THE UNIVERSITY OF MANITOBA

SEDIMENTOLOGY OF PLEISTOCENE CARBONATES FROM BIG PINE KEY, FLORIDA

by

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MASTER OF SCIENCE

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TABLE OF CONTENTS

LIST OF FIGURES

LIST OF FIGURES (Cont'd)

LIST OF TABLES

LIST OF PLATES

SEDIMENTOLOGY OF PLEISTOCENE CARBONATES FROM BIG PINE KEY, FLORIDA

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ABSTRACT

Cores and outcrop samples of Late Pleistocene carbonates from Big Pine Key were studied in detail in order to describe and interpret their stratigraphy, primary depositional facies, and submarine and subaerial diagenesis. The boreholes penetrated three stratigraphic units separated by distinct paleoexposure surface zones. From youngest to oldest these are the Q5, Q4, and Q3 units of Perkins, 1977.

Lithologically, the Q3 and Q4 units consist of a peloid-bioclast packstone to grainstone facies. The Q5 unit can be subdivided into three intergradational facies. From northwest to southeast across Big Pine Key, an ooid grainstone facies grades into a transitional facies, which, in turn, grades into a peloid-bioclast packstone to grainstone facies. This latter facies, which also constitutes much of the Q4 and Q3 units, is considered to be Key Largo Limestone. The grainstone and transitional facies of the Q5 unit are commonly referred to as the Miami Limestone. The ooid grainstone represents deposition in a high energy environment whereas the other facies indicate low energy. The presence of ooids in the Q4 unit suggests that previously undocumented shoals were developing at that time.

Submarine cementation in the Q5 unit of Big Pine Key is ubiquitous

and dominantly occurs in intraparticle spaces. Grain micritization, especially of ooids, is common and is primarily the result of two types of algal endoliths: a spherical to irregular form and a filamentous form.

Two types of surficial features related to subaerial exposure are recognized: a laminated crust and a micrite cement zone. Important components of the subaerial exposure surface zone include micrite cement, calcified filaments, random needle fibres, peloids, and fungal microborings. The calcified filaments are thought to originate from root hairs, and occur mainly as chasmoliths. Significant primary textural features are obliterated where endolithic forms are abundant.

Freshwater alteration of bioclasts and ooids is variable, but is typically less pronounced in the Q5 unit than in the older units. This is a reflection of the greater length of time that the older units have been subaerially exposed. The distribution of spar cement is patchy. The development of vug and channel porosity is closely associated with areas of minor cementation in grainstones. Meniscus pore-rounded cements also occur. Cross-bedding or burrowing controls the development of some of the vug and channel porosity, but the origin for the majority of these pores appears to be random dissolution. In mud-rich sediments, similar vug and channel pores develop in areas of intensive leaching of aragonitic mud. The nature of the diagenetic fabrics suggests that the vadose zone was the principle environment of diagenesis for the Big Pine Key carbonates. Effects attributable to the relatively recent incursion of the freshwater phreatic lenses are minor and localized.

The aragonitic Q5 and the calcitic Q4 and Q3 unit sediments have attained diagenetic grades IV and V of Land et al. (1967) and clearly demonstrate the path of mineralogical stabilization. The total porosity of the Q units is similar. However, porosity in the Q4 and Q3 units is mainly secondary in nature, whereas both primary and secondary porosity are common within the Q5 unit. The relatively higher secondary porosity of mud-rich sediments in the Q5 unit compared with grainstones suggests that the nature of the sediment itself is an important control on the evolution of porosity in early diagenesis. Accompanying this tendency toward rapid attainment of secondary pore fabrics in mud-rich sediments is a higher degree of mineralogical stabilization.

CHAPTER 1 - INTRODUCTION

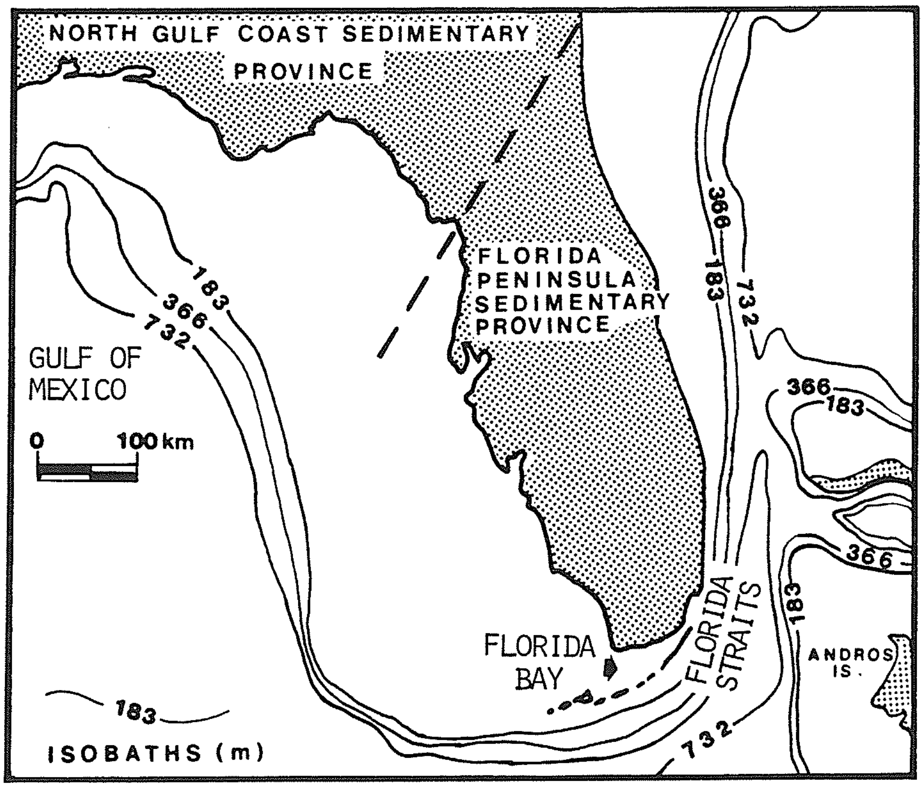
GEOLOGIC SETTING

The eastern part of the Gulf of Mexico Sedimentary Basin, including Florida, southern Georgia, southeastern Alabama, Cuba, and the Bahama Islands, can be subdivided into two sedimentary provinces: the dominantly clastic North Gulf Coast sedimentary province and the dominantly non-clastic (carbonate and evaporite) Florida Peninsula sedimentary province (Pressler, 1947, p. 1851; Perkins, 1977, p. 134; Figure 1-1). The latter approximates what is generally known as the Florida Plateau and is bounded on the east, south, and southwest by steep slopes and on the west by more gradual slopes which lead into the depths of the Gulf of Mexico (Figure 1-1). The Plateau includes most of the Florida Peninsula, all of Florida Bay, and a portion of the Gulf of Mexico.

The Florida Plateau, Bahama Banks, and Cay Sal Bank comprise the major positive features of a large carbonate province commonly referred to as the Florida-Bahama Province. The Atlantic Ocean basin, Cuba and other islands of the Antillean orogenic belt, and the Gulf of Mexico form the northeastern, southern, and western boundaries of the province. Deep (600 - 4000 m), broad-floored troughs such as the Florida Straits dissect the province (Figure 1-1).

The basement of the Florida Plateau comprises Paleozoic sedimentary and igneous rocks similar to those of the eastern United States

Figure 1-1. The Florida Plateau (outlined by contours) and its sedimentary provinces (modifed from Perkins, 1977).


NORTH GULF COAST SEDIMENTARY PROVINCE
FLORIDA PENINSULA SEDIMENTARY PROVINCE
GULF OF MEXICO
0 100 km
183
366
732
FLORIDA BAY
FLORIDA STRAITS
ANDROS IS.
183
ISOBATHS (m)

Piedmont Region and is thus interpreted as the southern buried extension of that region which continues under the coastal plain (Sheridan et al., 1966, p. 1976). The cover of sedimentary rock varies from approximately 1220 m (4000 ft) in north-central Florida to greater than 4570 m (15,000 ft) in southern Florida (Parker et al., 1955, p. 61). According to Applin (1951; cited in Perkins, 1977, p. 135), the oldest sedimentary rocks are at least Early Cretaceous in age and possibly Jurassic.

The origin of the steep marine slopes which bound the Florida Plateau and the other positive relief areas of the Florida-Bahama Province (Figure 1-1) is uncertain. Parker et al. (1955, p. 61) and Parker and Cooke (1944, p. 18) considered them to be the expression of fault scarps or monoclinal folds in the original basement complex, modified by subsequent sedimentation and erosion. Based on work in Northwest Providence Channel and Tongue of the Ocean, Andrews et al. (1970, p. 1076) supported the view (proposed by earlier workers) that these troughs originated as river valleys cut into the basement and that erosional processes have acted to preserve the relief differential. Ball (1967a, p. 266) attributed the positive and negative features of this province to post-Early Cretaceous downfolding and downfaulting which produced horst and graben structures. The relief differential was subsequently accentuated via accretion on the highs as sedimentation kept pace with regional subsidence while the deep areas were subject to very little sedimentation. Based on seismic refraction evidence, Sheridan et al. (1966, p. 1976) reasoned that the Paleocene and older seismic horizons which underlie the Florida Straits and Cay Sal Bank are without major vertical offsets and suggested that the relief of the banks and shelf edge is largely due to variations in post-Paleocene sediment thickness, especially those of the Eocene Epoch.

The emergence of the eastern part of the Florida Plateau led early workers (Parker and Cooke, 1944, p. 19; Parker et al., 1955, p. 61) to suggest that the plateau has been tilted westward. However, the possibilities of enhanced erosion of the western portion during stands of lower sea level or the preferential deposition of sediments in the east was acknowledged. These authors suggested that the time of tilting was Late Pliocene and possibly Early Pleistocene.

The youngest stratigraphic horizons of the Florida Plateau are largely undeformed (Parker and Cooke, 1944, p. 19; Perkins, 1977, p. 136). According to Perkins (1977, p. 137), the Pleistocene geologic record can be explained as a result of eustatic sea level shifts which produced the observed isopach thicknesses, shelf gradients, and facies patterns. Pre-Pleistocene relief is suspected to have greatly influenced the above, although the trend has been to progressively mask the relief differences via subsequent deposition.

PREVIOUS STUDIES

Sanford (1909) originally described and named the Pleistocene oolite of the southern Florida mainland and Keys as the 'Miami Oolite' and 'Key West Oolite', respectively. Cooke (1945) noted the cross-bedding in the Miami Oolite and described its distribution. He attributed the prominent oomoldic porosity in these limestones and calcite replacement of coral aragonite in Key Largo Limestones as the result of percolating groundwaters. Parker et al. (1955) discussed the existence of extensive solution cavities in the oolite of the Miami area.

Hoffmeister et al. (1967) divided the Miami Oolite into a 'bryozoan facies' and an 'oolitic facies.' They also compared the paleoenvironmental setting of the Atlantic Coastal Ridge oolite to the recent oolite platformward of Cat Cays in the Great Bahama Bank. They concluded that the ecological conditions which presently prevail over the Great Bahama Bank are similar to those when the Miami Limestone was deposited. In addition, the present shape and orientation of the lower Keys was thought to be due to the action of tidal currents on an unstable oolite. Ginsburg (1957) suggested that consolidation of the Miami Oolite occurred after removal from the marine environment. However, he apparently noticed a difference in the nature of the oolite above and below the water table. Where the oolite is still in the marine environment or above the water table, it is so friable that it can be easily disaggregated into individual ooids. Ginsburg also conducted a petrographic comparison of the Miami Oolite with the Mississippian Fredonia Oolite, where cementation was thought to have occurred only after removal from the marine environment. Friedman (1964) studied mineralogical and textural relationships in the oolite of Big Pine Key. He described the development of moldic porosity and the leaching of ooid cortex aragonite and replacement by calcite. Robinson (1967), in a study based on ooids from the Miami area (Pleistocene) and Bahamas (Holocene), described early diagenesis and its effects on pore size and distribution. Harris et al. (1979)

described microborings in ooids from the Atlantic Coastal Ridge and outlined the process whereby aragonite microboring molds are preserved.

