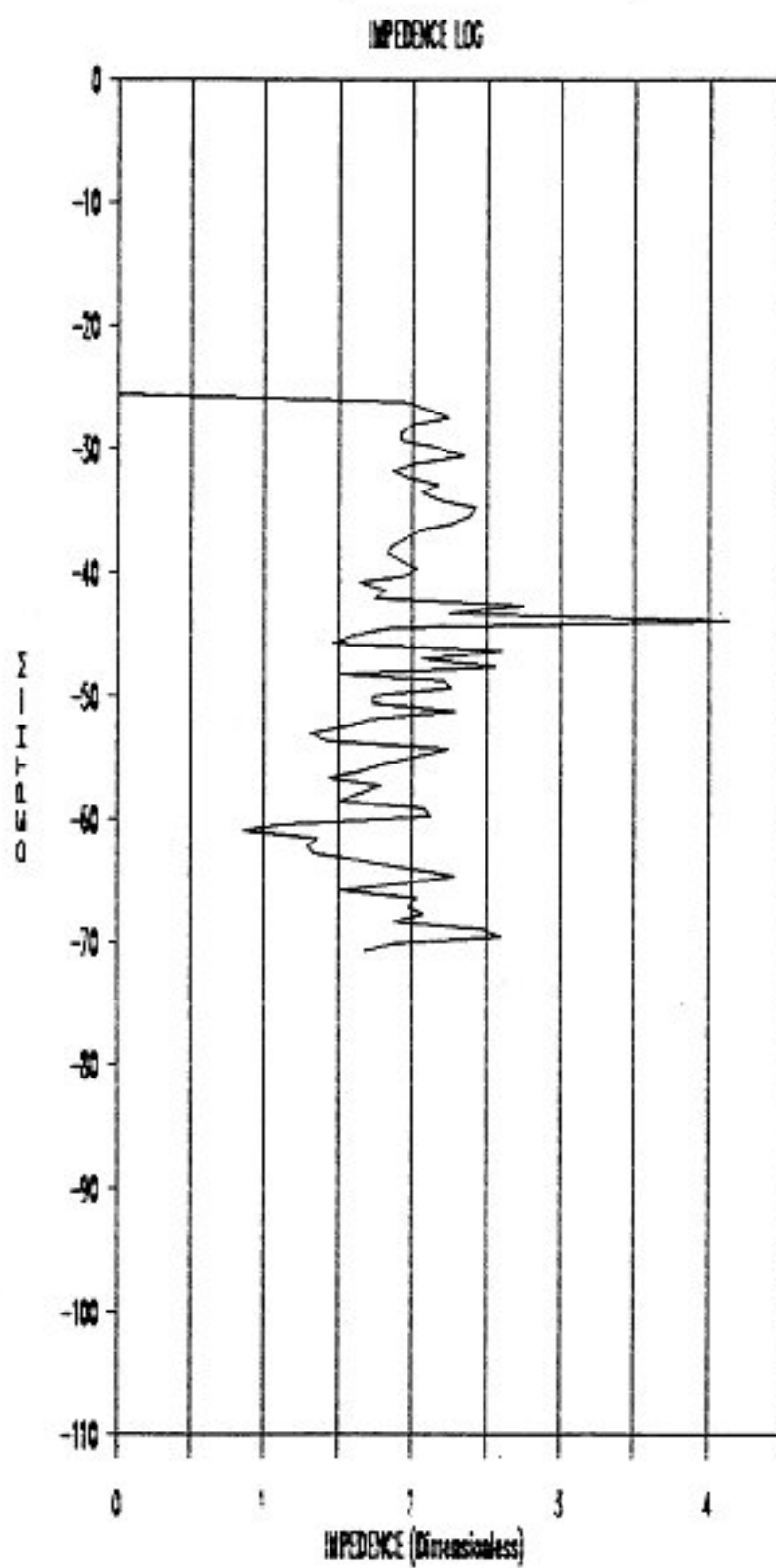


WELL 22

VELOCITY LOG

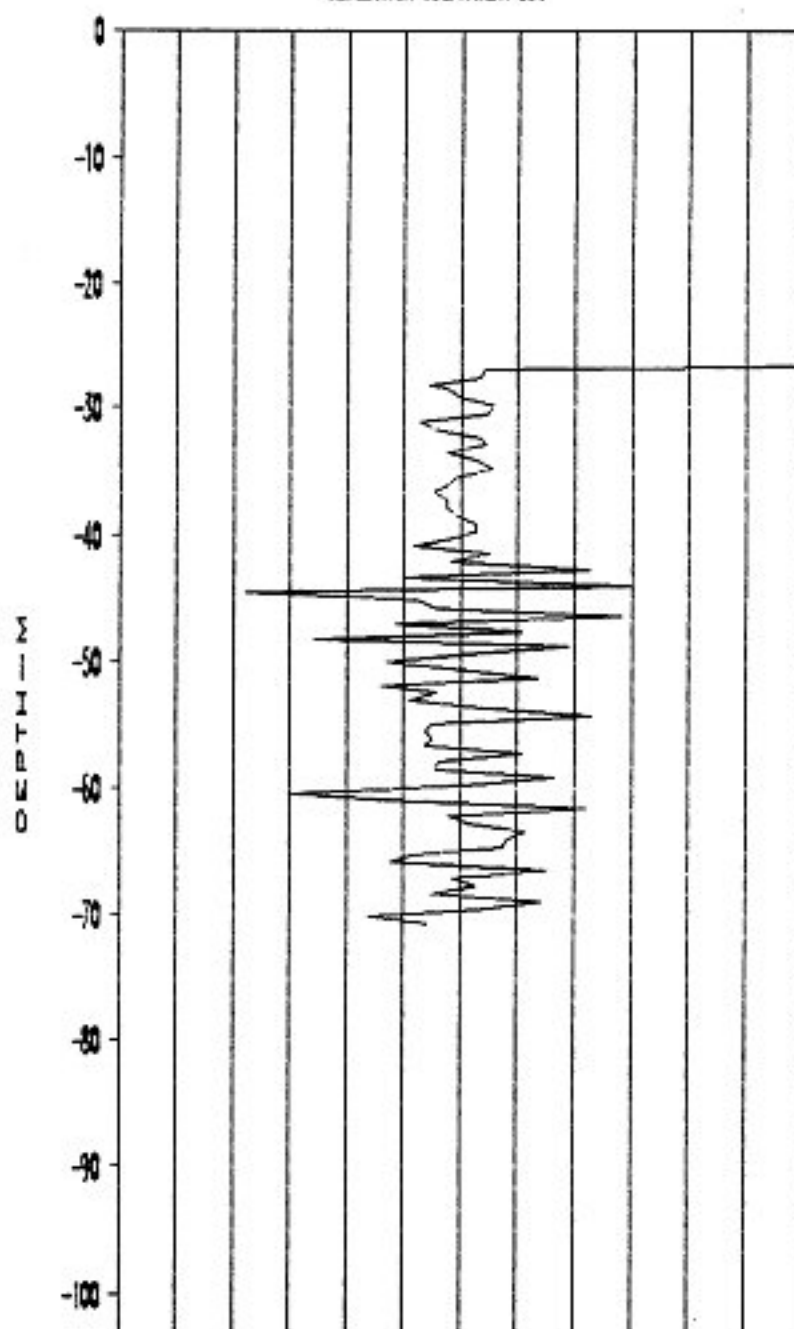


WELL 22 (SANIBEL CAUSEWAY)



WELL 22 (SANIBEL CAUSEWAY)