The coralline limestone which constitutes the middle and upper Florida Keys was described and named by Sanford (1909). Stanley (1966) described the composition and diagenesis of the Key Largo Limestone from the Windley Key quarry and Key Largo Waterway and concluded that the reef had a deep water origin and was similar to living *Montastrea*-dominated assemblages of the windward lower levels of platform edge reefs that contain *Acropora palmata*. Diagenesis was attributed to the effects of meteoric waters. Hoffmeister and Multer (1968) discussed the origin of the Florida Keys and concluded that they were formed as a line of patch reefs in a back reef area.

Multer and Hoffmeister (1968) described the caliche crusts of the lower Keys and attributed their formation to subaerial exposure. They designated three types of crusts: a microcrystalline rind, a dense laminated crust, and a porous laminated crust. Kahle (1977) studied laminated crusts of the lower Keys and described and interpreted their biologic constituents. He concluded that the majority of near-surface micrite was derived from replacement of sediment as a result of micritization and sparmicritization processes. Perkins (1977), based on numerous cores and exposures from south Florida, described the criteria used to recognize the subaerial exposure surfaces which separate the five Pleistocene stratigraphic Q units he recognized. A detailed discussion of the depositional features of each Q unit was presented.

PURPOSE OF STUDY

The Florida Keys, Florida Bay, and the associated reef tract are the dominant features of one of the most commonly visited carbonate provinces in the world. Whereas the Holocene sediments of the area have received a great deal of attention, detailed sedimentologic studies of the Pleistocene sediments are few in number (see previous section).

The purpose of this study is to:

1. Delineate, describe, and interpret the stratigraphy and primary depositional facies of the Late Pleistocene deposits at Big Pine Key.
2. Describe and interpret the diagenetic history of these deposits.
3. Evaluate the relationship between stratigraphy, depositional facies, and diagenetic history and porosity evolution.

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CHAPTER 2 - PLEISTOCENE HISTORY OF SOUTH FLORIDA

STRATIGRAPHIC SUBDIVISION OF THE FLORIDA PLEISTOCENE

The Pleistocene strata of south Florida were divided by Perkins (1977, p. 137) into five units, from the Q1 (oldest) to the Q5 (youngest; Figure 2-1). The terrigenous and carbonate sediments which comprise these units are largely marine in origin and were deposited during interglacial stages. The glacial stages are represented by freshwater deposits and alteration products. Details concerning depositional conditions, sediments, and relationships are discussed in a later section of this chapter.

The stratigraphic units are separated from one another by subaerial exposure surfaces formed during the glacial stages. Perkins (1977, p. 137) labelled these as discontinuity surfaces. These surfaces represent a relatively short span of geological time. Previously designated formations (e.g. Key Largo Limestone) transcend the discontinuity surfaces, making them, in places, intraformational. The Q unit terminology is time stratigraphic. However, each unit surface is diachronous because it progressively demarcates the lowering of the strand line accompanying the onset of glaciation.

The Q5 unit forms the bedrock over much of south Florida, including the Florida Keys (Figure 2-2). From approximately 60 kilometres north of Miami southward and including the lower Florida Keys, the bedrock is Miami Limestone (Hoffmeister et al., 1967, p. 178; 'Miami Oolite' of

Figure 2-1. Stratigraphic terminology for south Florida (modified from Perkins, 1977).

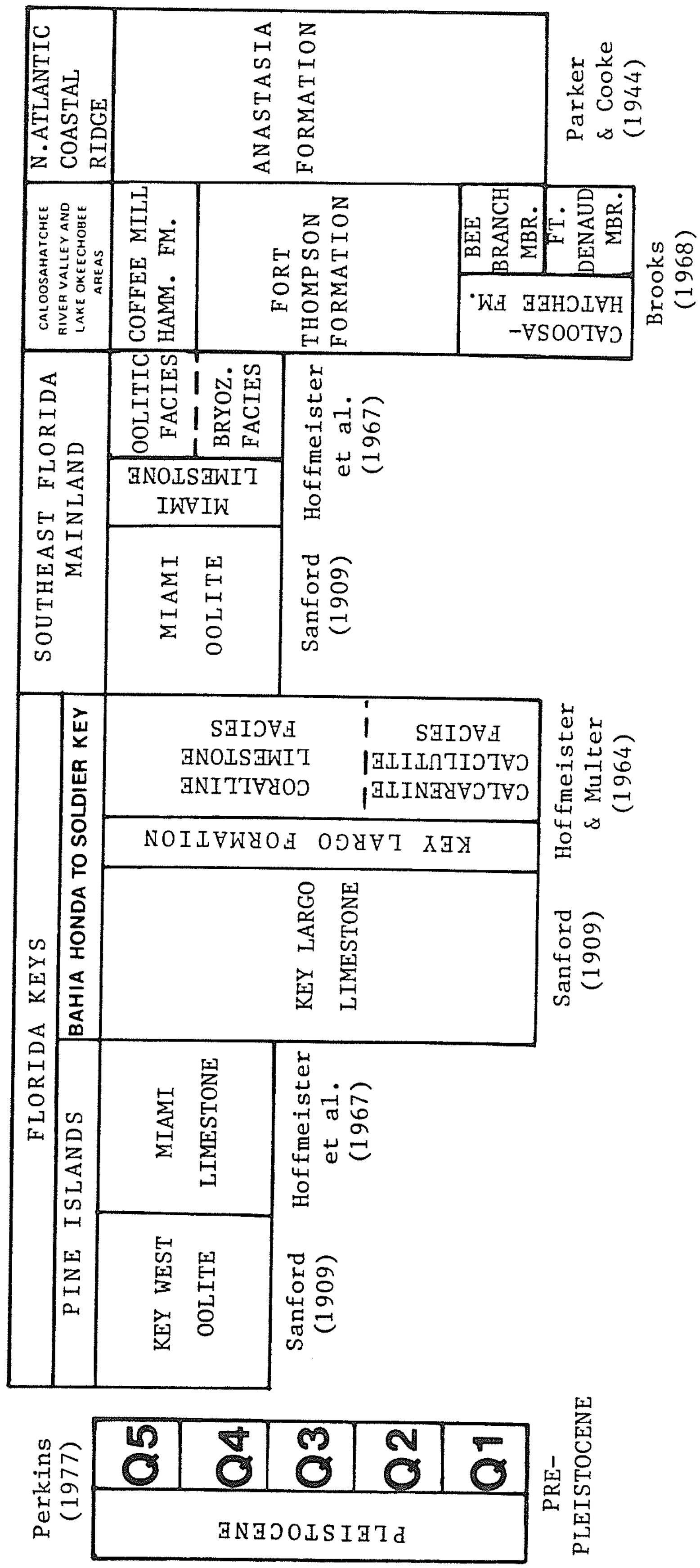
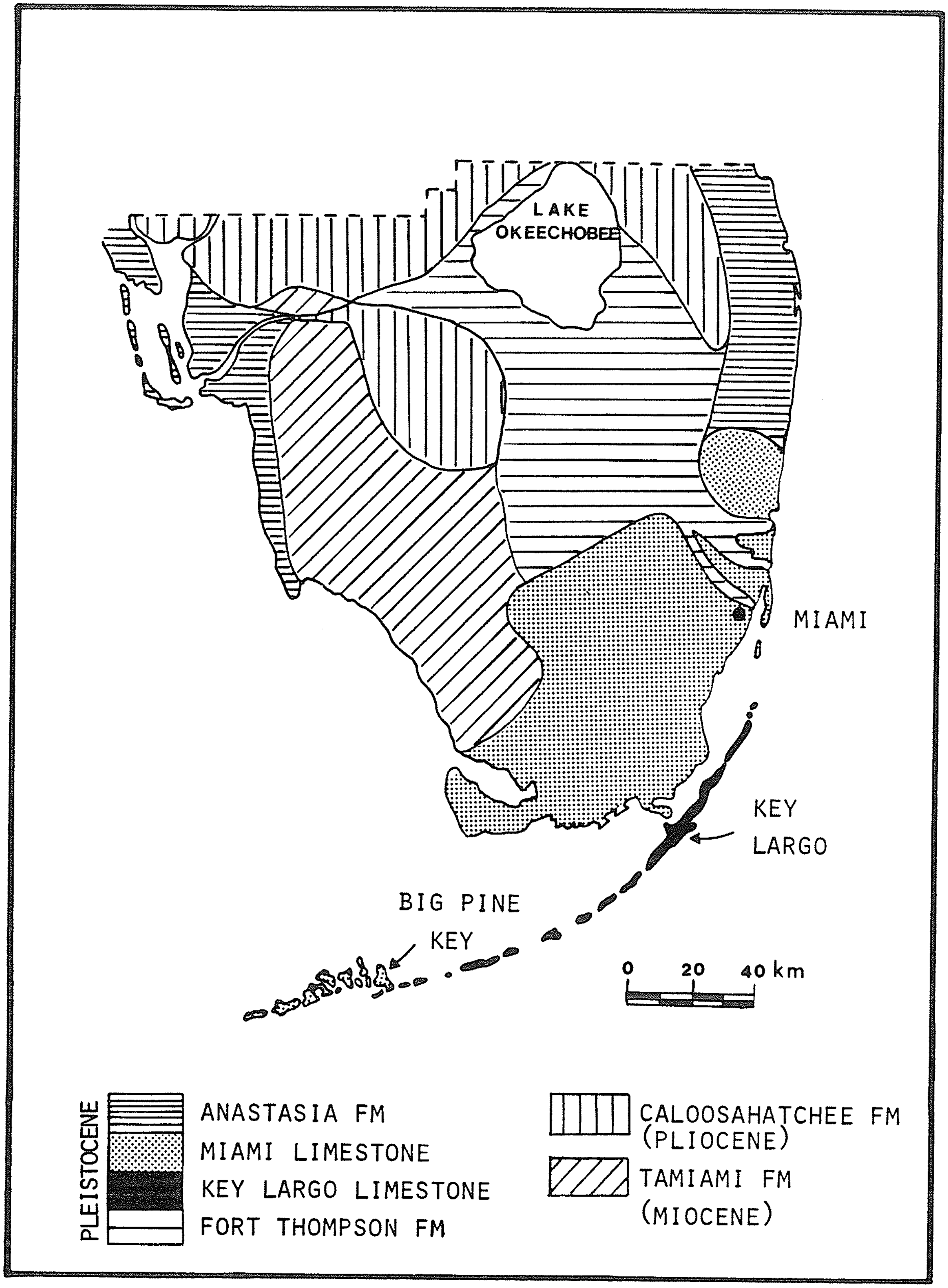

Perkins (1977)
PLEISTOCENE
Q5
Q4
Q3
Q2
Q1
PRE-PLEISTOCENE
FLORIDA KEYS
PINE ISLANDS
KEY WEST OOLITE
Sanford (1909)
MIAMI LIMESTONE
Hoffmeister et al. (1967)
BAHIA HONDA TO SOLDIER KEY
KEY LARGO LIMESTONE
Sanford (1909)
KEY LARGO FORMATION
CORALLINE LIMESTONE FACIES
CALCARENITE CALCILUTITE FACIES
Hoffmeister & Multer (1964)
SOUTHEAST FLORIDA MAINLAND
MIAMI OOLITE
Sanford (1909)
MIAMI LIMESTONE
OOLITIC FACIES
BRYOZ. FACIES
Hoffmeister et al. (1967)
CALOOSAHATCHEE RIVER VALLEY AND LAKE OKEECHOBEE AREAS
COFFEE MILL HAMM. FM.
FORT THOMPSON FORMATION
BEE BRANCH MBR.
FT. DENAUD MBR.
CALOOSA-HATCHEE FM.
Brooks (1968)
N. ATLANTIC COASTAL RIDGE
ANASTASIA FORMATION
Parker & Cooke (1944)

Figure 2-2. Geologic map of south Florida (modified from Perkins, 1977).


LAKE
OKEECHOBEE
MIAMI
KEY
LARGO
BIG PINE
KEY
0 20 40 km
PLEISTOCENE
ANASTASIA FM
MIAMI LIMESTONE
KEY LARGO LIMESTONE
FORT THOMPSON FM
CALOOSAHATCHEE FM
(PLIOCENE)
TAMIAMI FM
(MIOCENE)

Sanford, 1909). The Key Largo Limestone comprises the middle and upper Keys (Hoffmeister and Multer, 1968, p. 1490).

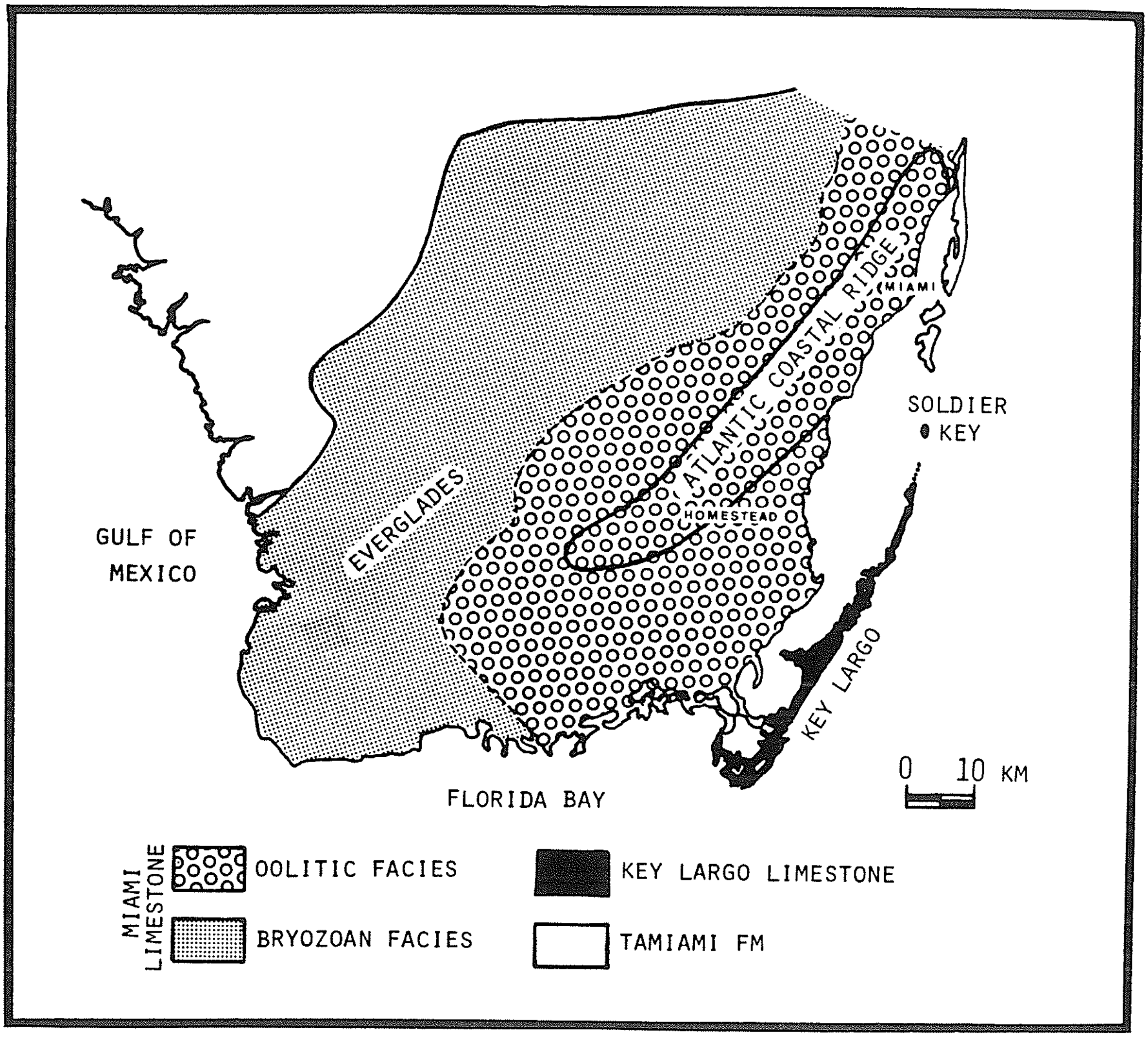
The Miami Limestone is subdivided into an oolitic and a bryozoan facies (Hoffmeister et al., 1967, p. 178; Figure 2-3). The oolitic facies occurs as the bedrock of the lower Florida Keys and also comprises the Atlantic Coastal Ridge which extends from Miami to Homestead, a distance of approximately 60 km (Halley et al., 1977, fig. 1). The bryozoan facies underlies the oolitic facies of the Atlantic Coastal Ridge and extends to the west and southwest to form the bedrock in much of the Everglades. The facies name was based on the occurrence of a large number of massive tubular cheilostome bryozoan colonies (*Schizoporella floridana* Osburn).

The Key Largo Limestone of the Q5 unit is composed of highly coralline sediments and can be observed to pass laterally into the Miami Limestone at only one location - the southeast point of Big Pine Key (Figure 2-2). Drill holes in the Florida Keys and adjacent mainland (e.g. Perkins, 1977; this work) demonstrate that the Key Largo Limestone constitutes not only the surface bedrock for the middle and upper Keys, but also extends southwestward along the Florida Keys where it underlies the oolitic facies of the Miami Limestone (Hoffmeister and Multer, 1968, p. 1490). Key Largo Limestone lithology also constitutes much, if not all, of the Q3 and Q4 units underlying the present-day Florida Keys.

CHRONOLOGY OF THE Q UNITS

The Key Largo and Miami Limestones of the Q5 unit have been dated by Osmond et al. (1965) and Broecker and Thurber (1965) using uranium series

Figure 2-3. Location of the Atlantic Coastal Ridge and distribution of the oolitic and bryozoan facies of the Miami Limestone (modifed from Hoffmeister et al., 1967).


GULF OF MEXICO
EVERGLADES
ATLANTIC COASTAL RIDGE
MIAMI
HOMESTEAD
SOLDIER KEY
KEY LARGO
FLORIDA BAY
0 10 KM
MIAMI LIMESTONE
OOLITIC FACIES
BRYOZOAN FACIES
KEY LARGO LIMESTONE
TAMIAMI FM

dating methods ($^{230}Th/^{234}U$). Corals and ooids from these formations have yielded ages of 120,000-140,000 years BP, suggesting contemporaneous deposition of the Miami and Key Largo Limestones of the Q5 unit. More recently, Mitterer (1975) used an amino acid dating technique to establish the Pleistocene chronology and average diagenetic temperatures experienced by fossils, particularly the bivalve Mercenaria. Application of the technique to Pleistocene deposits in Florida delineated 5 post-Miocene, probably Pleistocene, marine units deposited during interglacial times. The ages, from youngest to oldest, are 134,000 years BP, 180,000 years BP, 236,000 years BP, 324,000 years BP, and an undetermined age for the oldest unit. Perkins (1977, p. 140) proposed that these ages are applicable to the units Q5 through Q1, respectively, which he designated in south Florida (Figure 2-4).

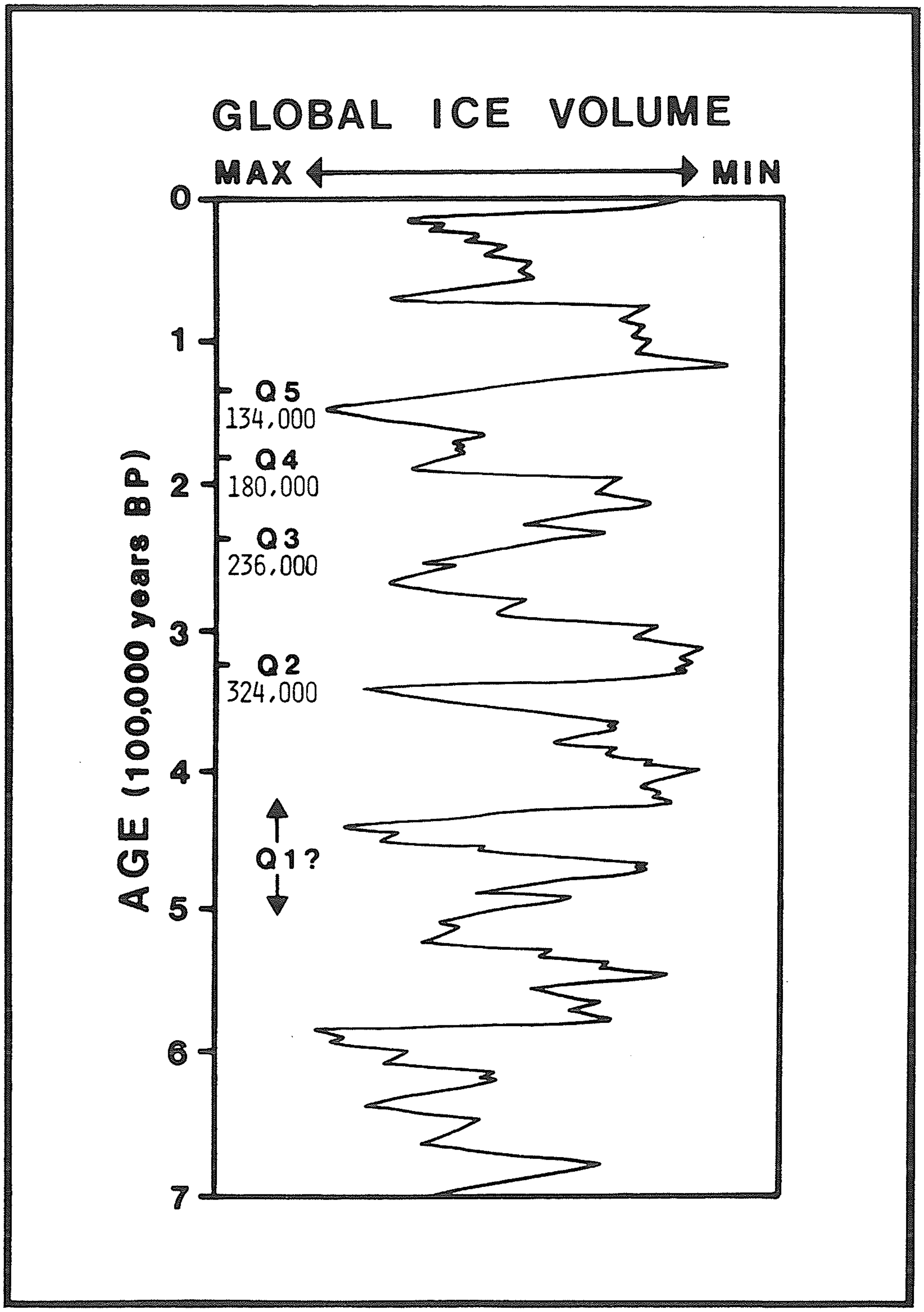
PLEISTOCENE DEPOSITION IN SOUTH FLORIDA

INTRODUCTION

Most of the following discussion is based on the work of Perkins (1977), who described the pre-Q unit paleotopography and the unit thickness, lithology, and ecologic and environmental facies of each of his five Q units in detail. Perkins (1977) based his study on 52 cores and 4 surface exposures distributed in the portion of south Florida extending from Lake Okeechobee southward to the reef tract of the southern Florida Keys. Two of his cores are from the Keys, and one core each from Florida Bay and offshore on the reef tract. The remainder are scattered about the mainland.