REFLECTION COEFFICIENT LOG

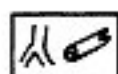


APPENDIX 4**CORRELATED SEISMIC AND
LITHOLOGIC CROSS SECTIONS
(REAR POCKET)**

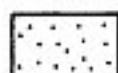
APPENDIX 5

VIBRACORE LOGS

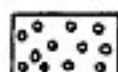
CORE LOG KEY



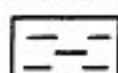
ROOTS



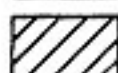
FINE-MEDIUM SAND



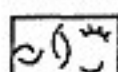
COARSE SAND/PEBBLES



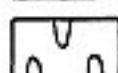
SILT



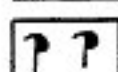
CLAY OR CLAYEY SAND



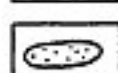
SHELLS OR SHELL FRAGMENTS



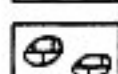
DISTINCT BURROWING OR MOTTLING



PHOSPHATE GRAINS



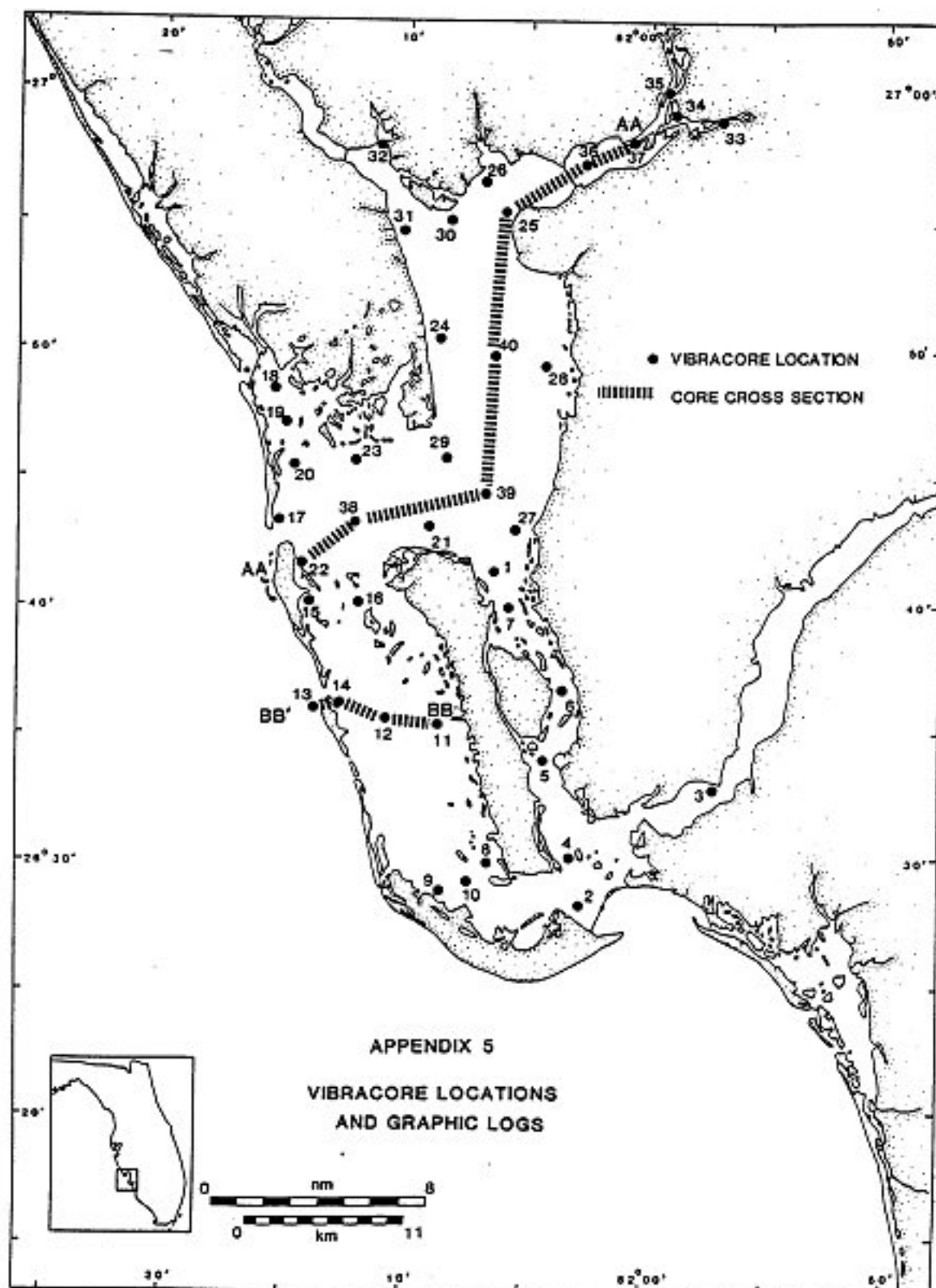
CEMENTED SAND

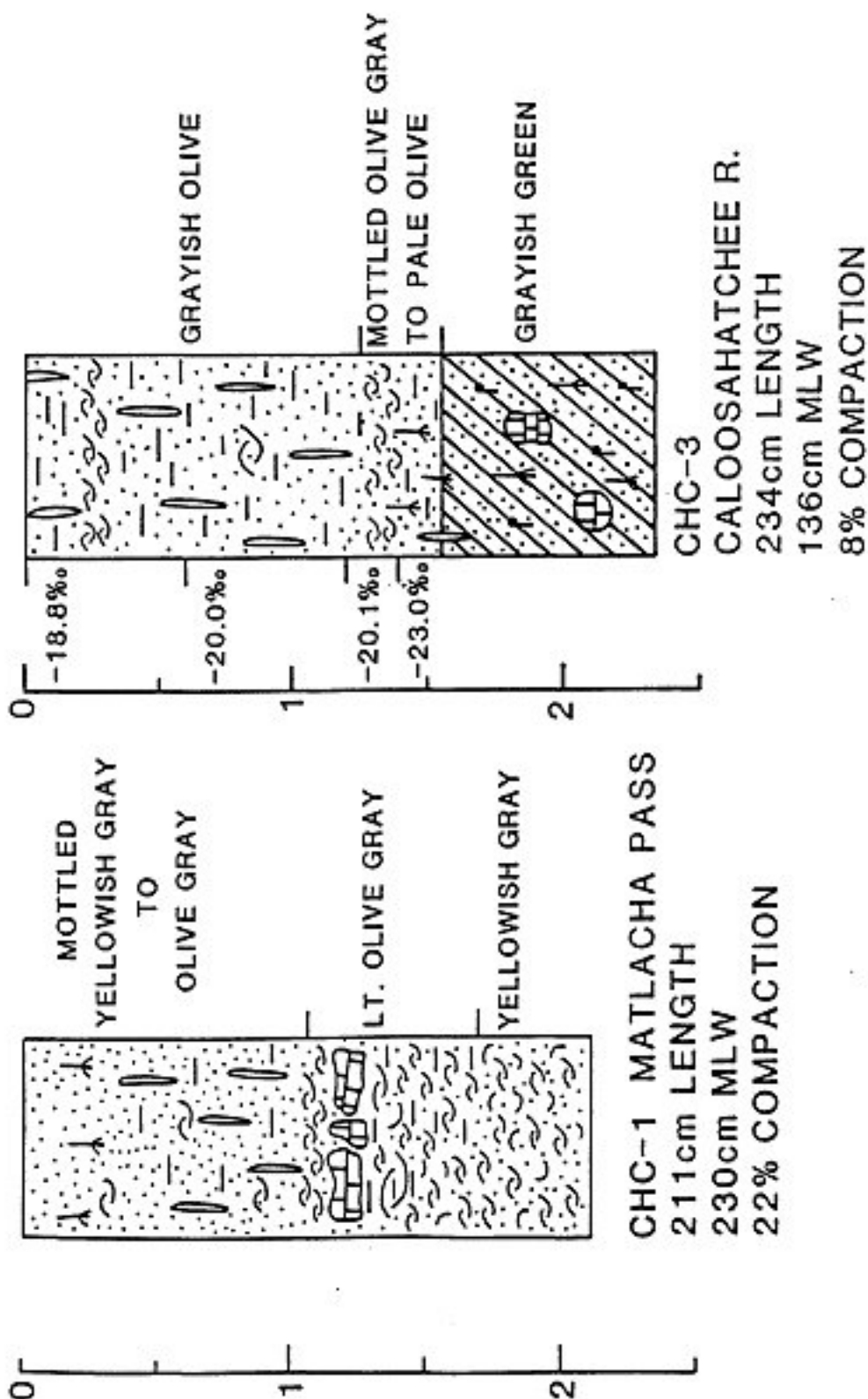


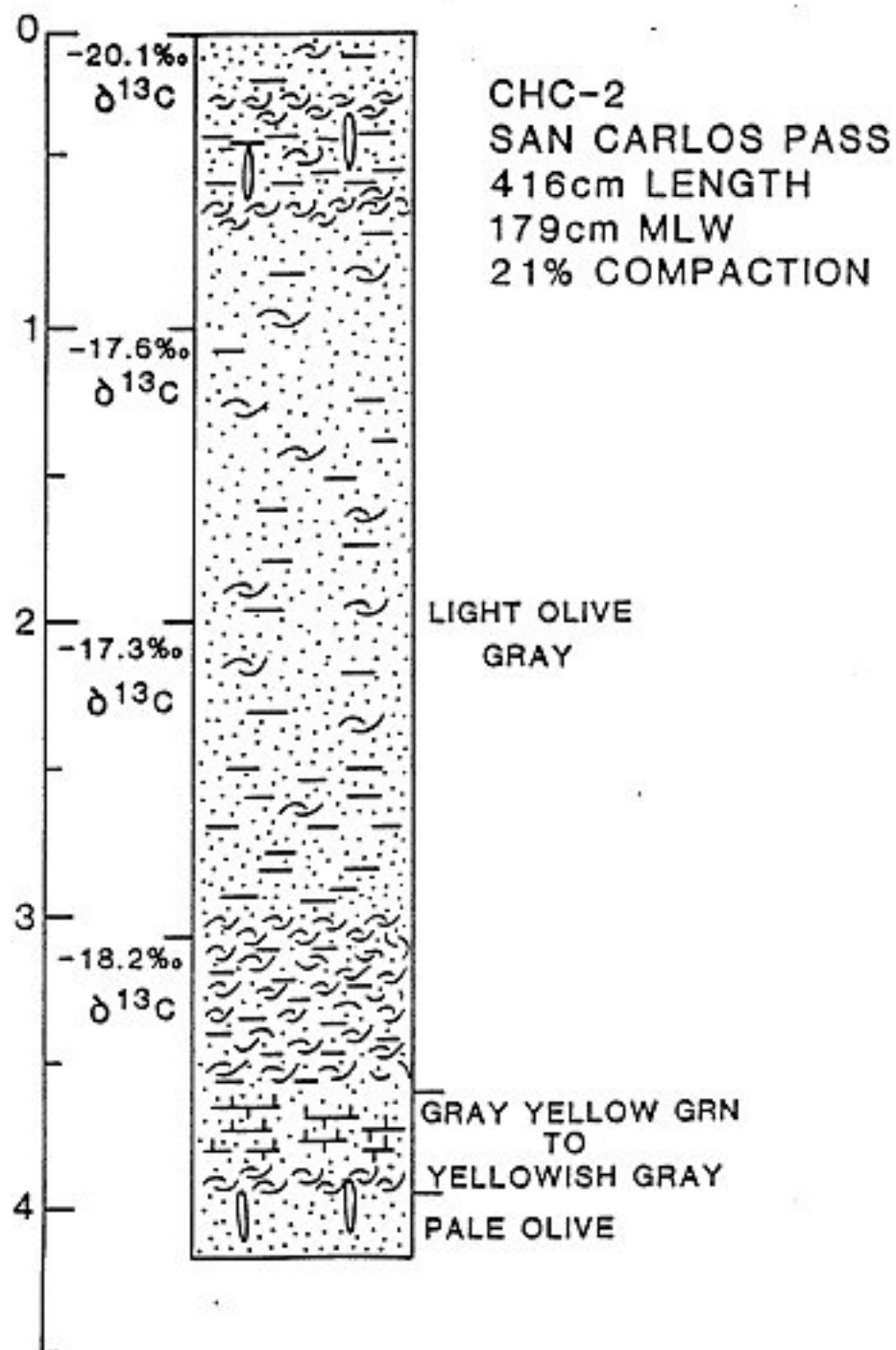
CALCAREOUS CRUST

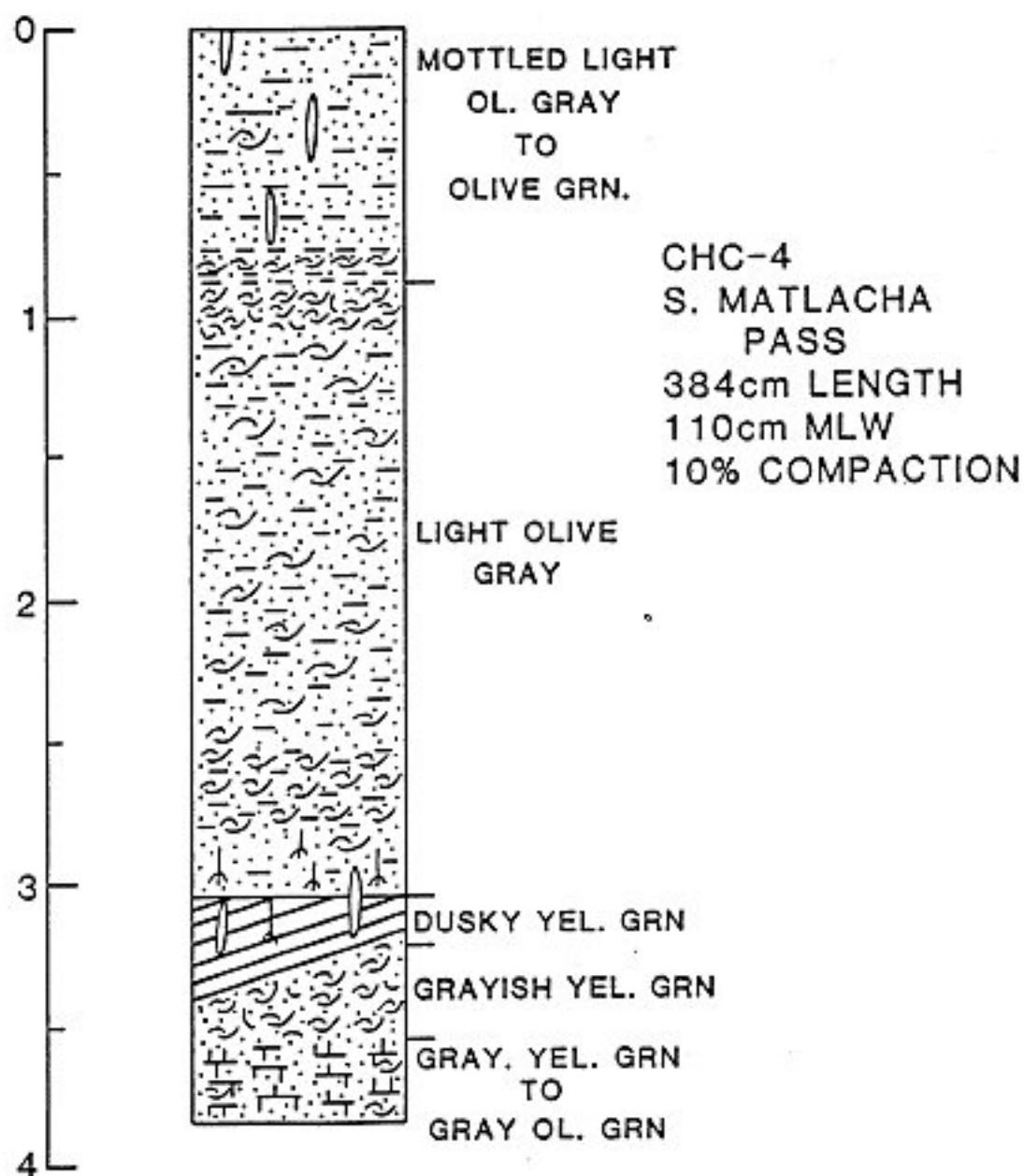
GRAPHIC
LITHOLOGY $\delta^{13}\text{C}$
(‰-PDB)

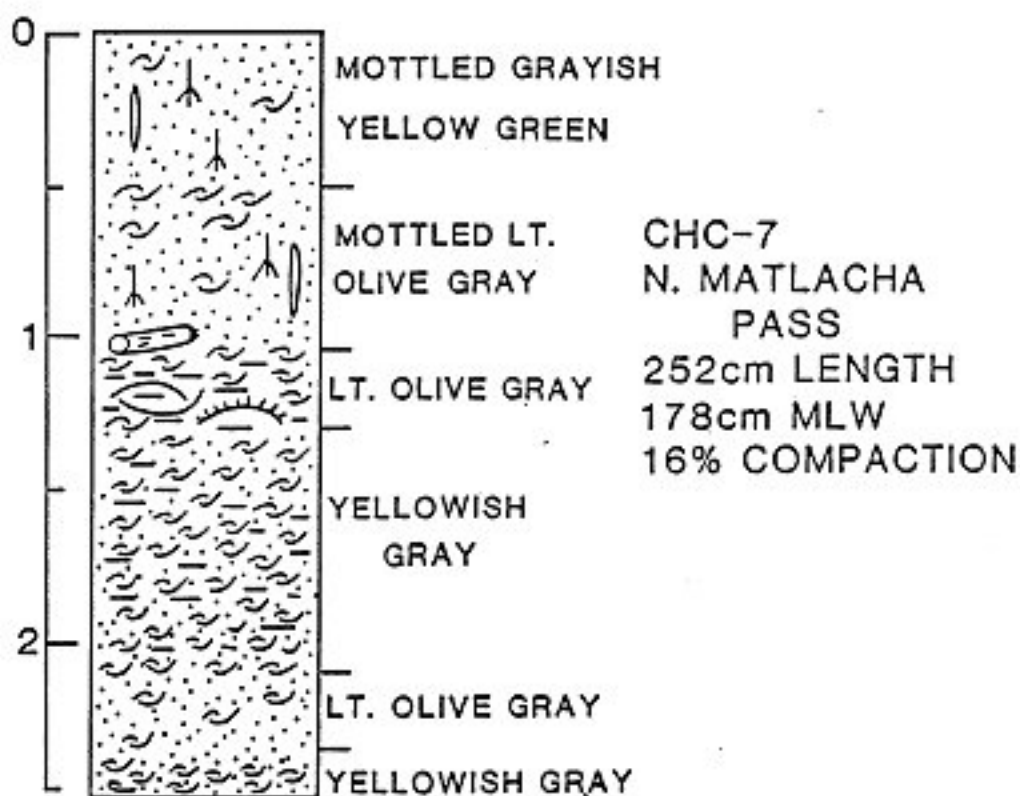
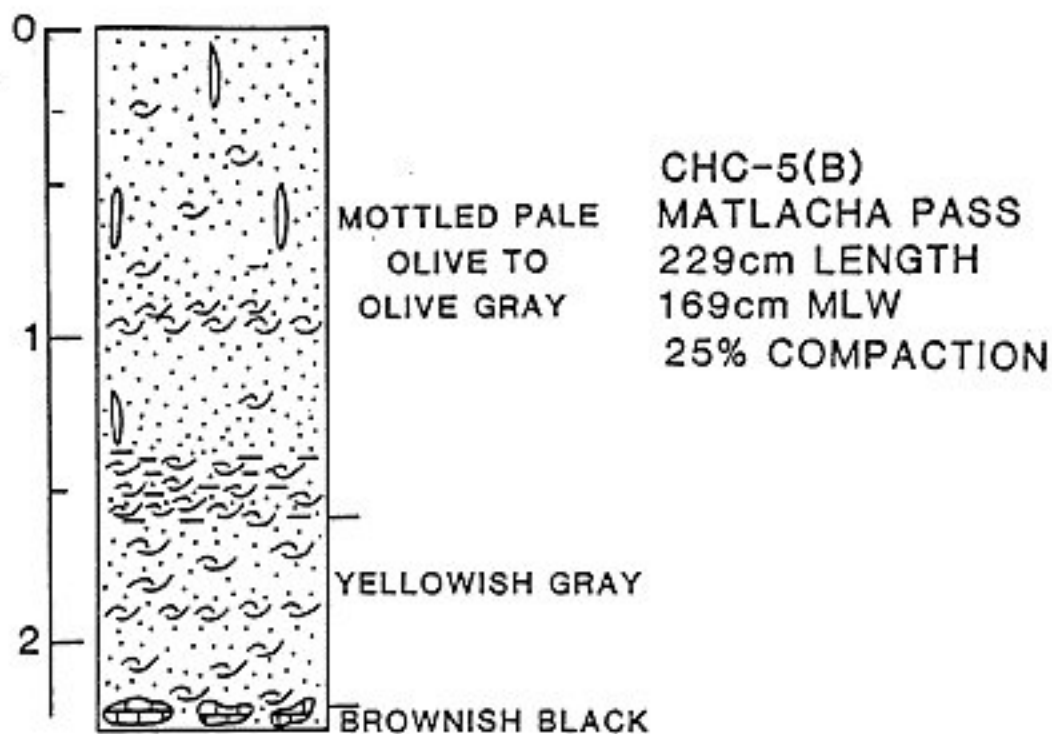
COLOR