In Perkins' (1977) work, ecologic suites and related environmental

Figure 2-4. Comparison of Q unit ages and global ice volume during the last 700,000 years. The Q unit ages are those of Mitterer, 1975 (modified from United States Committee for Global and Atmospheric Research Program, 1975).


GLOBAL ICE VOLUME
MAX
MIN
AGE (100,000 years BP)
0
1
2
3
4
5
6
7
Q5
134,000
Q4
180,000
Q3
236,000
Q2
324,000
Q1?

interpretations were based largely on similar living assemblages in south Florida (Table 2-1). Many elements of the Pleistocene biota have living counterparts which can be observed in their natural environment. The Pleistocene sediments closely reflect the environments which prevailed, because complete skeletons and disaggregated debris comprise a large proportion of the sediment. A typical Q unit depositional sequence begins with a shallow water phase which progressively deepens and then regresses back to shallow water as the sequence nears termination. The ecologic designation of the various marine units was determined by the most open (and deep) marine fauna (Perkins, 1977, p. 151).

The pre-Pleistocene basement comprises several positive and negative features which have greatly influenced the type and distribution of Pleistocene sediments. A paleotopographic map (Figure 2-5) shows the following major features: Brighton high, Allapatah lobe, Immokalee high, Cape Sable high, Okeechobee depression, and the Caloosahatchee depression.

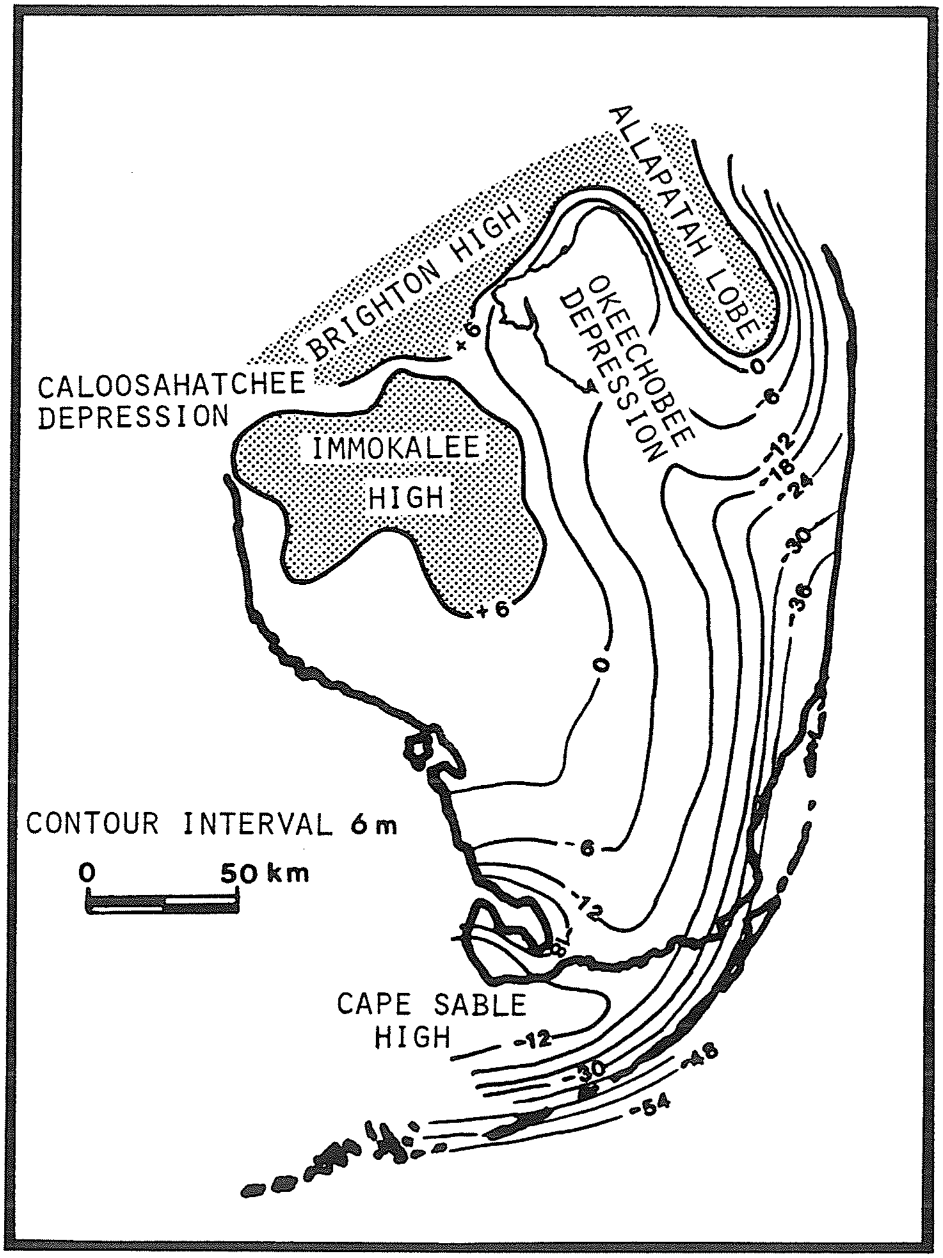
In general, Pleistocene deposition can be viewed as a constant battle between the detrital influx from the topographic highs and essentially *in situ* carbonate production which also varied, in many cases, according to the prevailing organic communities. Initially, Pleistocene deposition was largely influenced by the pre-Pleistocene paleotopographic features (Figure 2-5), but preferential infilling of depressions ultimately led to successively smaller amounts of pre-unit relief.

During the entire time of Pleistocene deposition, the northeastern part of Perkins' (1977) study area (Figure 2-2) was characterized by a beach-dune-lagoon-delta complex with its associated (mostly terrigenous)

Table 2-1. Present-day ecologic suites of south Florida (modified from Perkins, 1977).

ECOLOGIC SUITE	ASSEMBLAGE
I. Reef tract including patch reefs	*Montastrea*, *Diploria*, *Porites astreoides*, *Acropora cervicornis*, *Mussa*, *Millepora*, *Halimeda*, *Lithothamnion*, *Porites porites*, *Spondylus*
II. Behind inner reef tract to Keys, channels through Keys, Atlantic subenvironment of Florida Bay	*Porites divaricata*, *Manicina*, *Siderastrea*, *Solenastrea*, *Cladocora*, *Goniolithon*, *Schizoporella*, *Halimeda*
III. Florida Bay, nearshore shoals of southwest Florida coast	Predominantly molluscan (gastropods and pelecypods), *Chione cancellata* common
IV-A. Nearshore transitional; bar, beach, dune, lagoon complex of Florida east coast including tidal channels and inlets	Molluscan debris; local *Halimeda* and corals in tidal inlets
IV-B. Brackish to estuarine; bays and estuaries of southwest Florida coast coast	Oysters, *Rangia cuniata*
V. Freshwater deposits, Lake Okeechobee and Everglades	Pulmonate gastropods, *Helisoma* common

Figure 2-5. Pre-Pleistocene paleotopographic features of south Florida. Contour values relative to mean sea level (modified from Perkins, 1977).


ALLAPATAH LOBE
BRIGHTON HIGH
OKEECHOBEE DEPRESSION
CALOOSAHATCHEE DEPRESSION
IMMOKALEE HIGH
CONTOUR INTERVAL 6 m
0
50 km
CAPE SABLE HIGH
+6
0
-6
-12
-18
-24
-30
-36
-48
-54

sediments which largely comprise the Anastasia Formation. The Allapatah lobe and Immokalee high provided a continuous clastic influx which formed extensive shoals. However, the latter terminated its contribution at the end of Q4 sedimentation. The Okeechobee depression persisted throughout Pleistocene time and the mollusc fragment wackestones, packstones, grainstones, and quartz sandstones which accumulated are largely those of the Fort Thompson Formation (Figure 2-2).

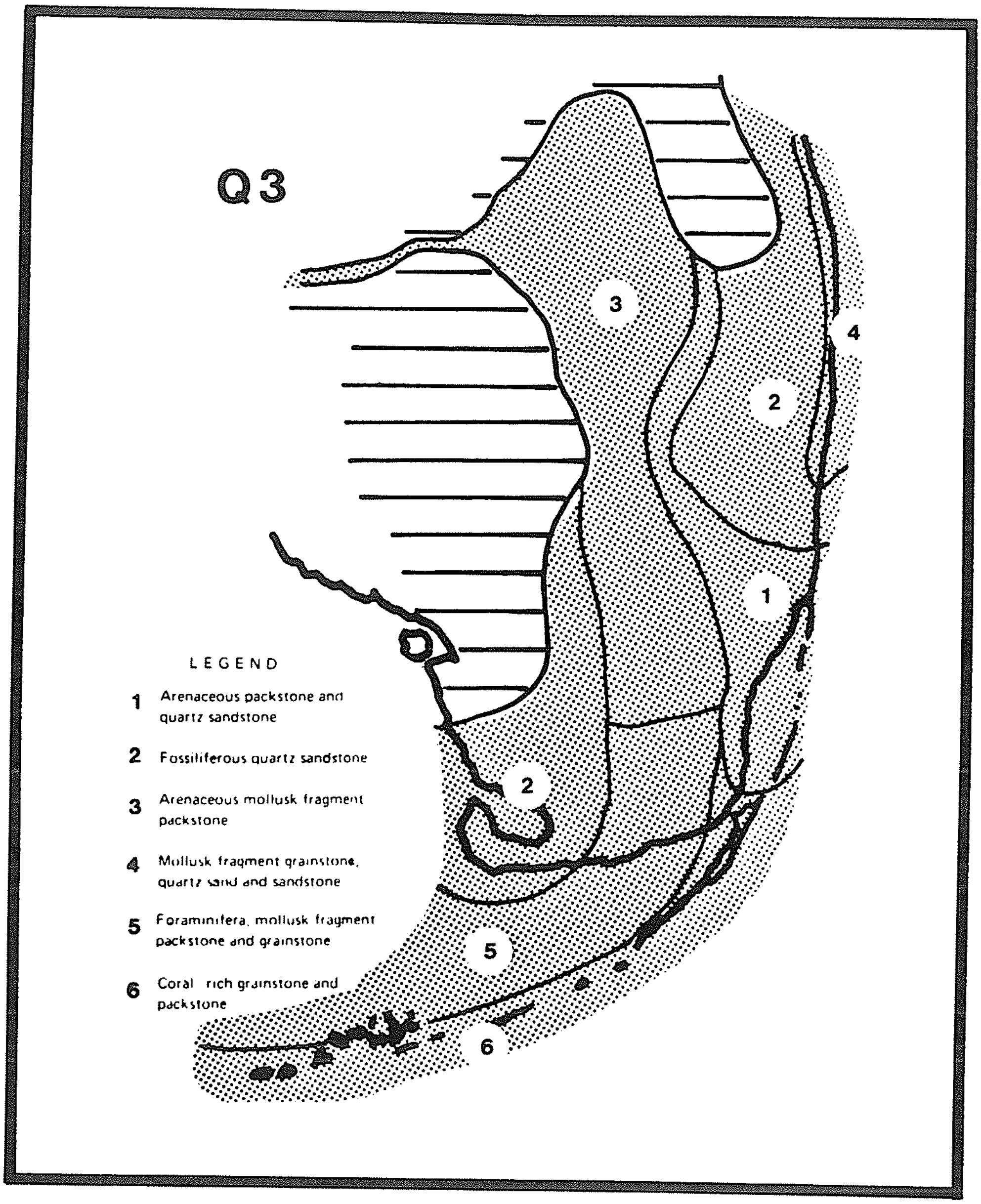
Q1 AND Q2 DEPOSITION

The conditions of Q1 and Q2 deposition were similar. In the area of the Keys, sediments consisted of arenaceous mollusc fragment packstones and fossiliferous quartz sandstones deposited in open marine conditions. Based on the shallow water sediments of the Q1 and Q2 units in the Caloosahatchee depression, maximum water depths at the site of Big Pine Key were 56 m (185 ft) and 46 m (150 ft) for Q1 and Q2 times, respectively. However, faunal evidence of such depths is lacking and only the shallow water ecologic suites that were deposited in water of 9 m (30 ft) or less are represented.

Q3 DEPOSITION

The Q3 unit marks the transition where the proportion of carbonate to clastic sedimentation was dramatically increased, possibly due to favourable climatic conditions or a reduction in the clastic influx which dominated Q1 and Q2 deposition. The approximate arc of the Florida Keys was characterized by a coral reef tract consisting of highly coralline grainstones and packstones (Figure 2-6). The fauna included <u>Montastrea</u>,

Figure 2-6. Q3 unit depositional facies. Ruled area indicates land (modified from Perkins, 1977).


Q3
LEGEND
1 Arenaceous packstone and quartz sandstone
2 Fossiliferous quartz sandstone
3 Arenaceous mollusk fragment packstone
4 Mollusk fragment grainstone, quartz sand and sandstone
5 Foraminifera, mollusk fragment packstone and grainstone
6 Coral rich grainstone and packstone
3
4
2
1
2
5
6

Diploria, Porites astreoides, Porites porites, and coralline algae (ecologic suite I, Table 2-1). Matrix constituents included Halimeda and peneroplid and miliolid foraminifera. Adjacent to the reef tract and occupying much of the Florida Bay area and beyond, foraminiferal and mollusc fragment packstones and grainstones were deposited in an open marine environment (ecologic suite II, Table 2-1).

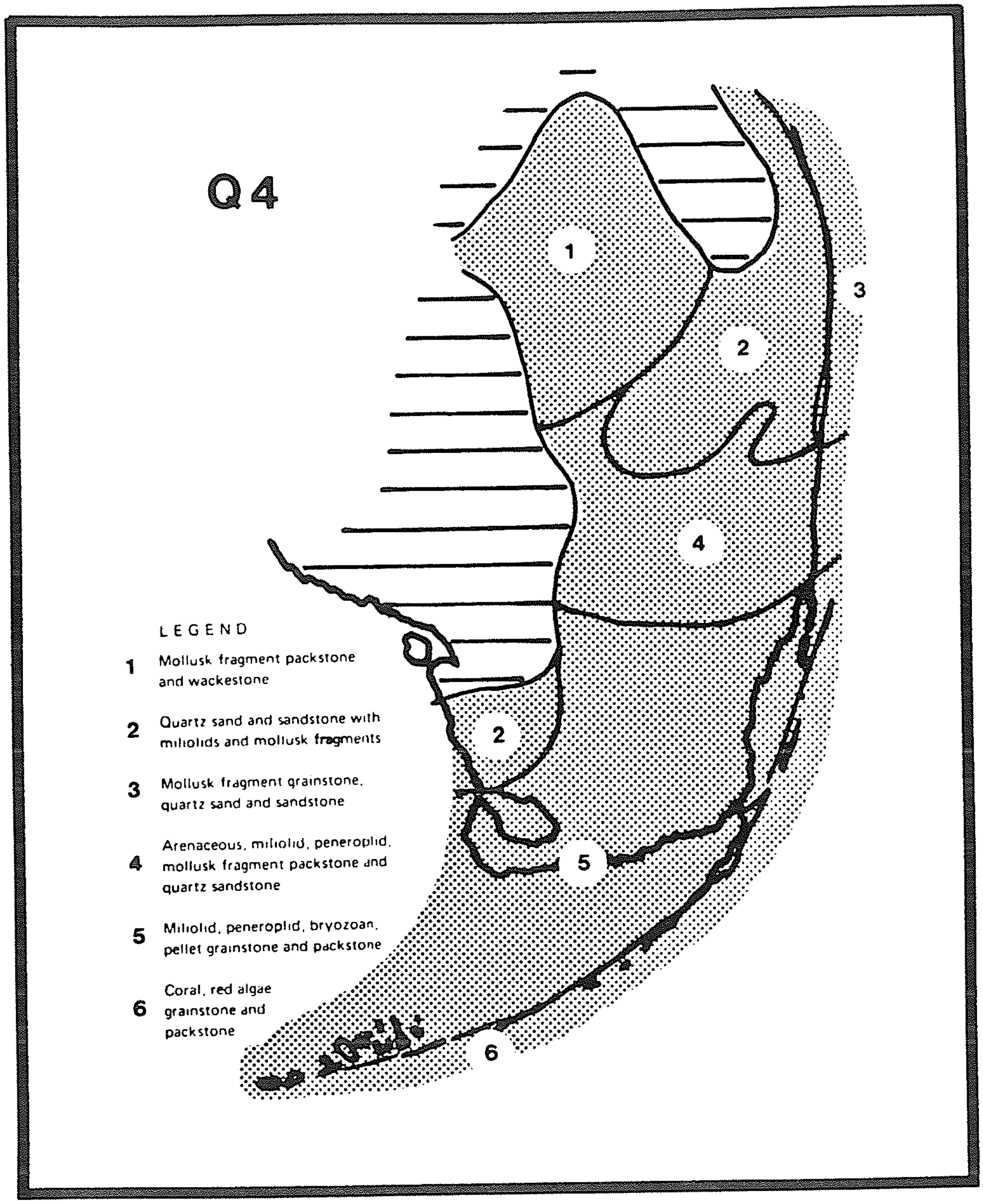
The construction of bioherms in the area of the Keys led to a very thick Q3 unit at this location. This unit reached a maximum thickness of approximately 30 m (97 ft) in a core from Big Pine Key (Perkins, 1977, core no. 56).

At the peak of the Q3 transgression, the maximum water depth at the site of Big Pine Key was estimated to be 17 m (55 ft).

Q4 DEPOSITION

In the area occupied by much of the Everglades, Florida Bay, and extending to the lower Keys, a highly burrowed (Callianassa) peloidal grainstone and packstone facies which included miliolid and peneroplid foraminifera, local Manicina, Porites, and Schizoporella existed in an open marine environment with conditions very similar to those of Great Bahama Bank (ecologic suite II, Table 2-1; Figure 2-7). These sediments ('bryozoan facies' of Hoffmeister et al., 1967) grade southward into highly coralline and red algal packstones and grainstones characterized by Montastrea, Diploria, Halimeda, and coralline algae. The most seaward control point of Perkins' (1977) study was a core drilled from Little Molasses Island where the sediments were composed of highly coralline and red algal packstones and grainstones with Halimeda and Spondylus. These

Figure 2-7. Q4 unit depositional facies. Ruled area indicates land (modified from Perkins, 1977).


Q4
1
2
3
4
2
5
6
LEGEND
1 Mollusk fragment packstone and wackestone
2 Quartz sand and sandstone with miliolids and mollusk fragments
3 Mollusk fragment grainstone, quartz sand and sandstone
4 Arenaceous, miliolid, peneroplid, mollusk fragment packstone and quartz sandstone
5 Miliolid, peneroplid, bryozoan, pellet grainstone and packstone
6 Coral, red algae grainstone and packstone

sediments, however, represent a deeper water environment than the similar sediments to the west.

Sedimentation during Q4 time is thought to have taken place in water depths of approximately 9 m (30 ft).

Q5 DEPOSITION

Q5 unit sediments (Figures 2-8, 9) were deposited on a relatively featureless, gently seaward dipping platform. The coralline facies of Q4 deposition produced a linear arcuate high (present site of the Florida Keys). Seaward of this high the slope gradient to the outer shelf margin was approximately 3 m/km (17 ft/mi).

In the northeastern part of Perkins' (1977) study area the beach-dune-lagoon-delta complex persisted as it had during all of the Pleistocene. Laterally along depositional strike it graded into northeast-southwest trending oolitic sediments which consisted of ooid and pellet grainstones and packstones with miliolids, peneroplids, Halimeda, and locally Schizoporella ('oolitic facies' of Hoffmeister et al., 1967). Open marine conditions existed here with water depths of approximately 9 m (30 ft). The sediments are cross-bedded in places, but locally intense bioturbation may have destroyed all traces of bedding. Callianassa and annelid burrows are common.

Another oolitic facies (approximately east-west trending) containing less mud developed at the site of the present lower Florida Keys (Figure 2-8).

The central portion of the arcuate high produced by the Q4 coralline facies existed as a subaerial high during early Q5 transgression. Plat-

Figure 2-8. Q5 unit depositional facies. Ruled area indicates land (modified from Perkins, 1977).