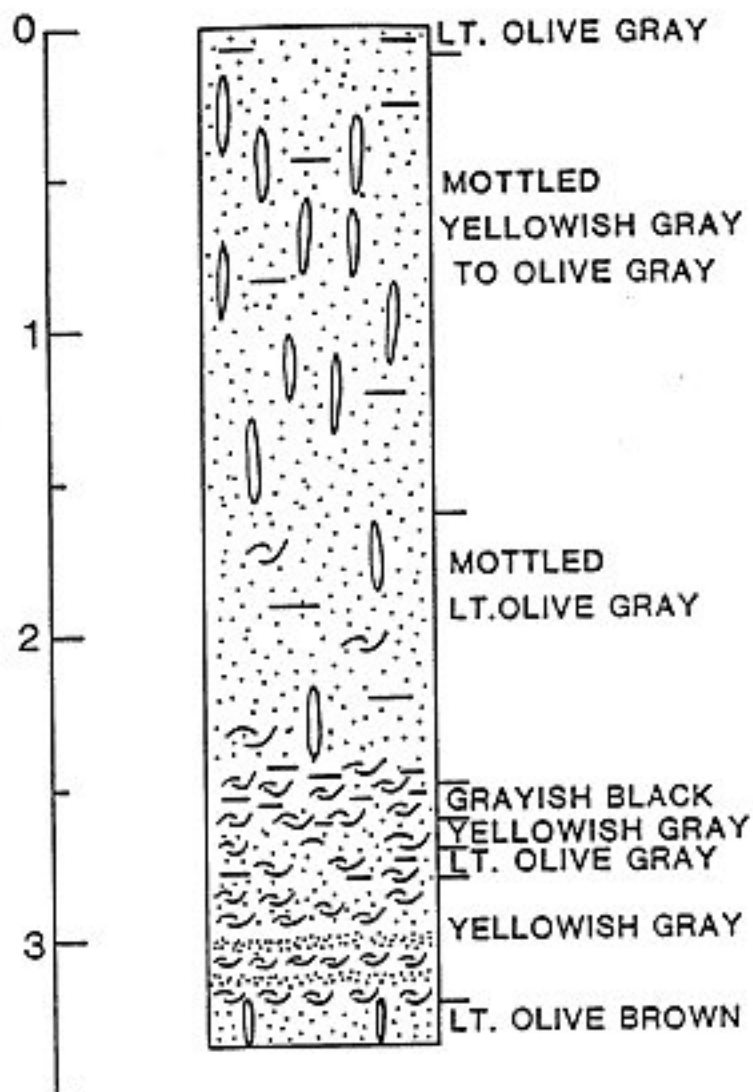




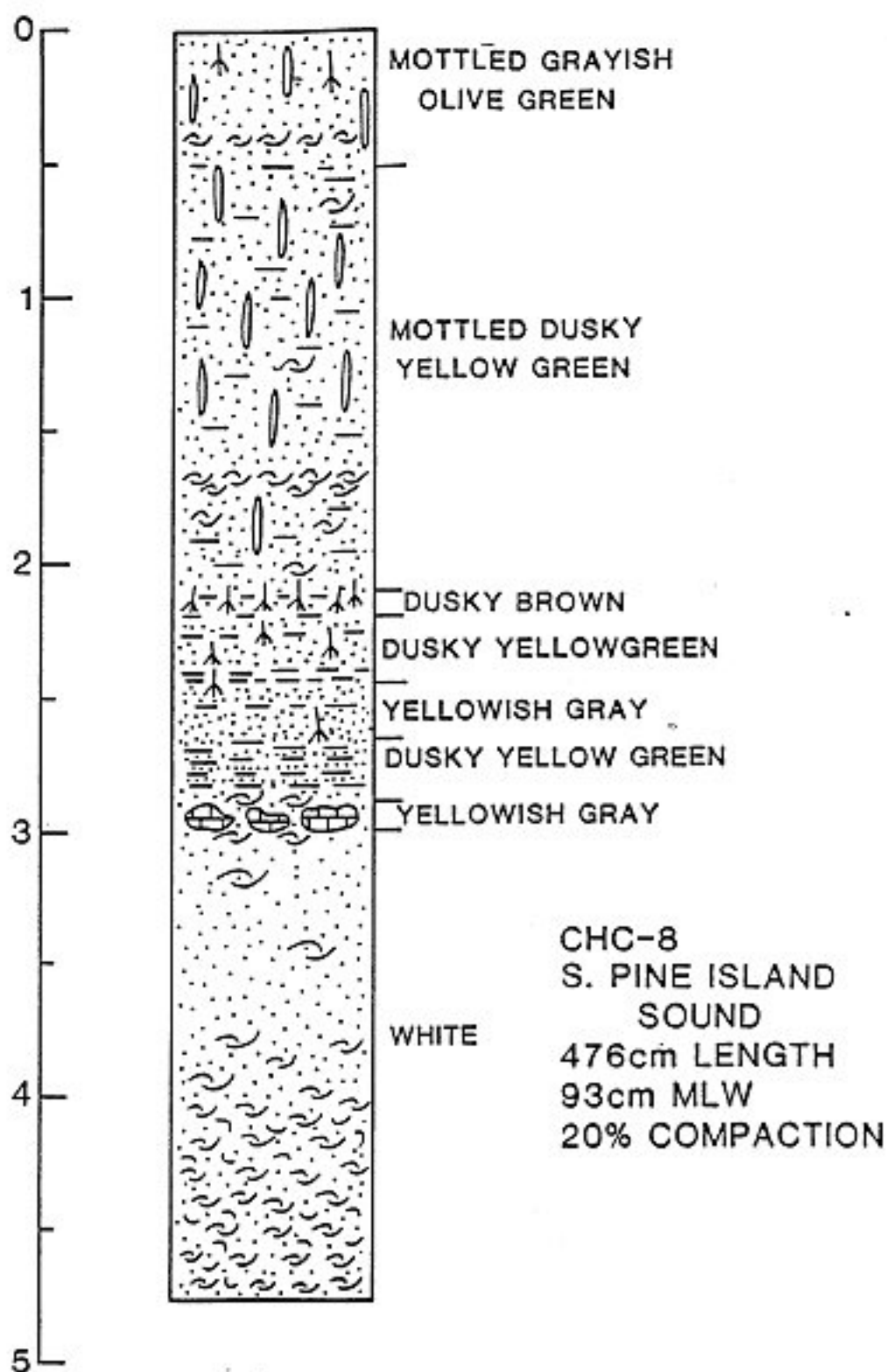


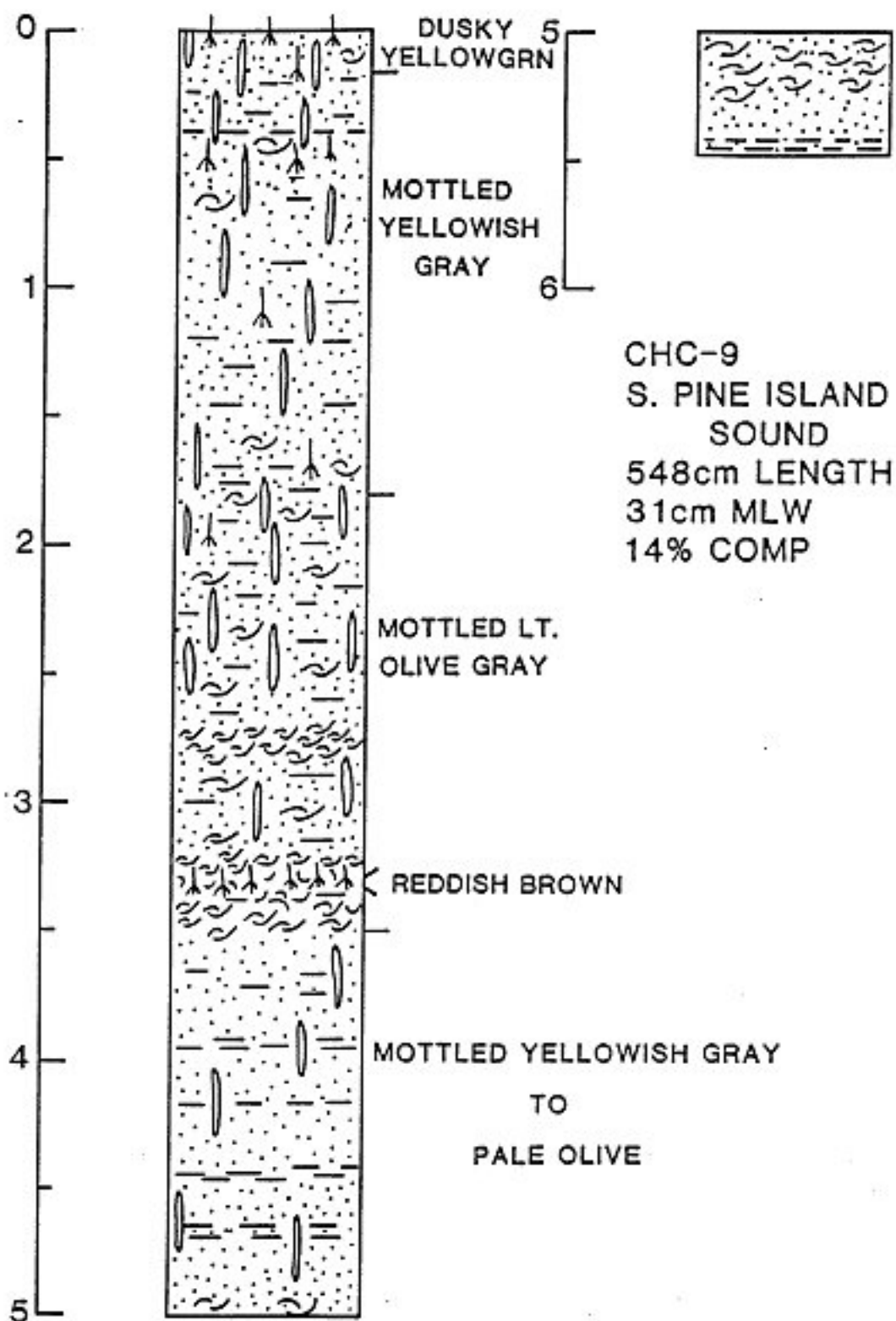


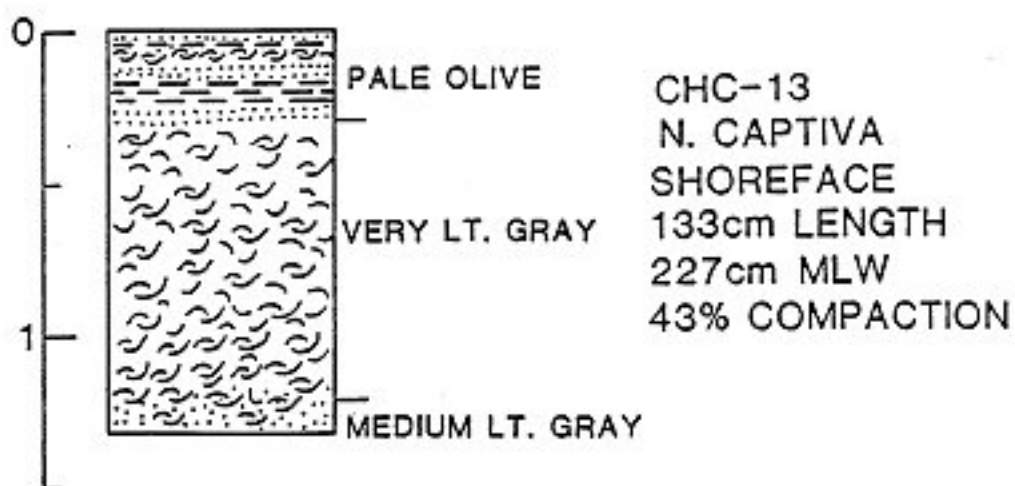
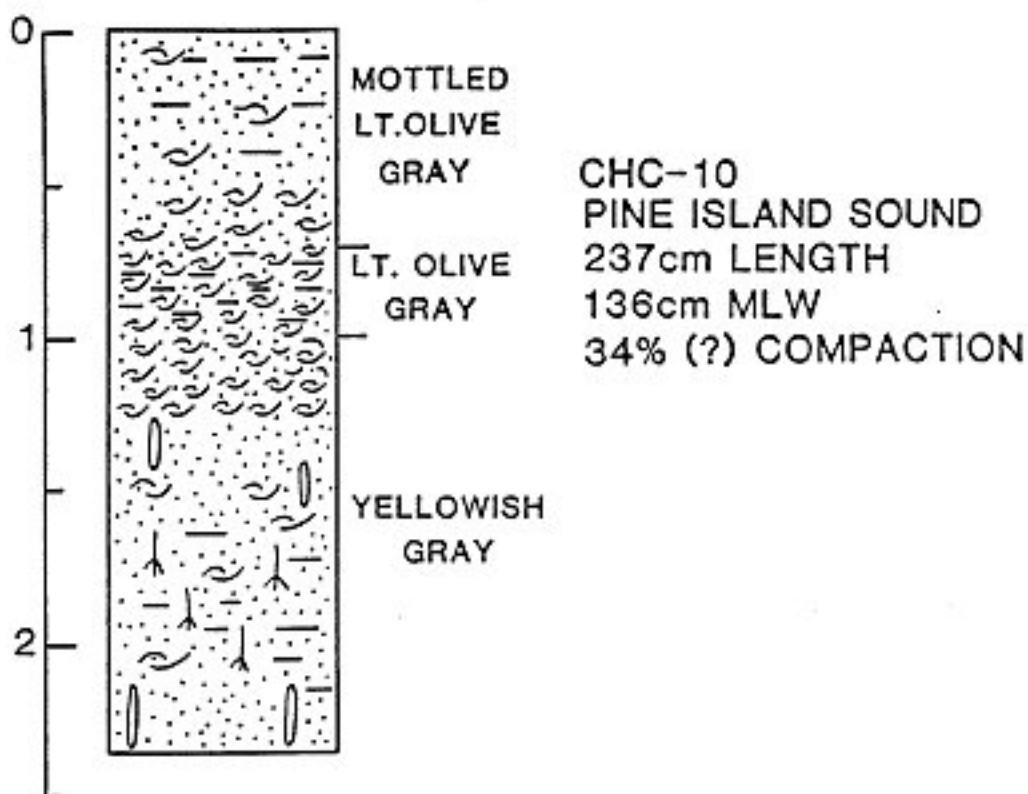


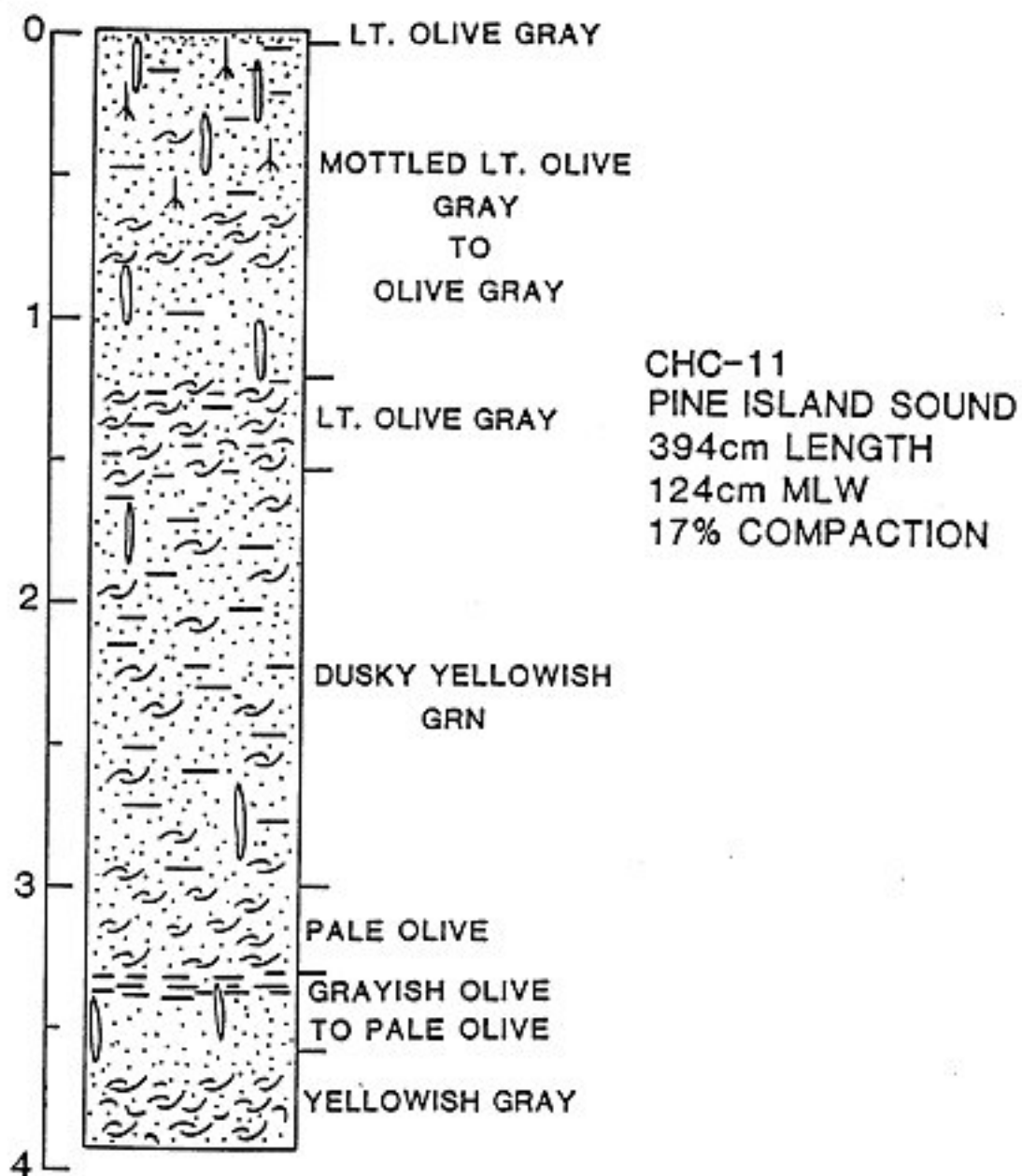


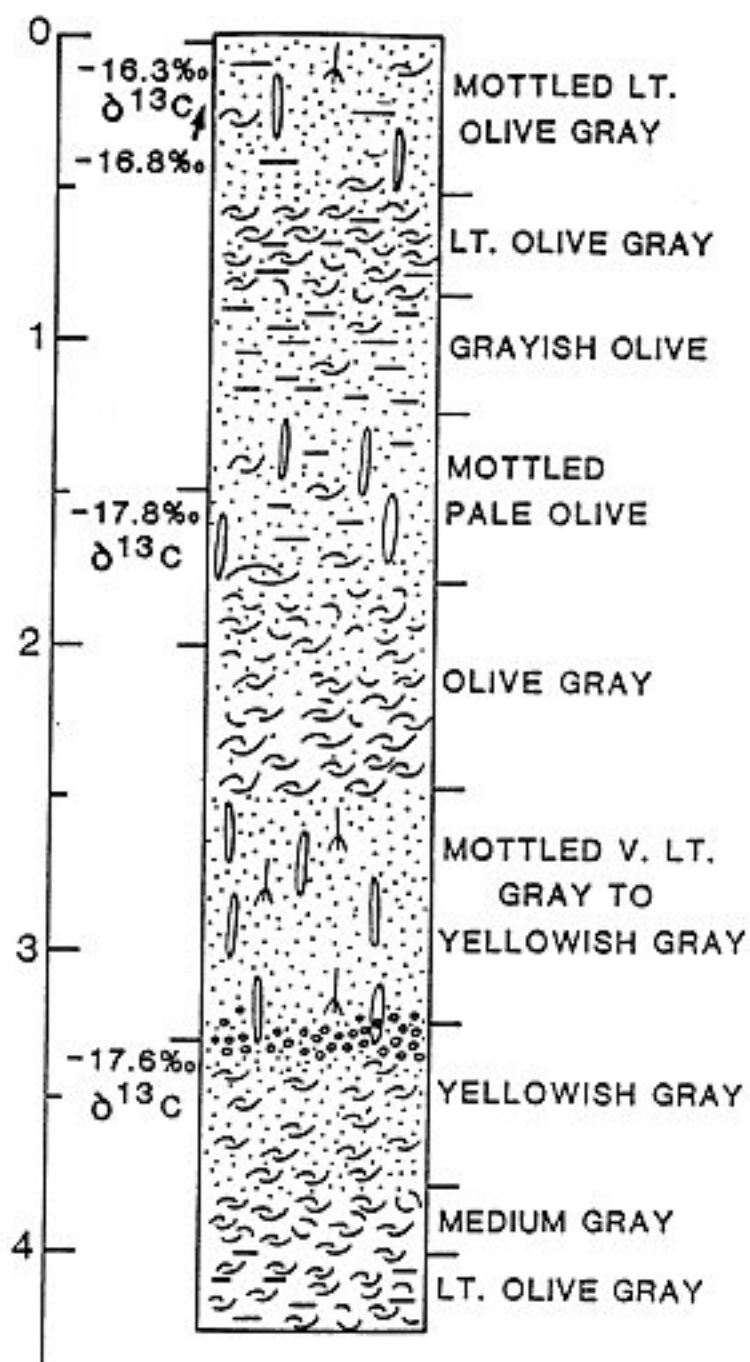
CHC-6
MATLACHA PASS
334cm LENGTH
128cm MLW
25% COMPACTION



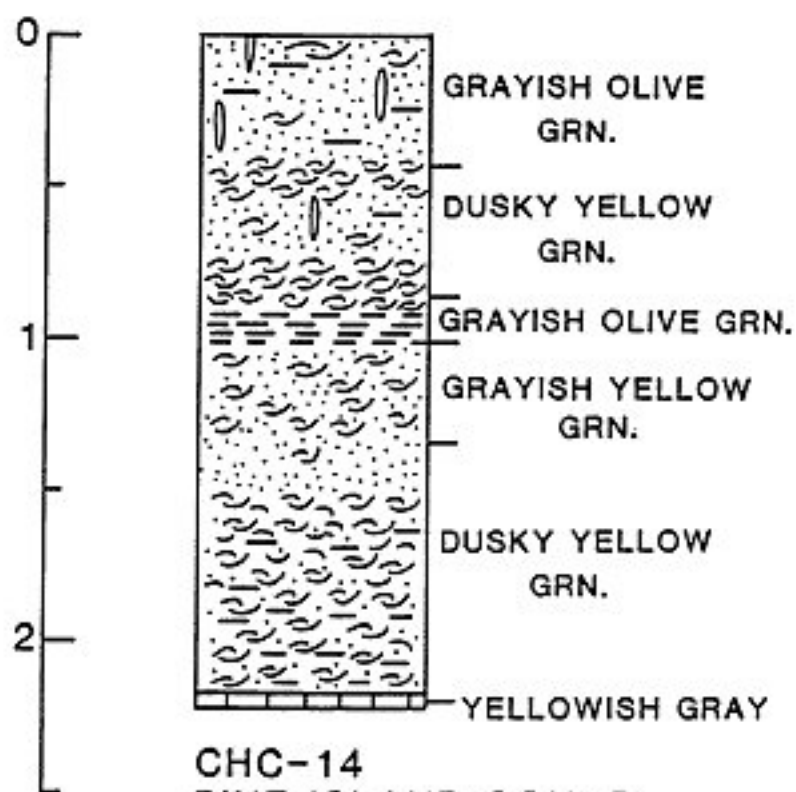




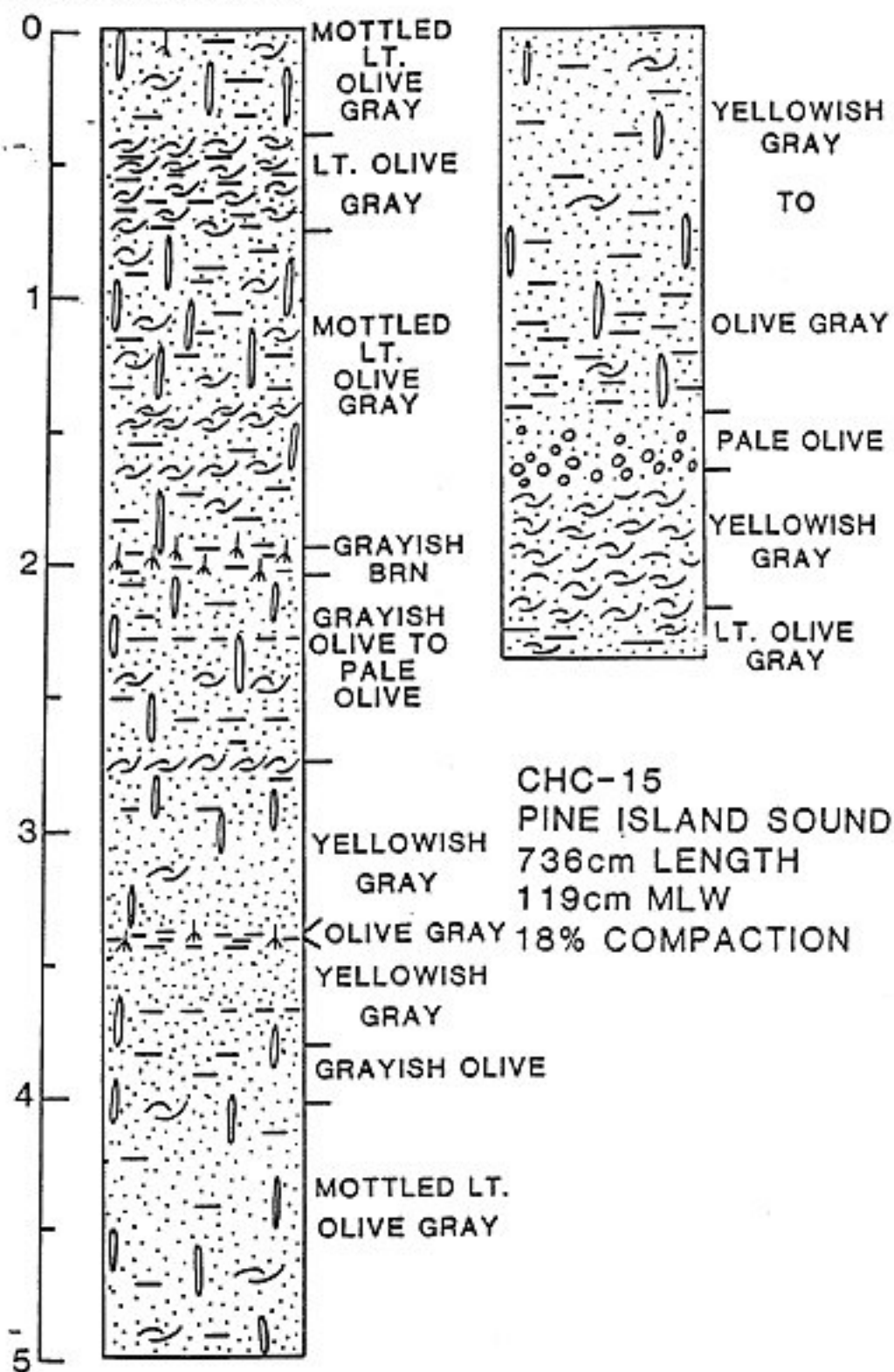


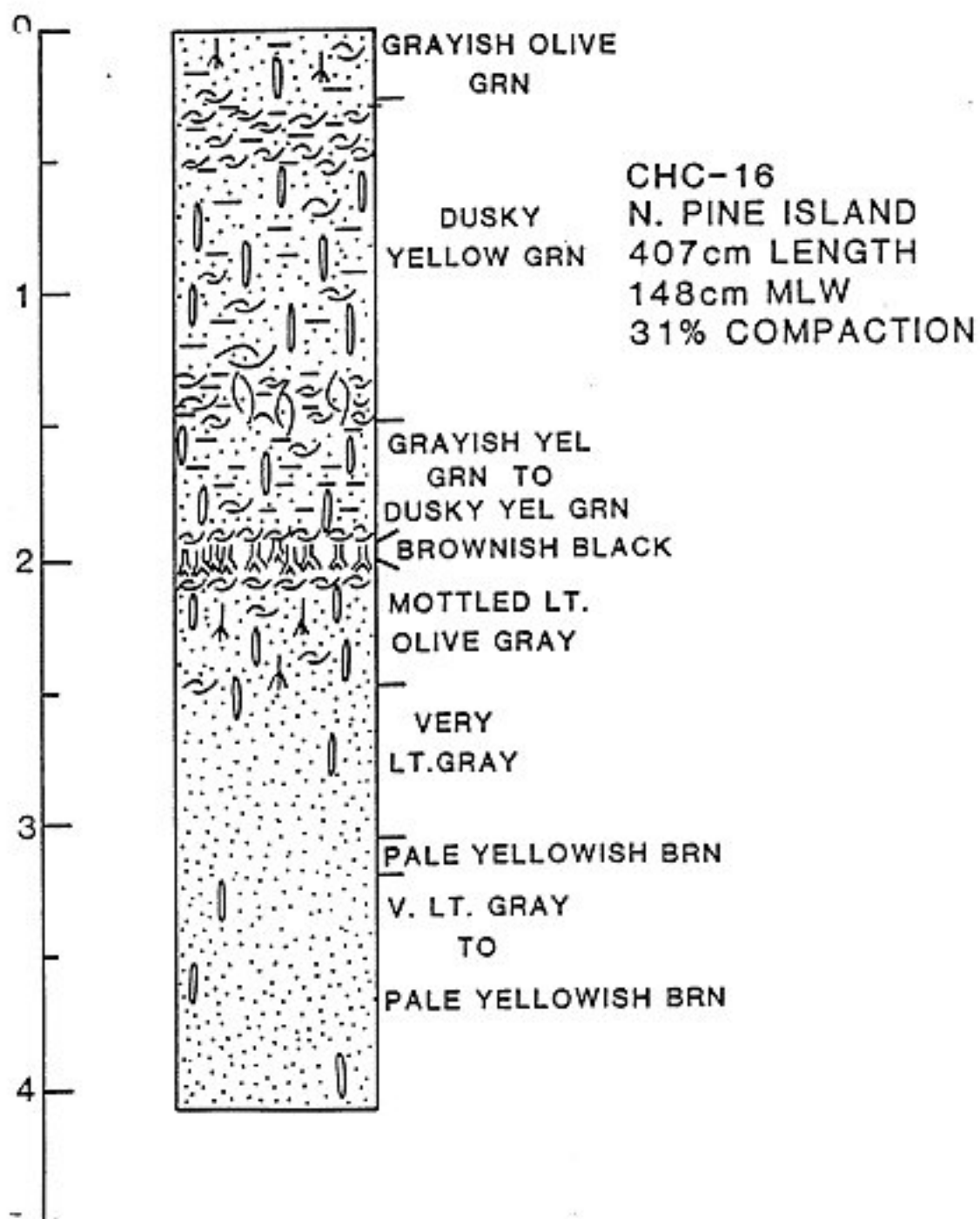


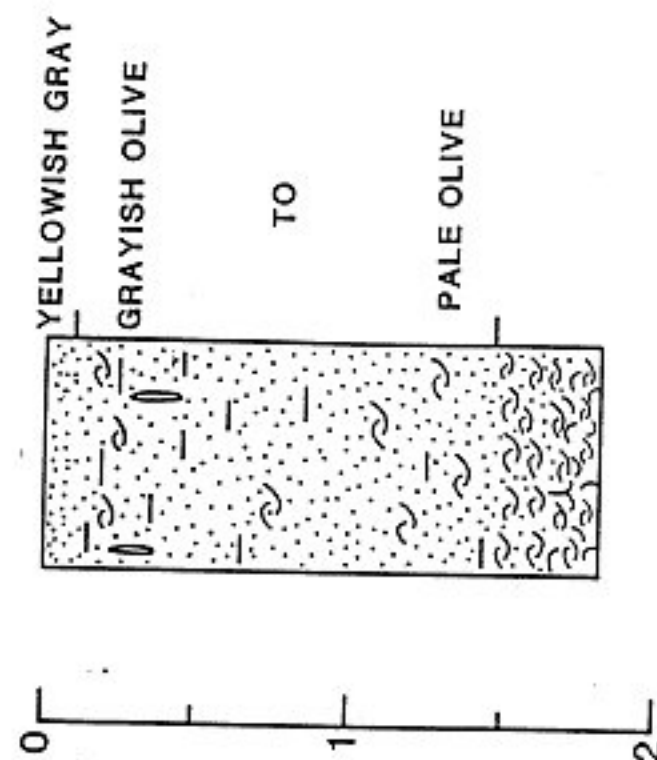
CHC-12
PINE ISLAND SOUND 158cm MLW
425cm LENGTH 29% COMPACTION