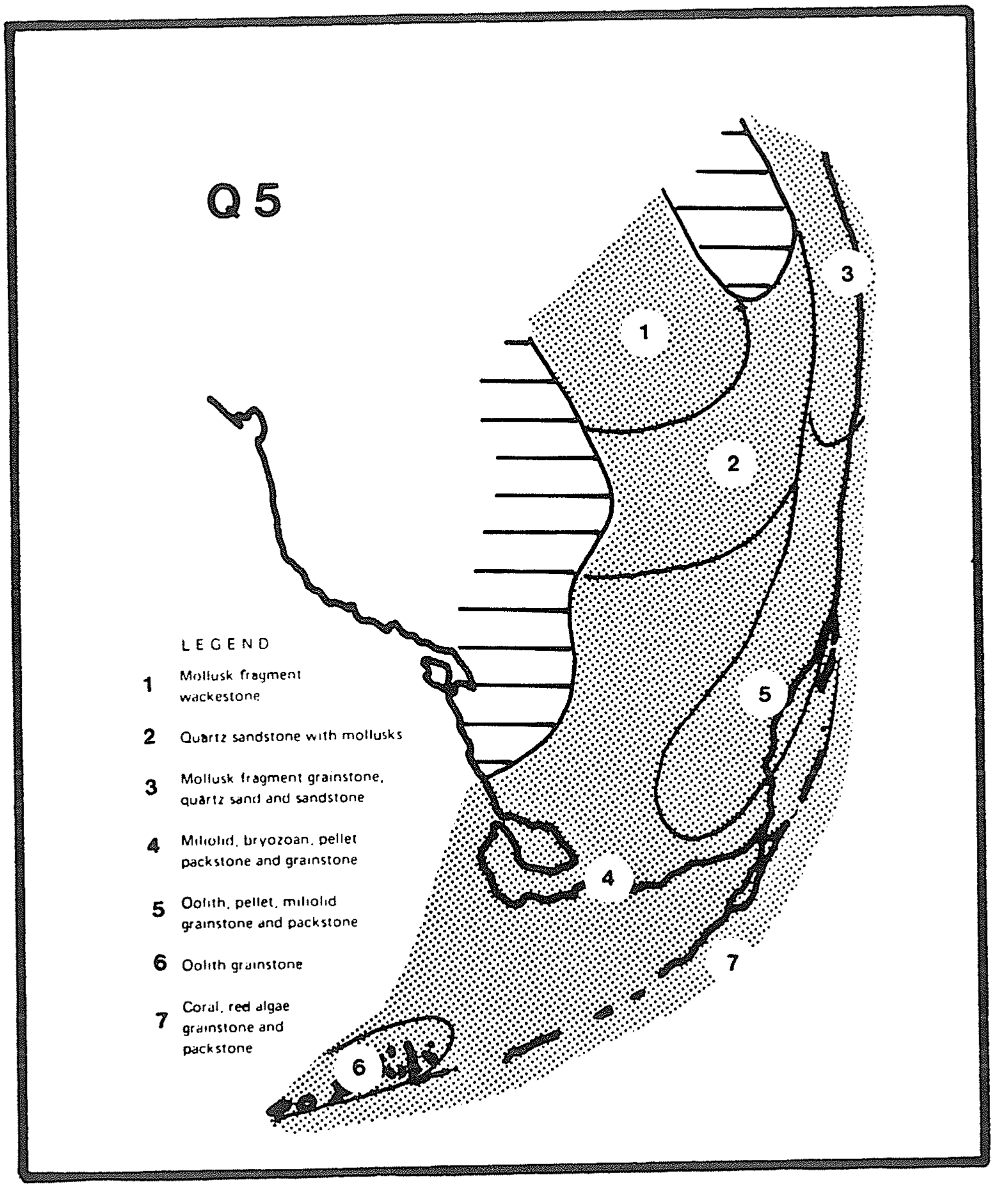
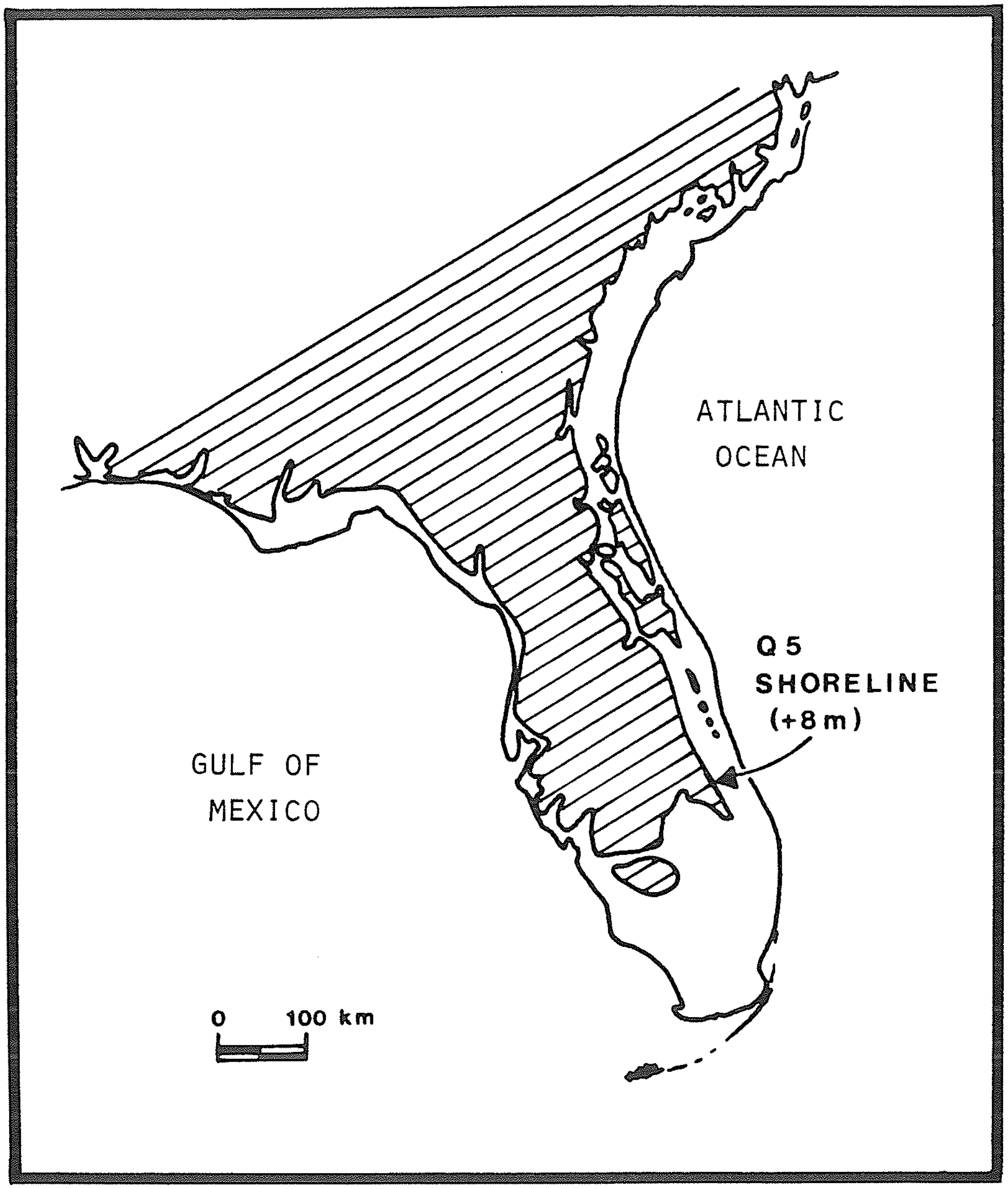

Q5
LEGEND
1 Mollusk fragment wackestone
2 Quartz sandstone with mollusks
3 Mollusk fragment grainstone, quartz sand and sandstone
4 Miliolid, bryozoan, pellet packstone and grainstone
5 Oolith, pellet, miliolid grainstone and packstone
6 Oolith grainstone
7 Coral, red algae grainstone and packstone
1
2
3
4
5
6
7

Figure 2-9. Approximate shoreline position at the height of Q5 transgression. This position corresponds with that previously designated by Cooke (1945) as the Pamlico shoreline (modified from Cooke, 1945).


ATLANTIC
OCEAN
Q 5
SHORELINE
(+8 m)
GULF OF
MEXICO
0 100 km

formward, the high was flanked by ooid shoals. The high decreased in elevation northward and southward away from the central portion. With rising sea level the high became a substrate for renewed coral growth, and highly coralline sediments similar to that of Q4 time were deposited. They consisted of various corals, red algae, and foraminifera in grainstones and packstones.

The origin of the Key Largo Limestone and its relationship to the ooid shoals of the Q5 unit have yet to be satisfactorily explained. Based mainly on a literature survey, the origin enigma is discussed in Appendix A. Although this discussion deals mainly with the Q5 unit Key Largo reef, the Key Largo reefs of the Q3 and Q4 units presumably share a common origin with the Q5 unit reef.

MIAMI LIMESTONE: DEPOSITION AND POSITION IN THE Q5 FRAMEWORK

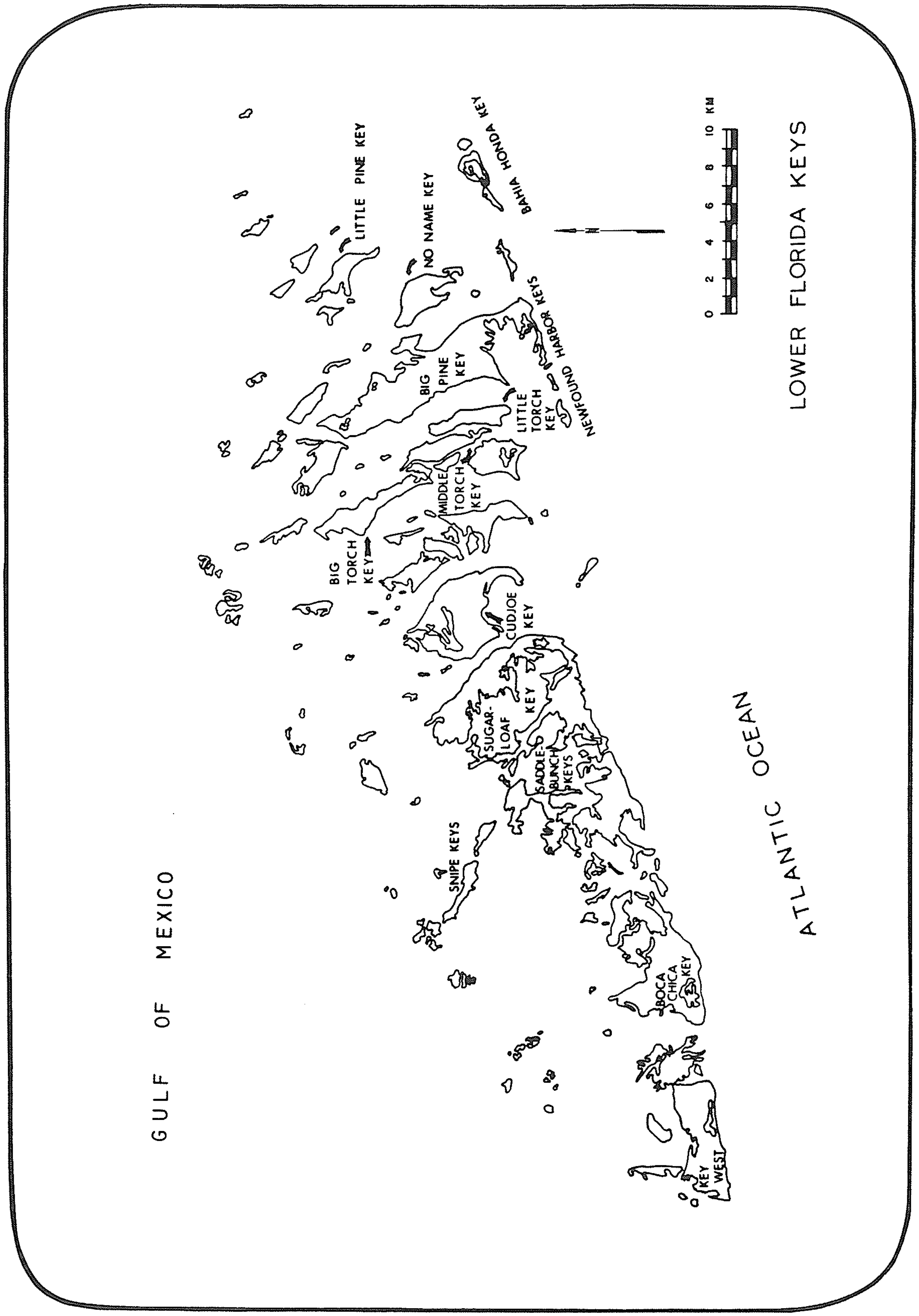
During Q5 deposition, ooid shoals accumulated slightly platformward at the extremeties of the Key Largo reef and the peloidal grainstones and packstones of Q4 deposition persisted, framed on the south by ooid shoals (lower Florida Keys), on the east by the Key Largo reef, on the north by ooid shoals (Atlantic Coastal ridge), and on the west by the Immokalee high.

The ooid shoals forming the Atlantic Coastal Ridge grade along depositional strike into the grainstones and sandstones of the beach-dune-lagoon-delta complex (Anastasia Formation). The ridge consists of broad areas approximately 4.8 km (3 mi) across and more than 2.7 m (9 ft) above sea level, separated by 0.8 km (0.5 mi) wide depressions which extend for several kilometres and are generally less than 2.1 m (7 ft) above sea level. The ridge decreases in elevation from 4.6 m (15 ft) near Miami to 3 m (10 ft) near its southern end (Halley et al., 1977, p. 519; Figure 2-3). In some areas, low sea cliffs are cut into the ridge, the most famous occurring at Silver Bluff (Parker and Cooke, 1944, p. 53-54). The ridge is well cross-bedded in its upper portions and the oolite has a maximum thickness of approximately 10 m (34 ft). Its base reaches approximately 3 m (10 ft) below sea level, where it overlies the bryozoan facies.

According to Hoffmeister et al. (1967, p. 178) this oolitic deposit is largely confined to the ridge, although oolitic sheets do overlie the bryozoan facies as a thin cover which extends several kilometres to the west (Figure 2-3). The cross-beds may dip up to 30° and strike in any direction. Cross-bedding becomes less distinct in a westward direction across the ridge. The dominant grains of the oolitic facies are ooids with subordinate peloids and bioclasts. The sediments gradually become less oolitic away from the ridge until, at its margin, ooids may constitute approximately 10% of the sediment with the bulk consisting of peloids and bioclasts.