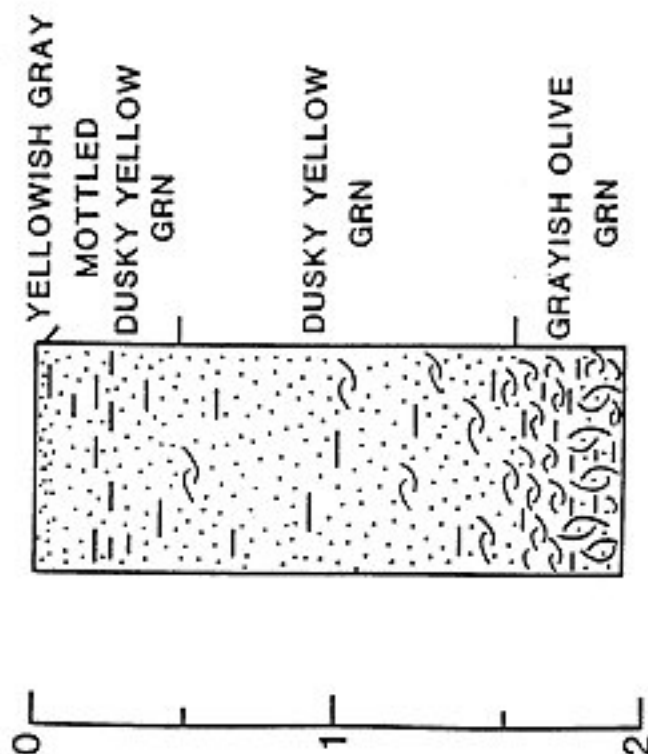
CHC-14
PINE ISLAND SOUND
221cm LENGTH
256cm MLW
18% COMPACTION



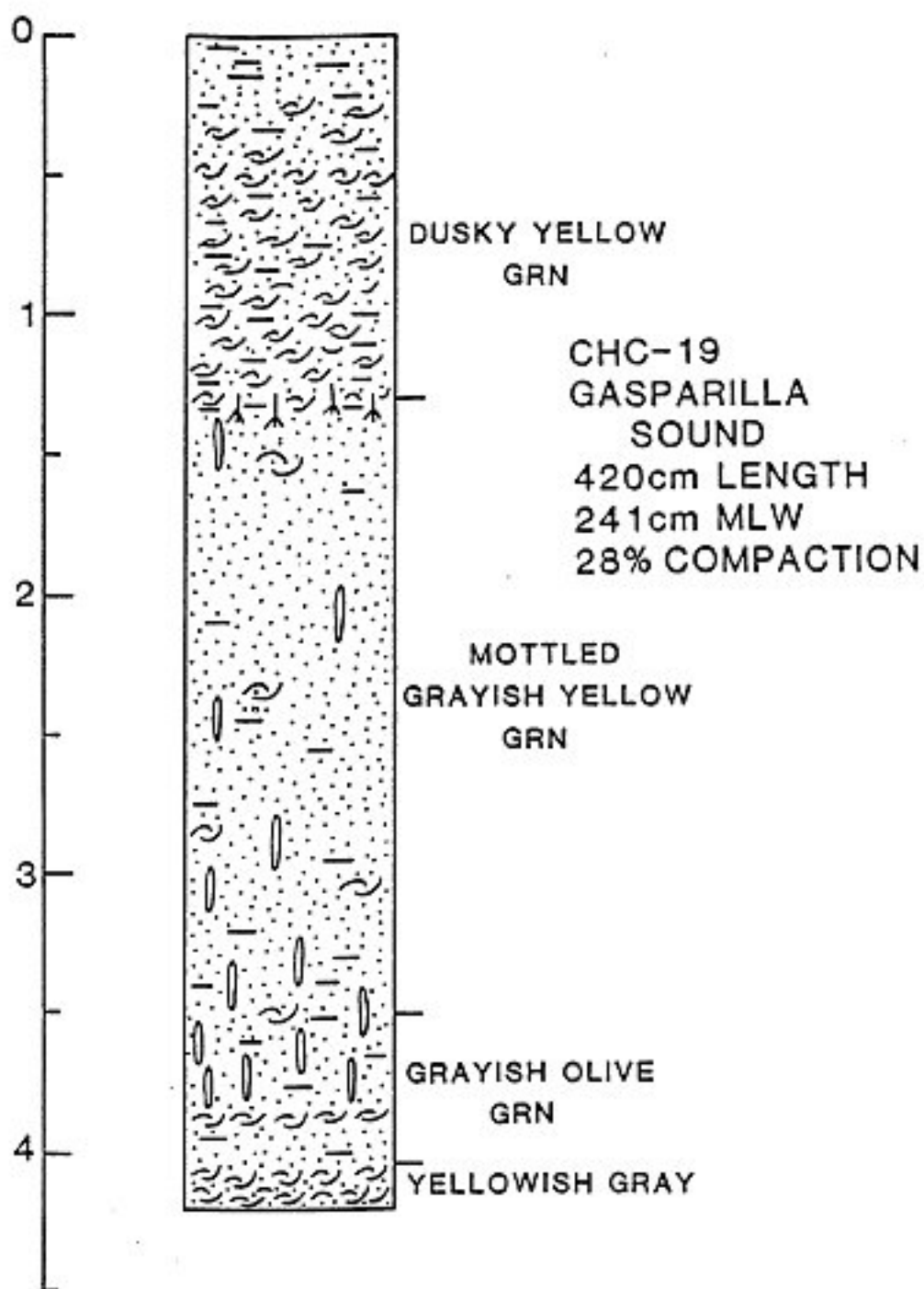


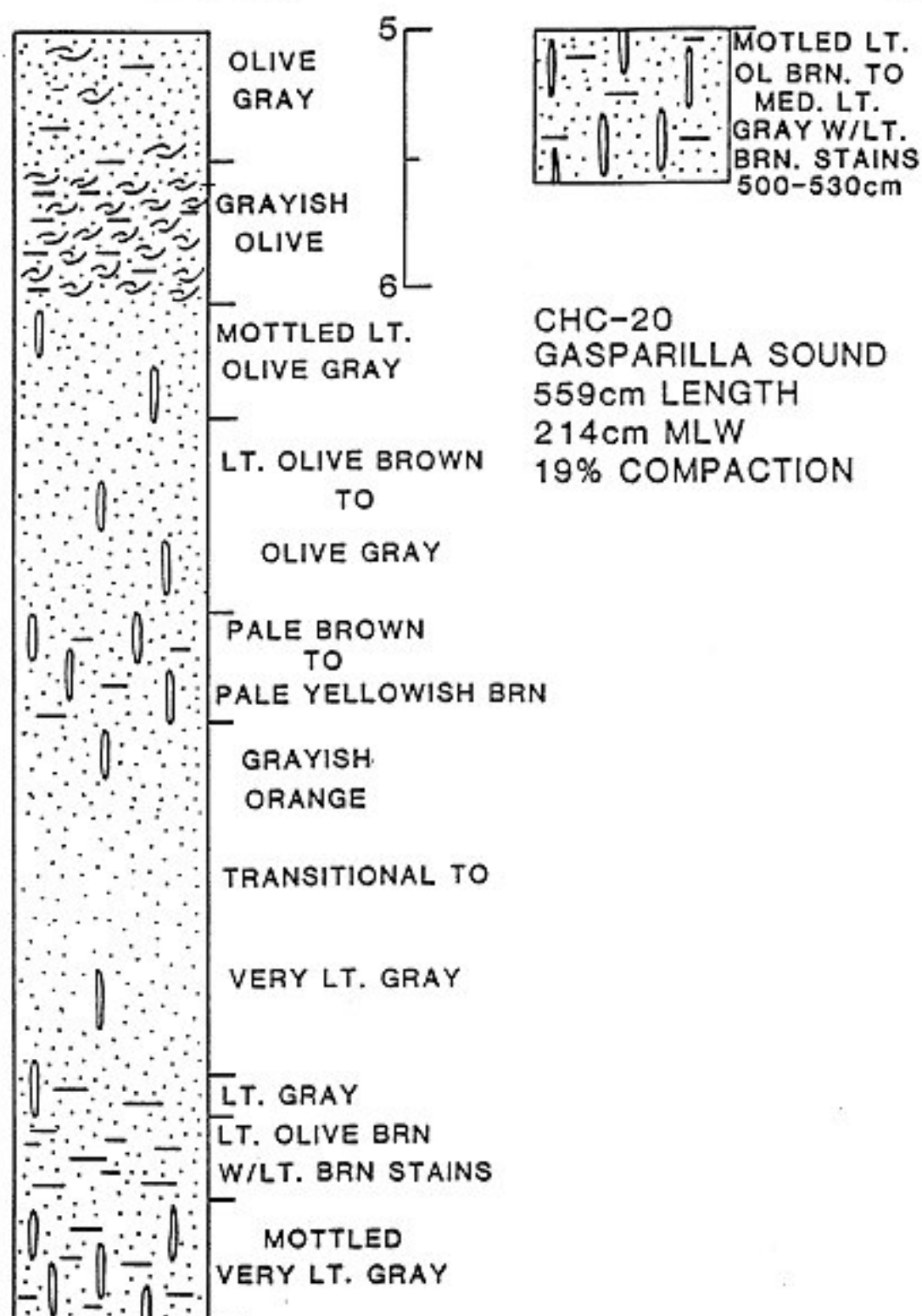


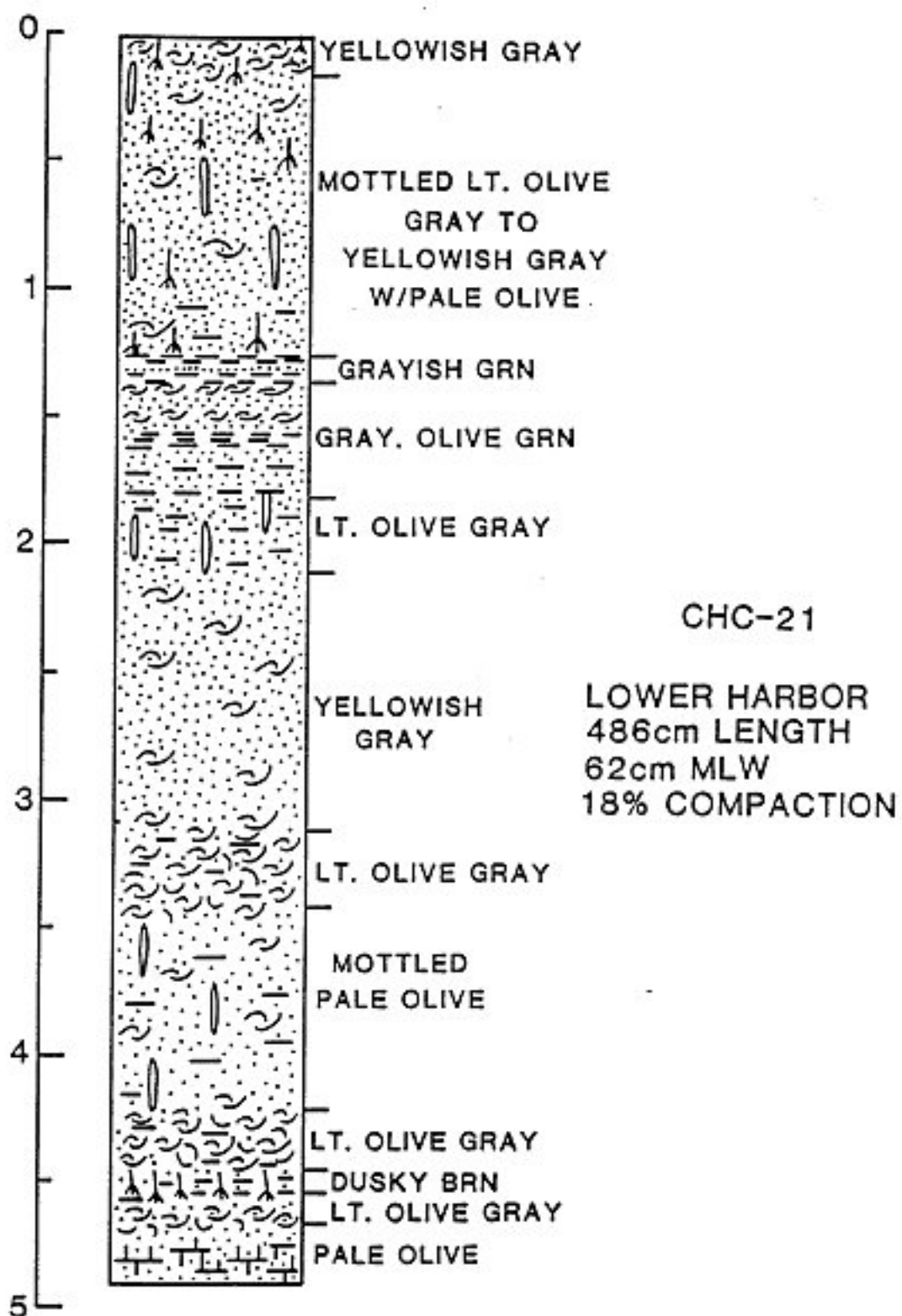
CHC-17
 GASPARILLA SOUND
 182cm LENGTH
 212cm MLW
 53% COMPACTION(?)

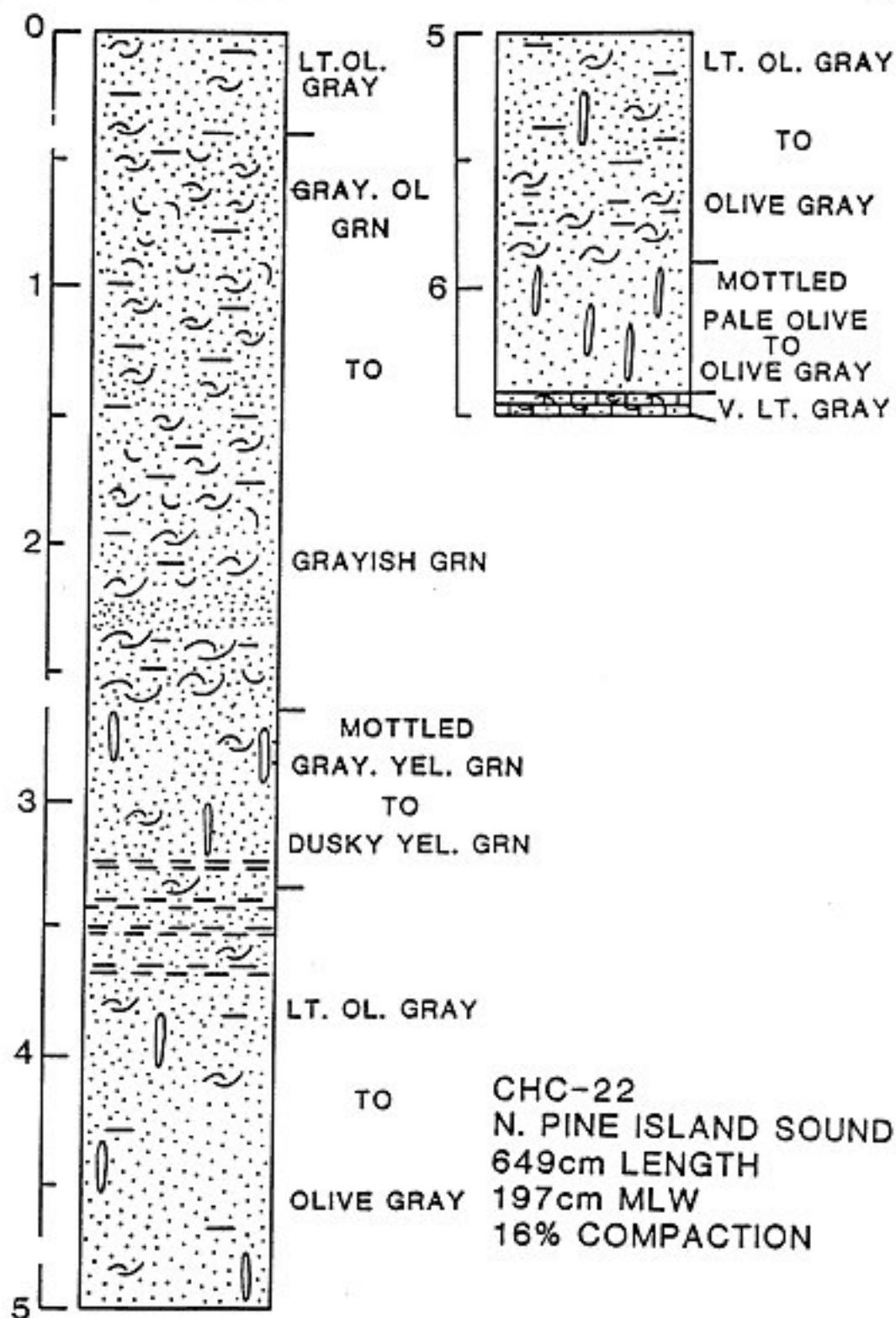


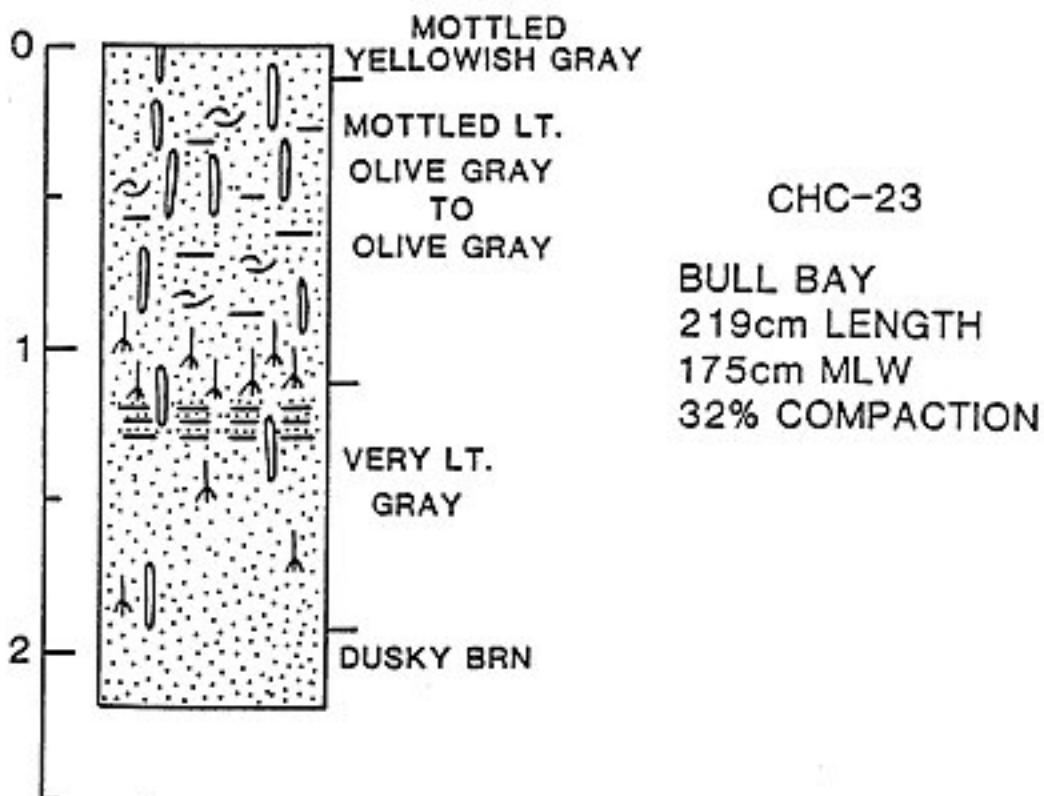
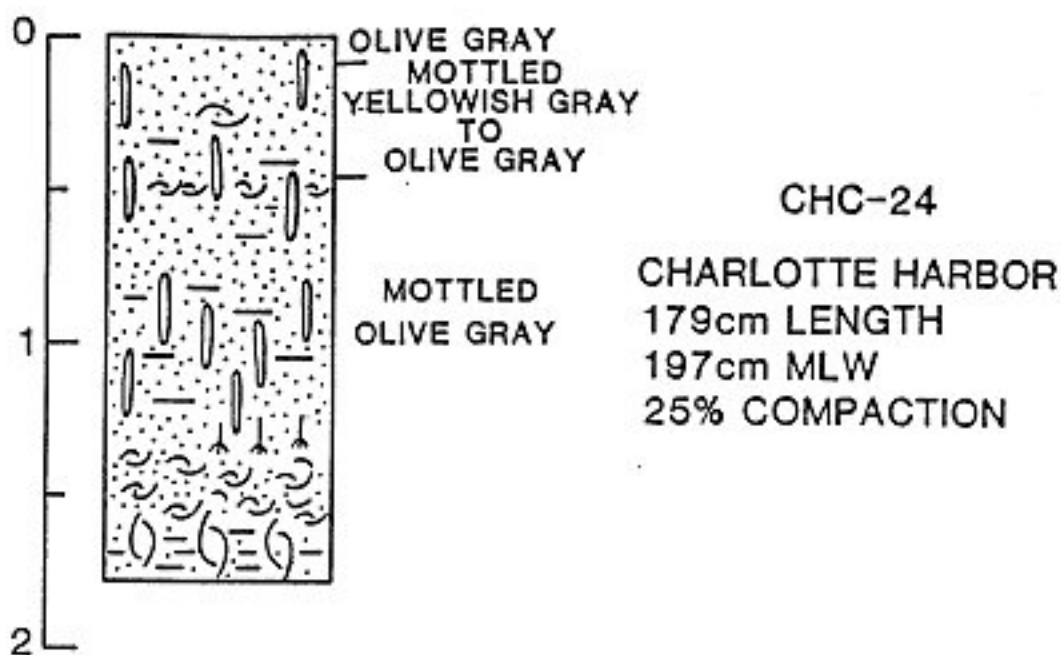
CHC-18
 GASPARILLA SOUND
 194cm LENGTH
 238cm MLW
 32% COMPACTION