The lower Florida Keys comprise an approximately east-west oriented oolitic mound (Q5) with many of the same characteristics as the Atlantic Coastal Ridge (Figure 2-10). Geometrically, the lower Keys form a bar which decreases in width from Big Pine Key to Key West. The channels which dissect the bar and connect the Atlantic Ocean with the Gulf of Mexico do so at approximately right angles. The lower Keys resemble a tidal bar belt (see Ball, 1967b) and are interpreted as having such an

Figure 2-10. The lower Florida Keys.


LOWER FLORIDA KEYS
GULF OF MEXICO
ATLANTIC OCEAN
0 2 4 6 8 10 KM
N
BAHIA HONDA KEY
LITTLE PINE KEY
NO NAME KEY
BIG PINE KEY
HARBOR KEYS
NEWFOUND
LITTLE TORCH KEY
MIDDLE TORCH KEY
BIG TORCH KEY
CUDJOE KEY
SUGAR-LOAF KEY
SADDLE-BUNCH KEYS
SNIPE KEYS
BOCA CHICA KEY
KEY WEST

origin (Hoffmeister et al., 1967, p. 189). Big Pine Key is the largest of the southern Keys and is approximately 14 km long and 2-3 km wide. Present day tidal passes occupy the positions of the Pleistocene relicts.

The extent of the lower Keys oolite is approximated by the area of the present Keys. The oolite grades to the north and northeast into peloidal grainstones and packstones ('bryozoan facies' of Hoffmeister et al., 1967; Figure 2-8).

RELATIONSHIP BETWEEN Q5 OOLITE AND KEY LARGO REEF

Geographically, the oolitic deposits of the Atlantic Coastal Ridge and the lower Florida Keys are part of the same arcuate trend as the Key Largo reefal sediments. However, the oolitic deposits lie slightly farther back on the platform. The factor(s) which caused the Key Largo 'reef' to nucleate where it did (paleotopography of the Q3 and Q4 unit surfaces and subsequent exaggeration from biohermal construction) also exerted an influence that determined approximately where the ooid shoals could form.

Big Pine Key marks the surface transition between the Miami and Key Largo Limestones of the Q5 unit. Westward of Big Pine Key, no emergent Key Largo Limestone lies seaward of the Keys. The oolite overlaps reefal material in the area of Big Pine Key with a transitional northward dipping contact (see Hoffmeister and Multer, 1964, p. 60; this work, Chapter 3). This is the only surface contact in the lower Keys. The oolite thickens northward at a right angle to regional strike in the lower Keys. Hoffmeister and Multer (1964, p. 60) reported a thickness of 6 m (20 ft) from a borehole in the centre of Key West.

It is this oolite and its transition to the coralline Key Largo Limestone of the Q5 unit and the nature of the Q3 and Q4 units of Big Pine Key to which attention will now be drawn.

CHAPTER 3 - PRIMARY DEPOSITIONAL FACIES AND STRATIGRAPHY OF BIG PINE KEY

INTRODUCTION

This study is based on core information from 9 boreholes and from 22 outcrop samples from Big Pine Key (Figures 3-1, 2). Boreholes BP-1 through BP-7 were made available by the United States Geological Survey Energy Resources Division (Miami). Boreholes MO-2 and MO-6 were made available by the USGS Water Resources Division (Miami). All cores are 4.8 cm diameter except BP-1 which is 3.5 cm. Average borehole depth is 8.2 m. Core recovery was variable (Figure 3-3). Intervals of high secondary porosity (large dissolution vugs and channels; discussed in Chapter 6) typically led to poorer recovery than those intervals with less intensive development of secondary porosity.

In order to provide continuity to information obtained from drill cores, surface exposures in Big Pine Key were sampled. Average hand specimen size is approximately 1 kilogram. Typically, a hard caliche surface crust overlies the 'fresh' rock. Because of low relief and essentially flat-lying strata, outcrops on Big Pine Key are almost totally confined to small exposures approximately 1 m or less above sea level exposed in various boat slips, drainage ditches, and more rarely in freshwater ponds.

Discontinuity surfaces in the core permitted the recognition of three stratigraphic units in the study area. These are the Q5, Q4, and Q3 units of Perkins (1977). Three major facies are recognized within

Figure 3-1. Borehole and highway locations on Big Pine Key.

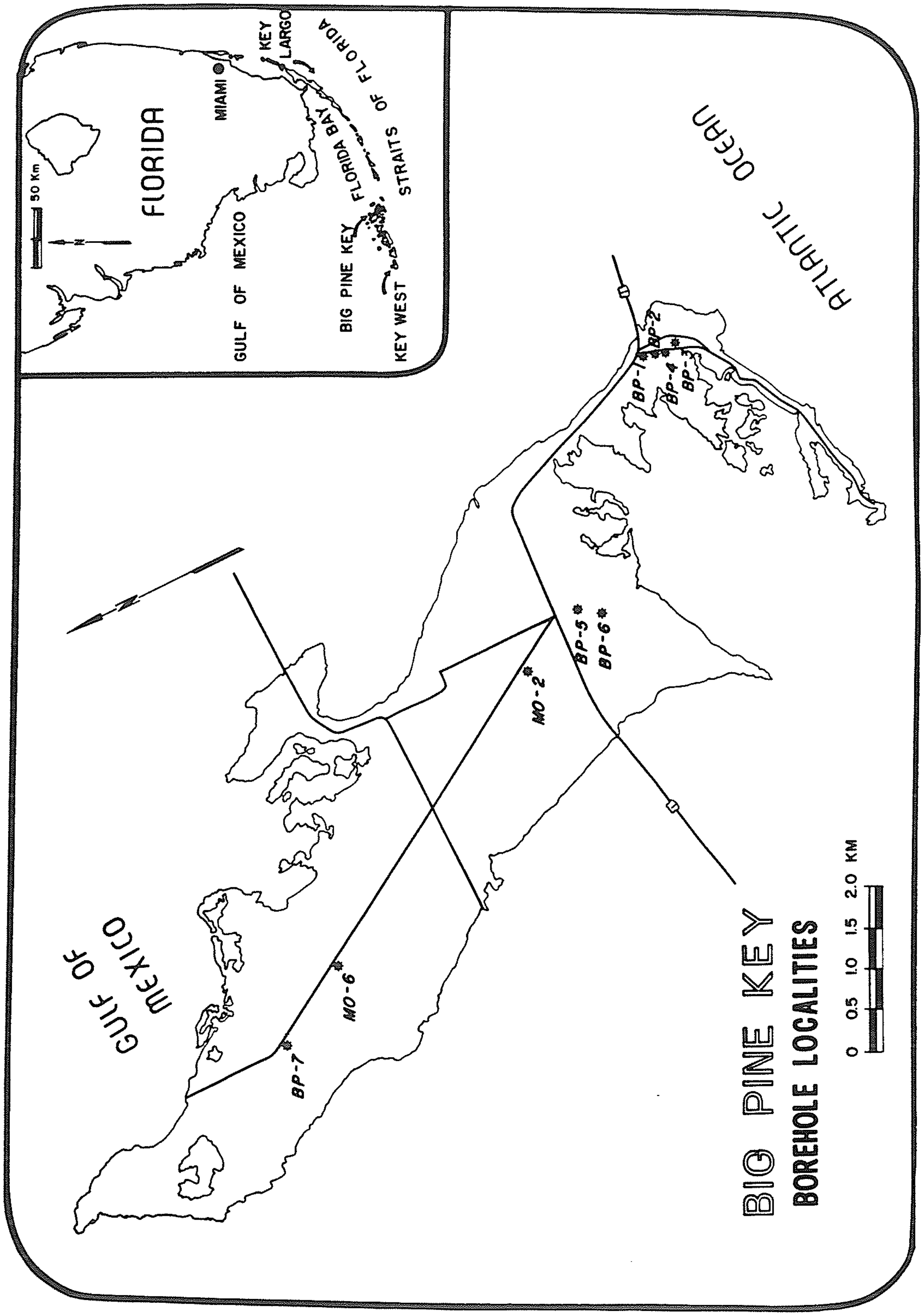

FLORIDA
MIAMI
KEY
LARGO
FLORIDA BAY
STRAITS OF FLORIDA
GULF OF MEXICO
BIG PINE KEY
KEY WEST
50 Km
N
ATLANTIC OCEAN
GULF OF MEXICO
BP-1
BP-2
BP-3
BP-4
BP-5
BP-6
BP-7
MO-2
MO-6
N
BIG PINE KEY
BOREHOLE LOCALITIES
0 0.5 1.0 1.5 2.0 KM

Figure 3-2. Approximate contact of the Miami Limestone and the Key Largo Limestone at the southeast point of Big Pine Key. Map also shows borehole and hand specimen locations.