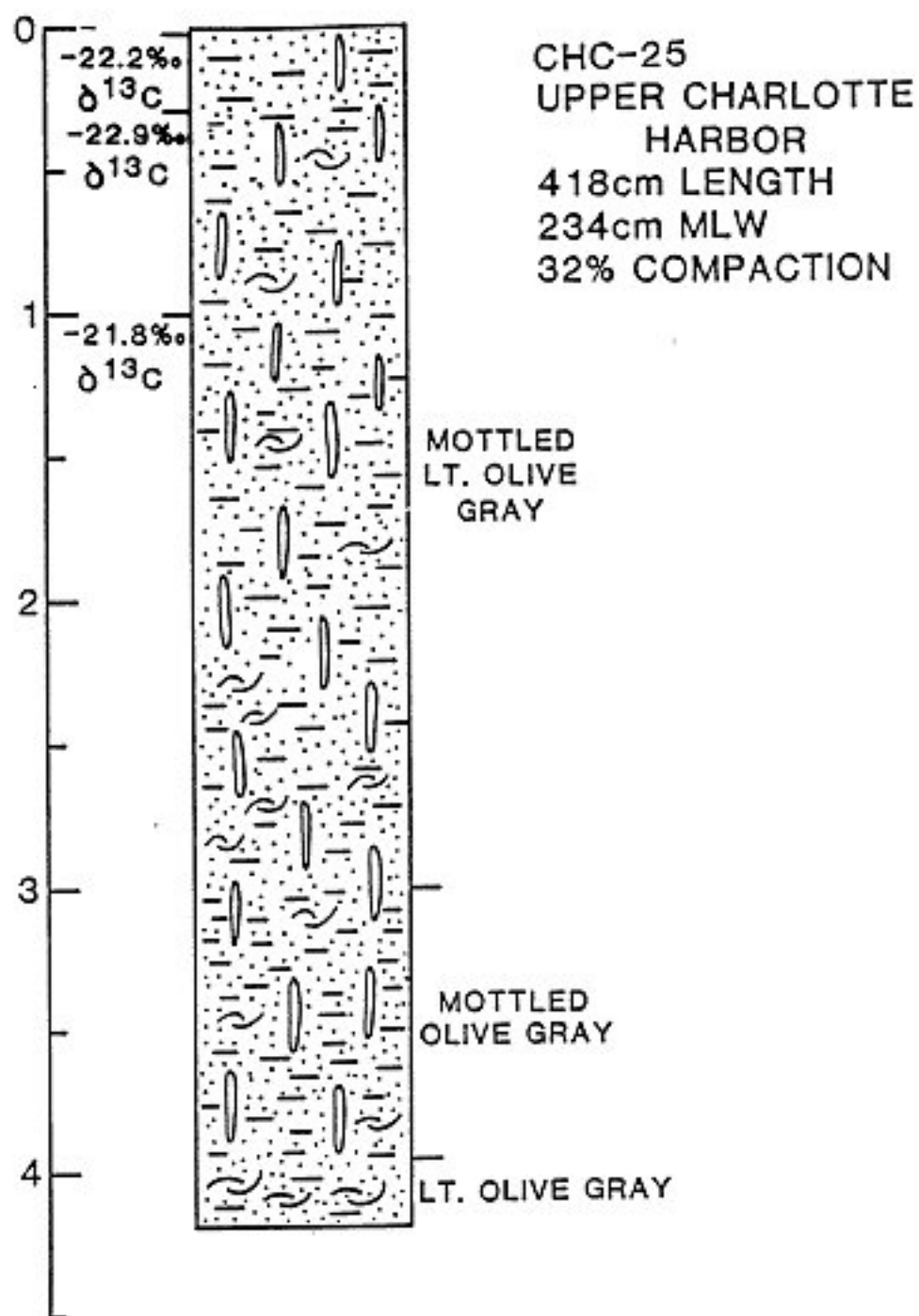


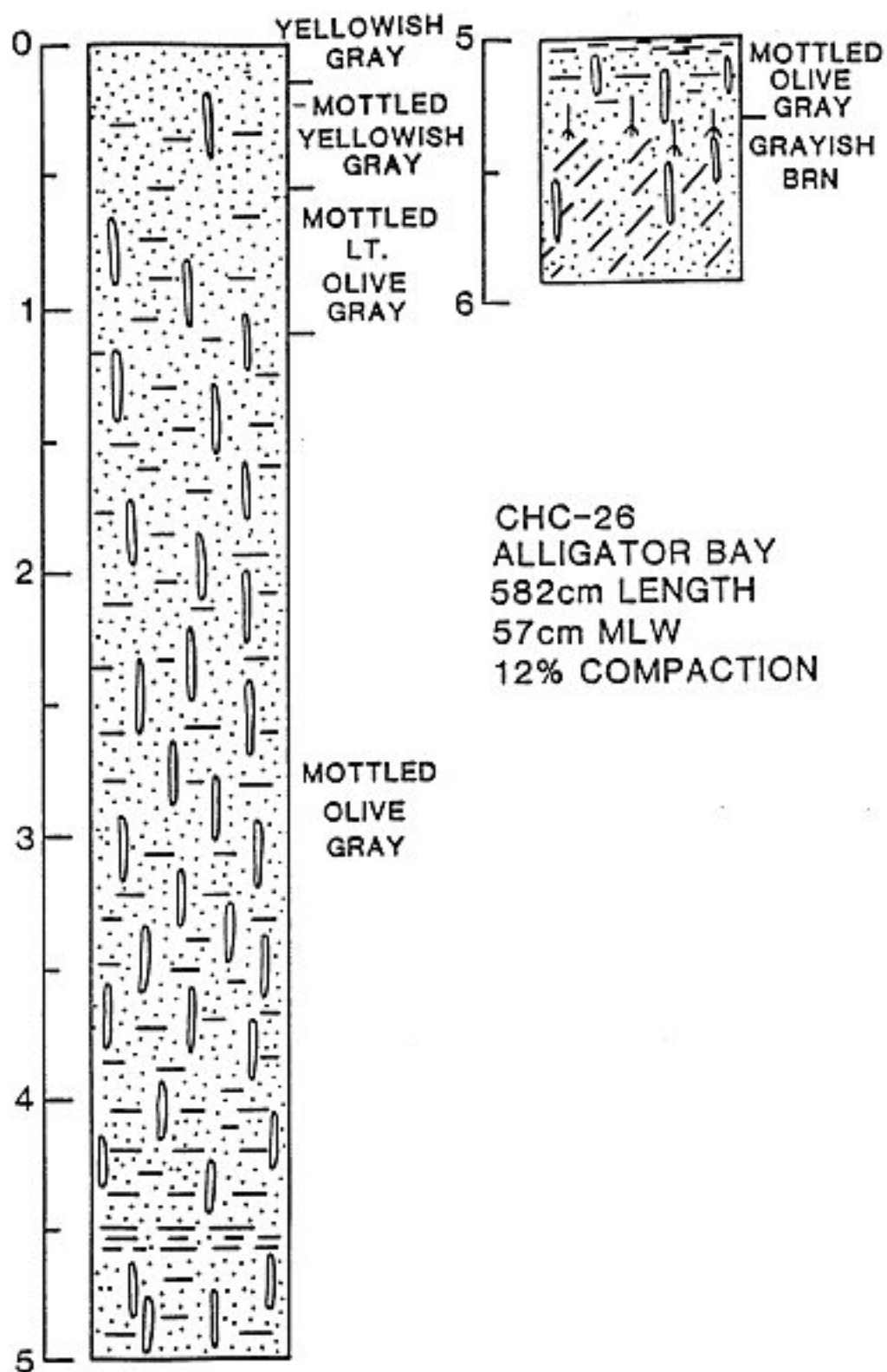


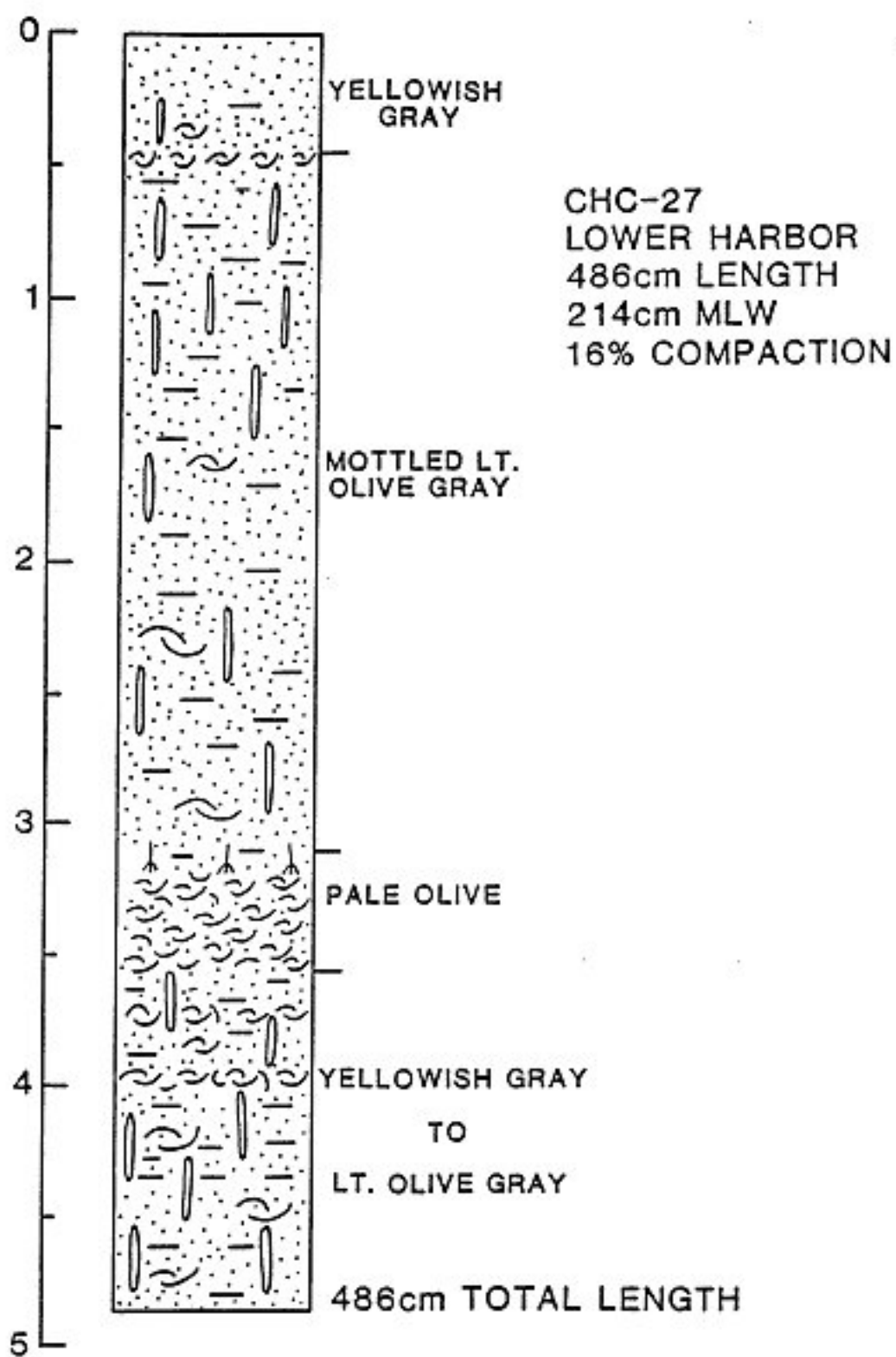




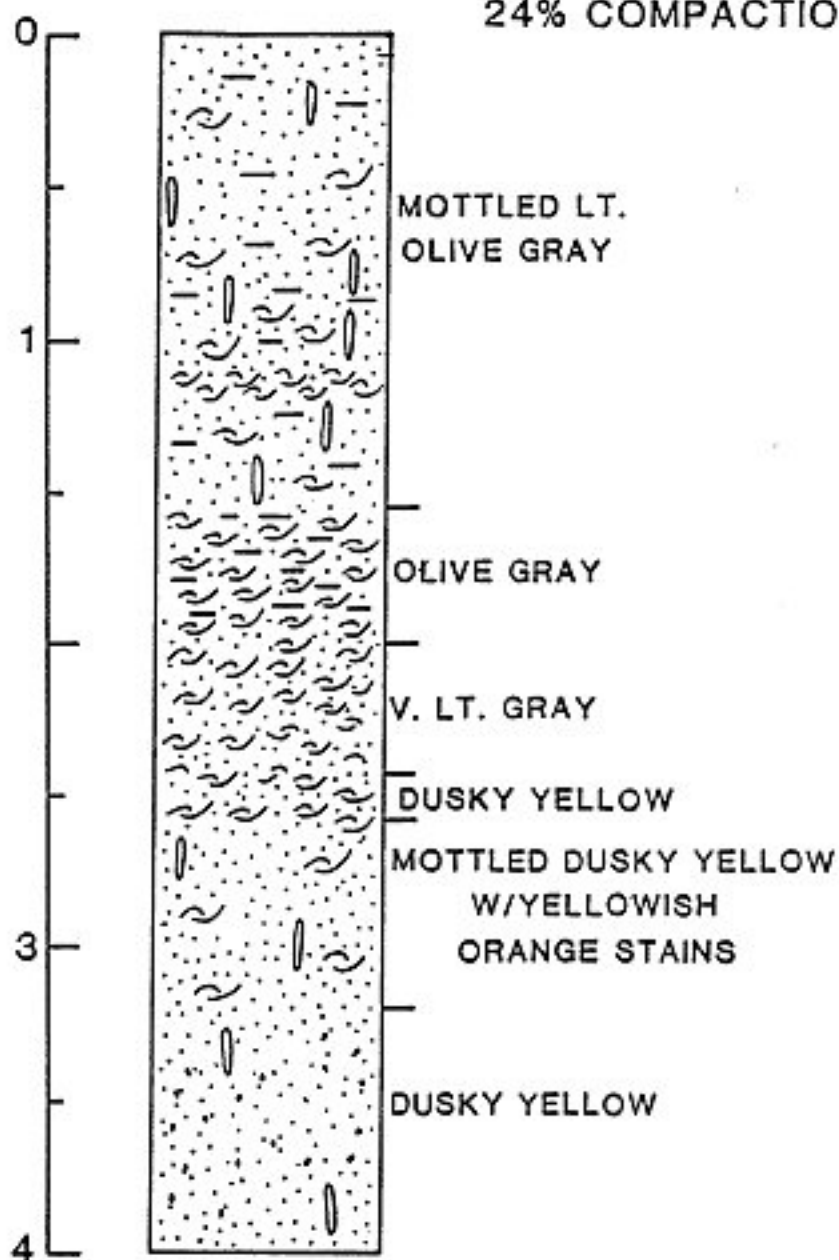


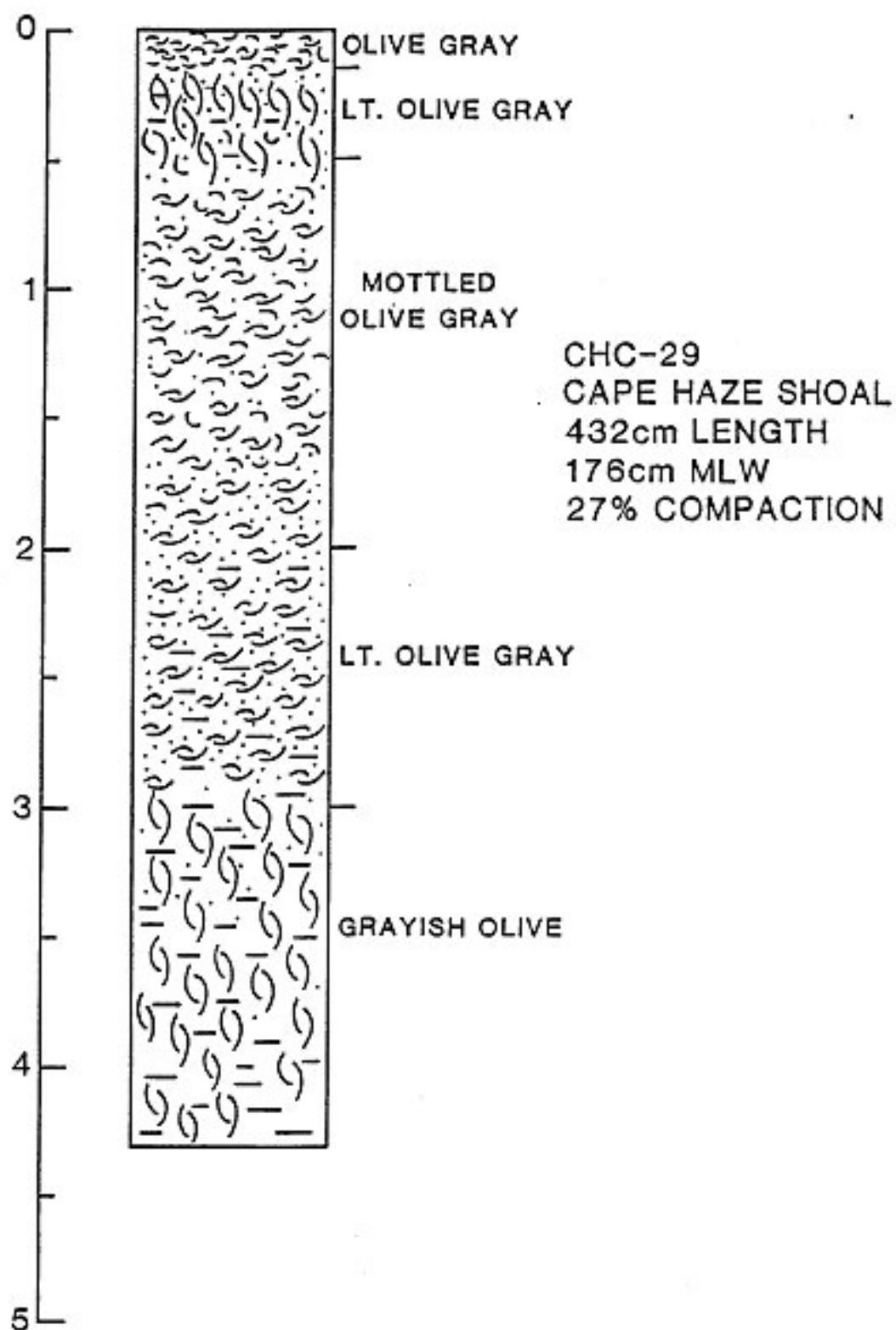


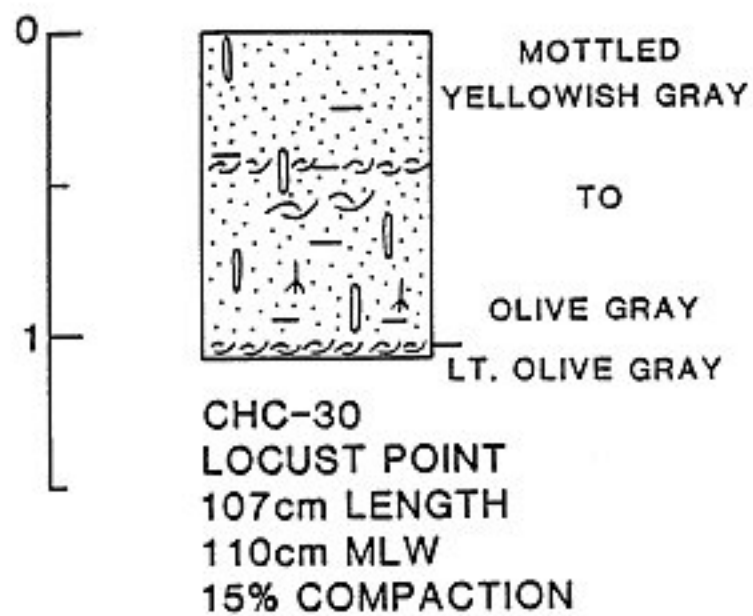


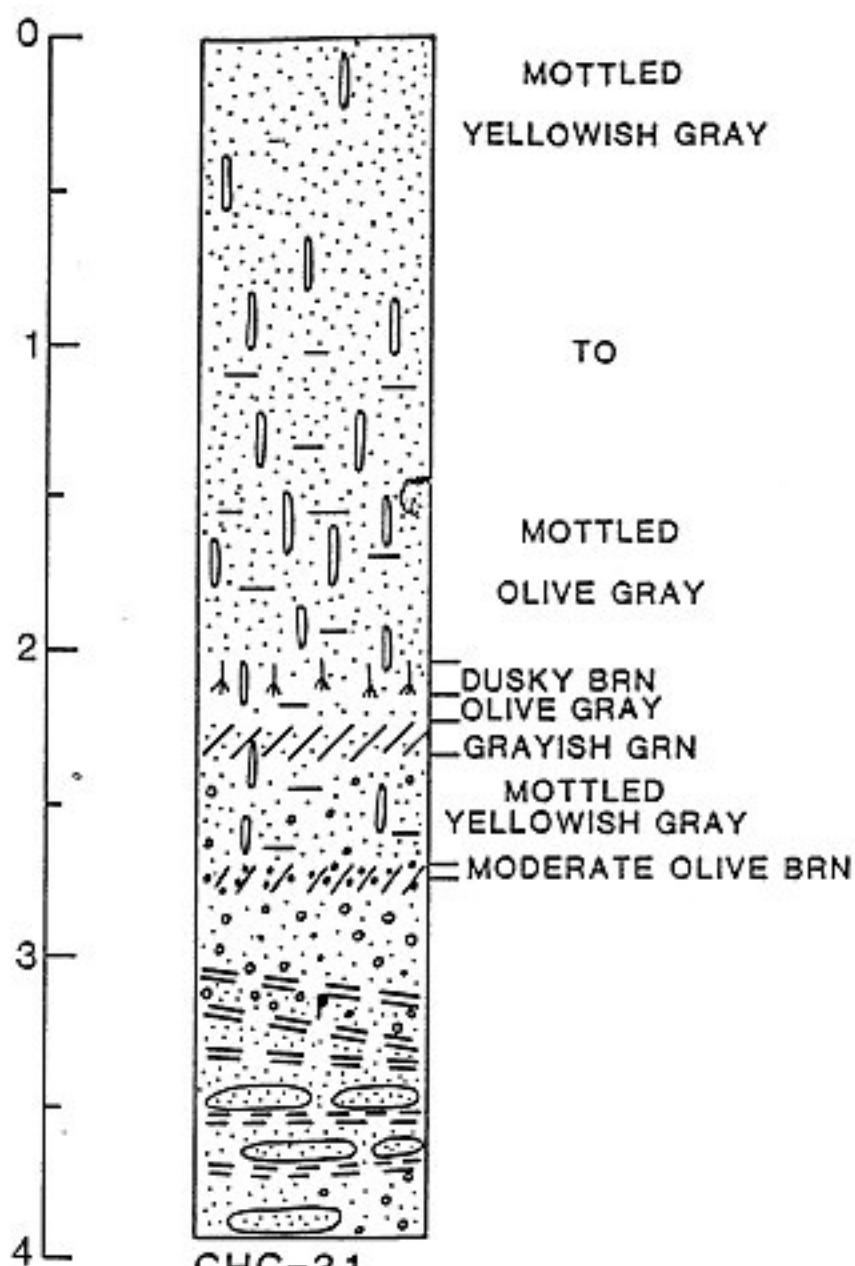


CHC-28
CHARLOTTE HARBOR
401cm LENGTH
233cm MLW
24% COMPACTION

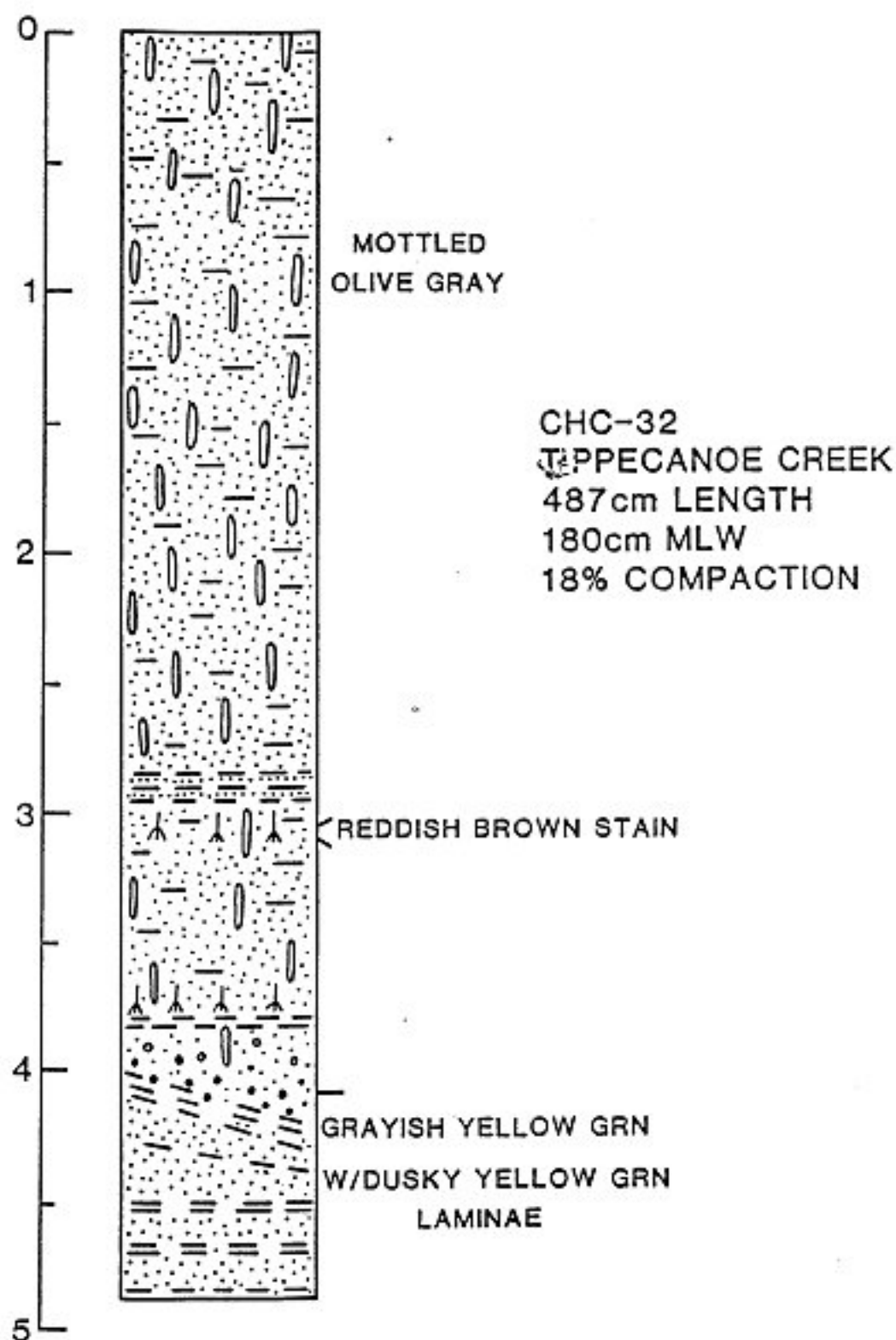


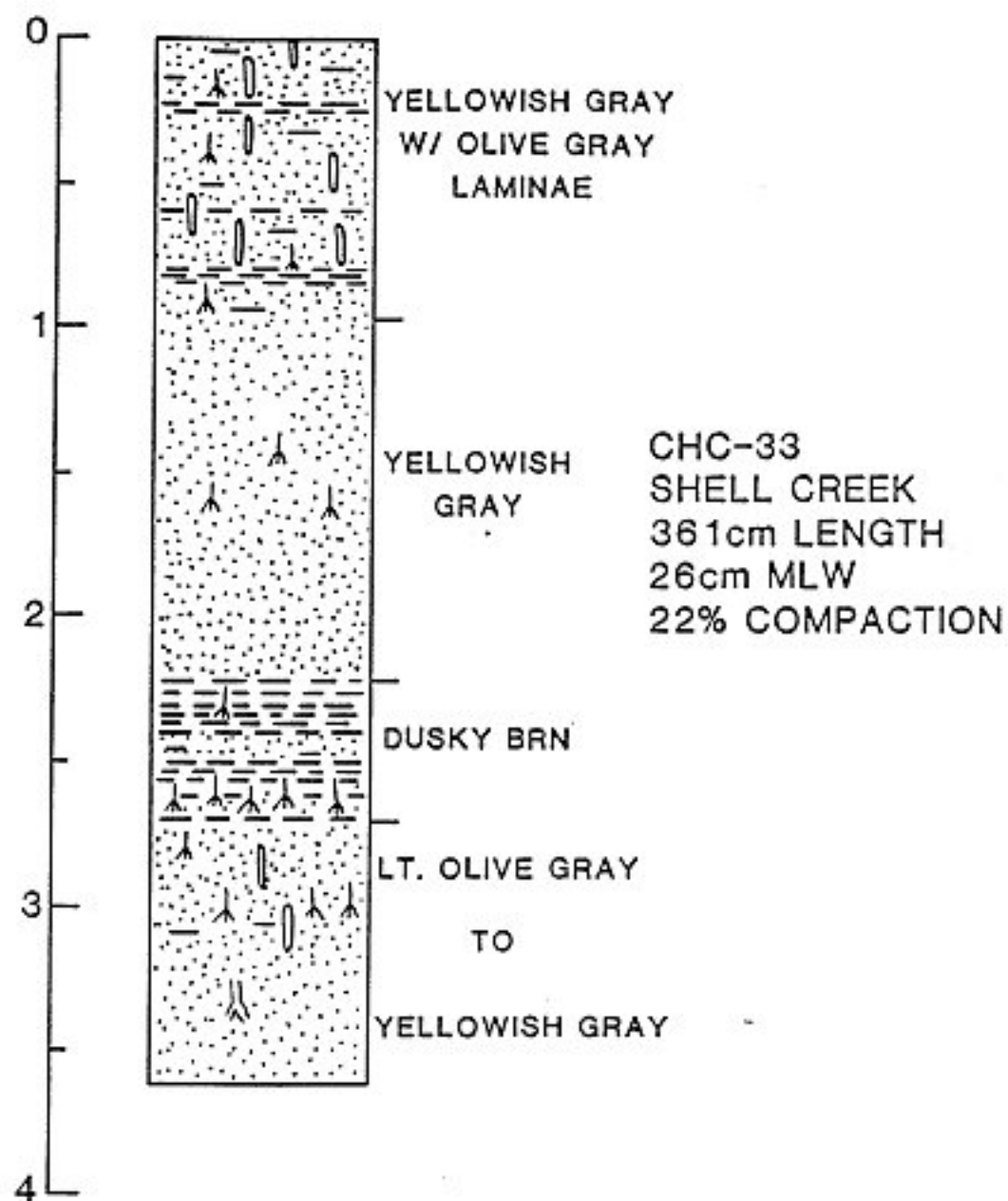


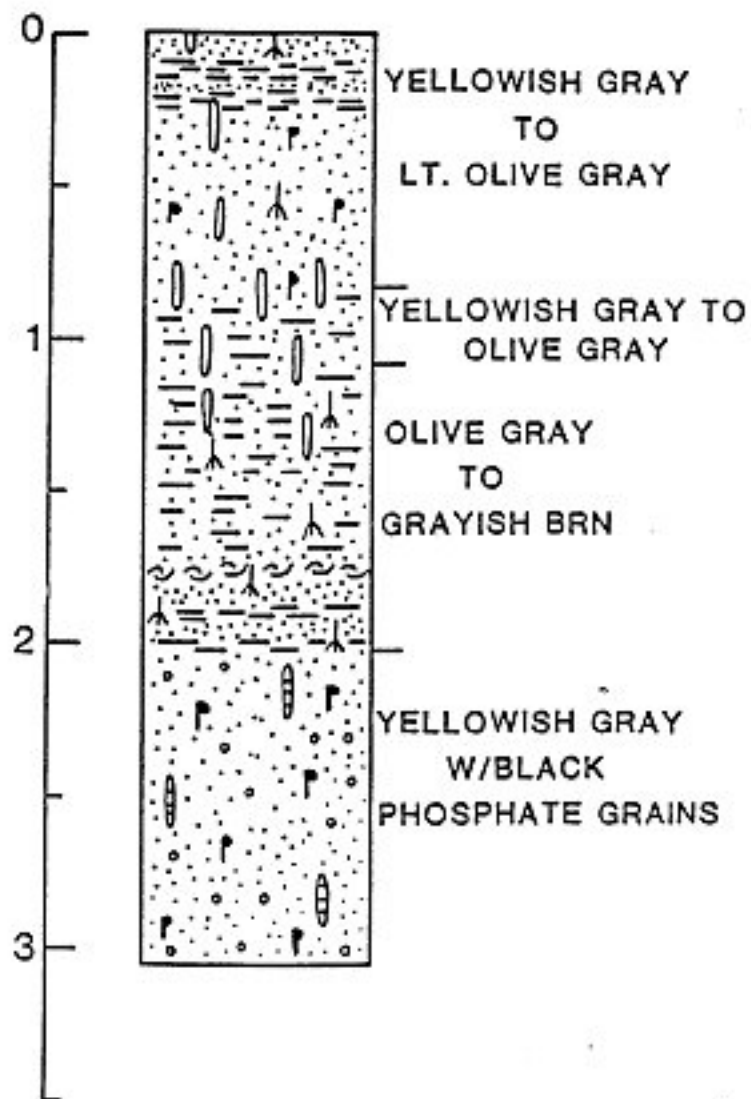




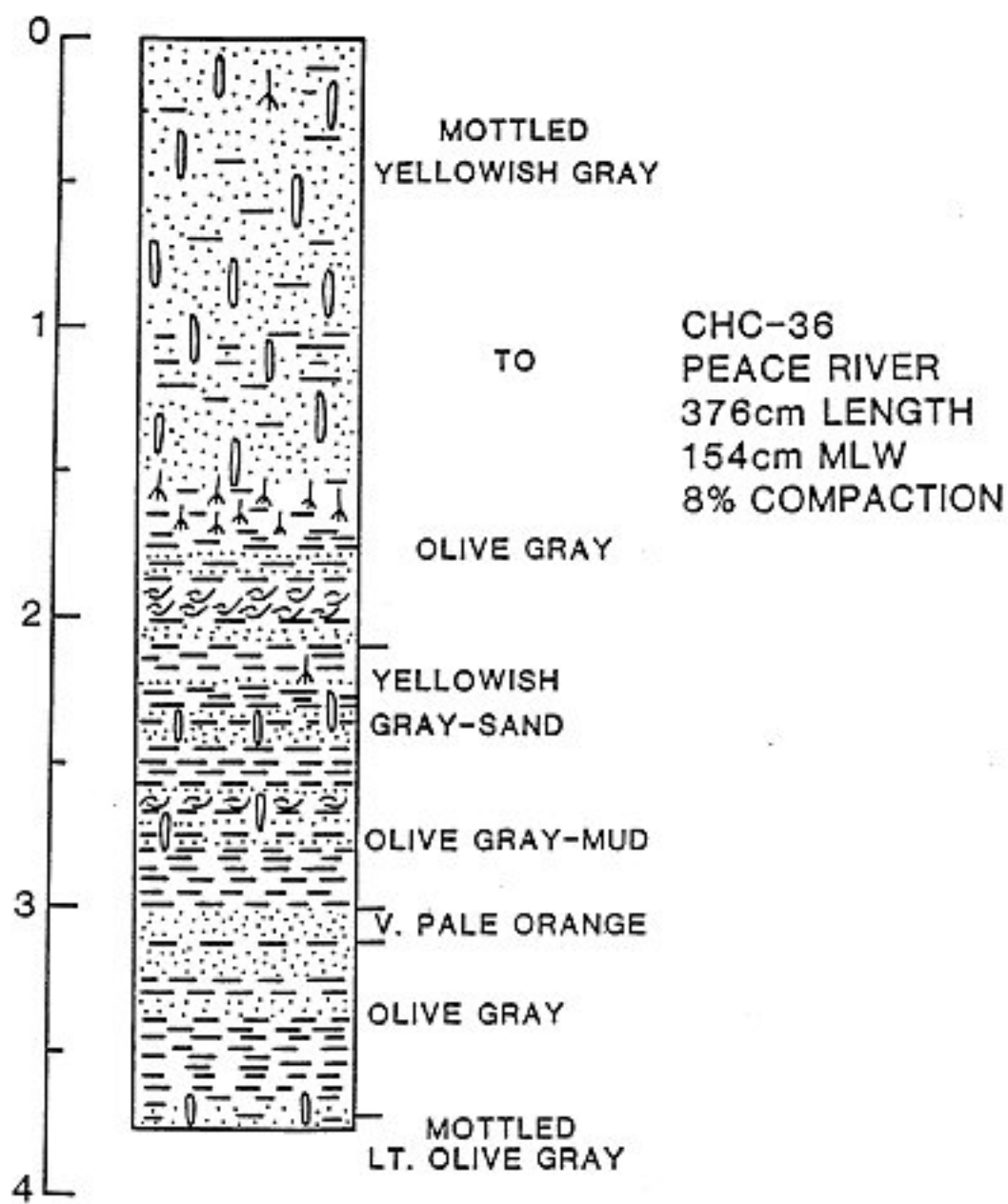
CHC-31
N. CHARLOTTE HARBOR
394cm LENGTH
169cm MLW
33% COMPACTION







CHC-34
SHELL CREEK
306cm LENGTH
86cm MLW
34% COMPACTION



CHC-37
PEACE RIVER
346cm LENGTH
59cm MLW
22% COMPACTION