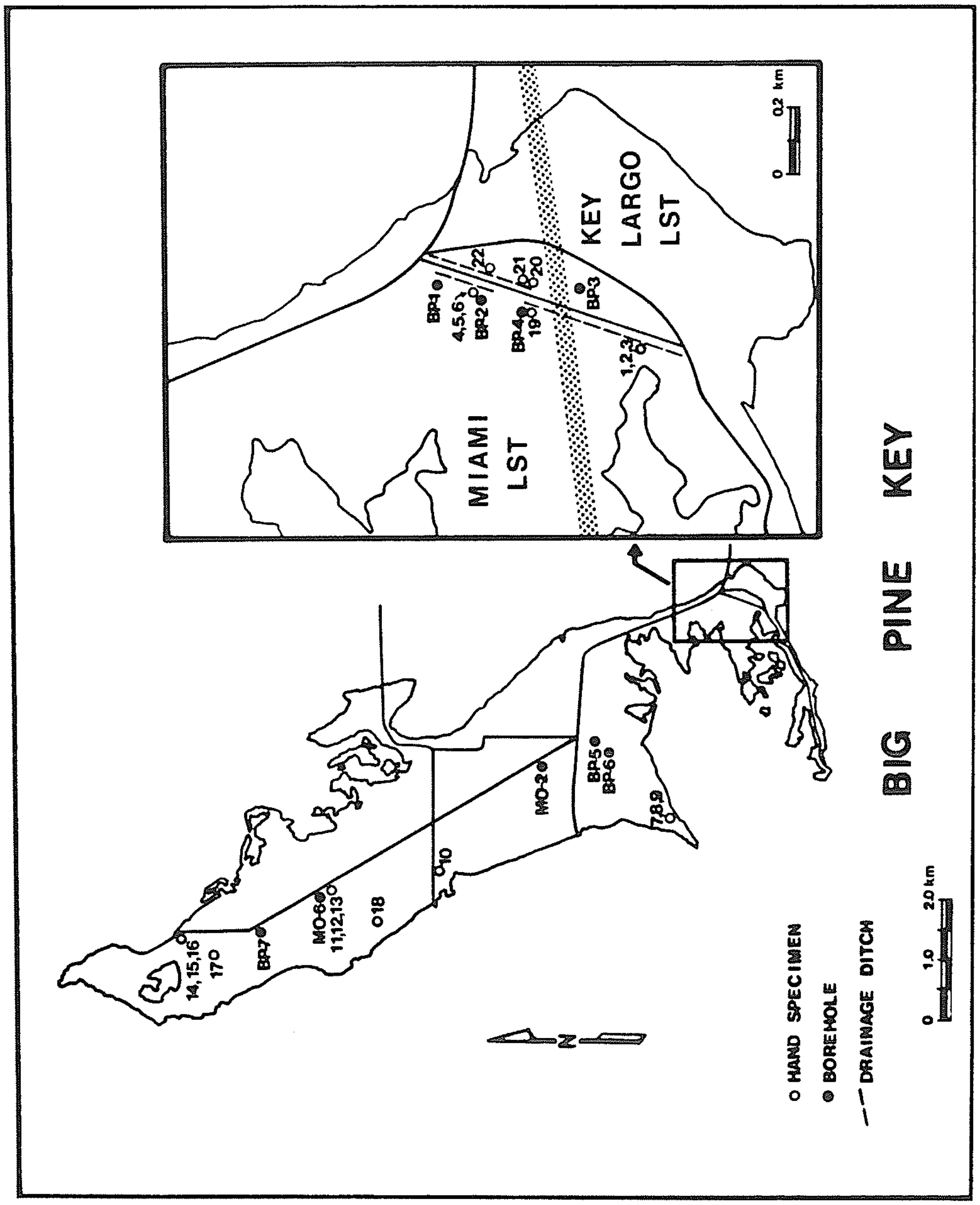
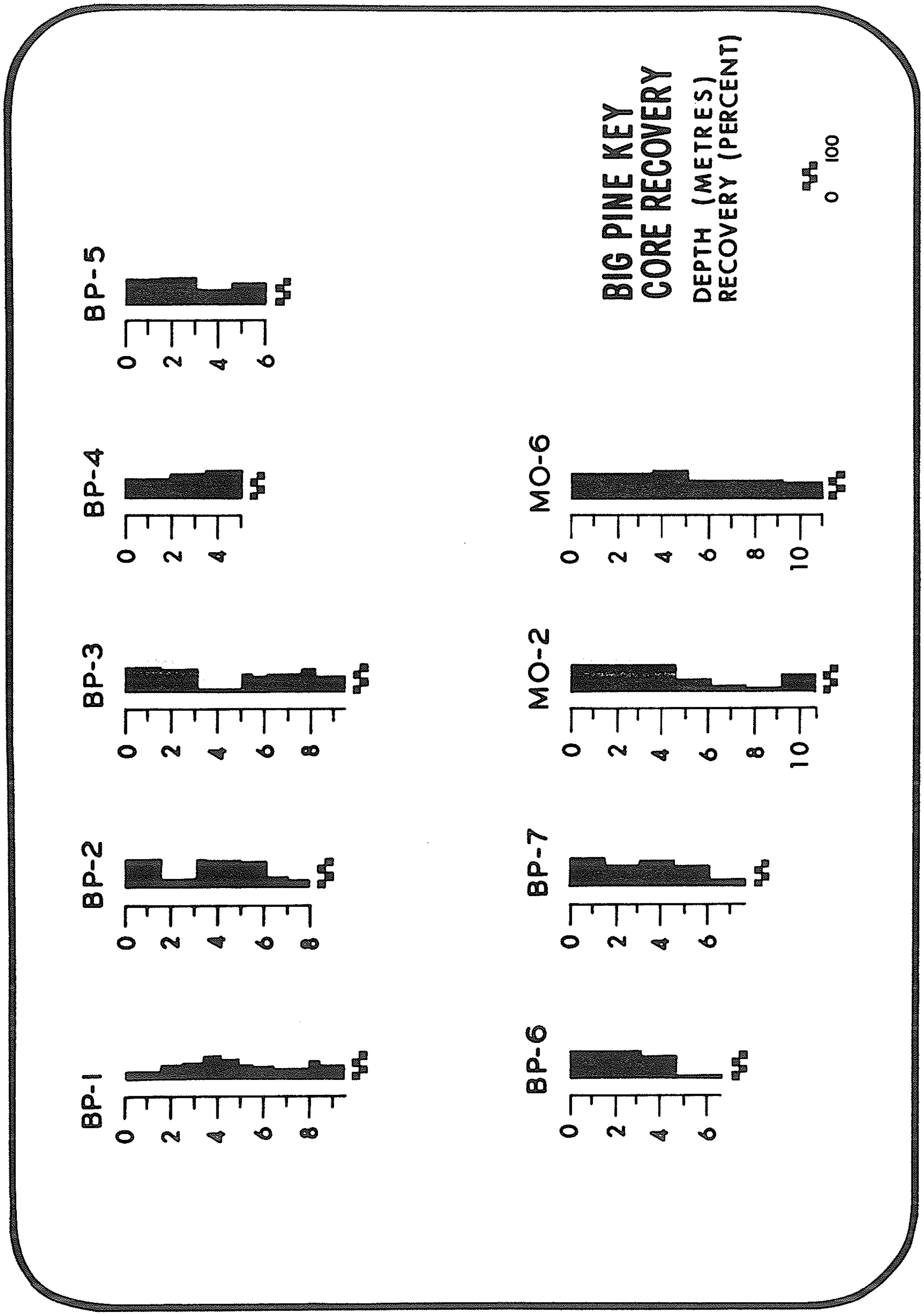

BIG PINE KEY
MIAMI LST
KEY LARGO LST
BP1
4,5,6
BP2
22
BP4
19
21
20
BP3
1,2,3
0 0.2 km
14,15,16
17
BP7
MO-6
11,12,13
18
10
MO-2
BP5
BP6
7,8,9
N
HAND SPECIMEN
BOREHOLE
DRAINAGE DITCH
0 1.0 2.0 km

Figure 3-3. Core recovery.


BIG PINE KEY
CORE RECOVERY
DEPTH (METRES)
RECOVERY (PERCENT)
0 100
BP-1
BP-2
BP-3
BP-4
BP-5
BP-6
BP-7
MO-2
MO-6

these units.

DEPOSITIONAL FACIES

INTRODUCTION

Facies names are based on Dunham's (1962; Figure 3-4) classification of carbonate rocks because:

1. Predominant grain types are noted.
2. The classification reflects depositional texture by denoting grain and mud volume-fabric relationships.
3. It permits naming of rock type using a minimal amount of component volume data.

The last point mentioned is an important factor because the weight of the effects of diagenesis on the rock name is subdued. This is especially significant because, as will be seen in Chapter 6, the observed mud content is probably less than that which was present initially. Evidence for dissolution and/or neomorphism of micrite to microspar is abundant, especially in those rocks which are mud-rich. The combination textural designation 'grainstone to packstone' reflects the sediment variability, typically present even on a thin section scale, and also allows for some diagenetic effects. The first part of the combination term reflects the rock type that appears to be predominant. Grain size classifications are based on Folk (1974) using the Wentworth size class terminology (Table 3-1).

OOID GRAINSTONE FACIES

Components: Included as ooids in this facies are superficial ooids and

Table 3-1. Grain size classes (modified from Folk, 1974).

Figure 3-4. Classification of carbonate sediments according to depositional texture (modified from Dunham, 1962).

	(MILLIMETRES)
	4
GRANULE	
	2
VERY COARSE SAND	
	1
COARSE SAND	
	0.50
MEDIUM SAND	
	0.25
FINE SAND	
	0.125
VERY FINE SAND	
	0.0625
SILT	
	0.0039
CLAY	

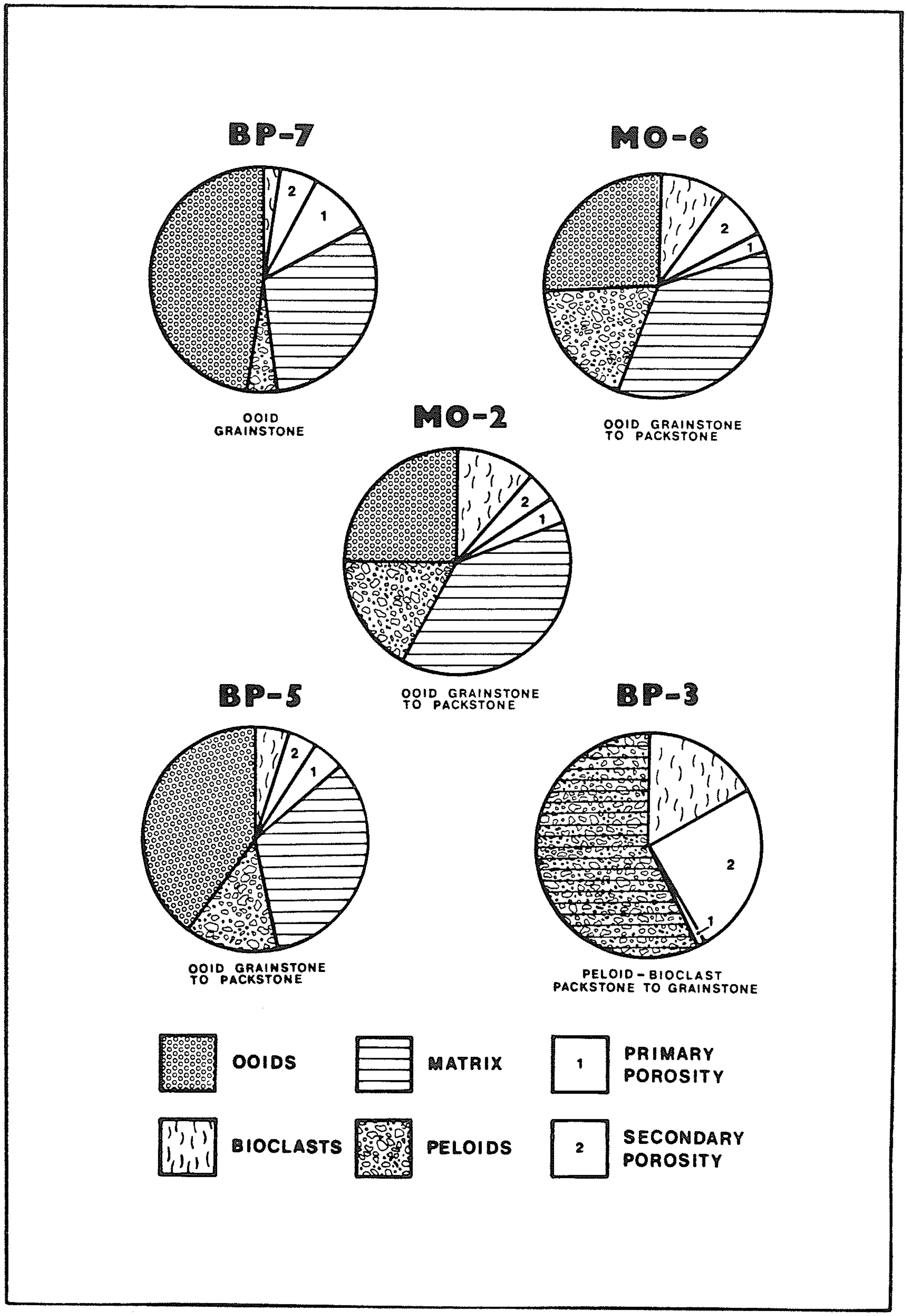
<table>
<tr><td colspan="5">DEPOSITIONAL TEXTURE RECOGNIZABLE</td><td rowspan="5">DEPOSITIONAL TEXTURE NOT RECOGNIZABLE</td></tr>
<tr><td colspan="4">ORIGINAL COMPONENTS NOT BOUND TOGETHER DURING DEPOSITION</td><td rowspan="4">ORIGINAL COMPONENTS BOUND TOGETHER</td></tr>
<tr><td colspan="3">CONTAINS MUD</td><td rowspan="3">LACKS MUD AND IS GRAIN-SUPPORTED</td></tr>
<tr><td colspan="2">MUD-SUPPORTED</td><td rowspan="2">GRAIN-SUPPORTED</td></tr>
<tr><td>GRAINS
<10%</td><td>GRAINS
>10%</td></tr>
<tr><td>MUD-STONE</td><td>WACKE-STONE</td><td>PACK-STONE</td><td>GRAIN-STONE</td><td>BOUND-STONE</td><td>CRYSTALLINE CARBONATE</td></tr>
</table>

multiple ooids. Subordinate amounts of uncoated pellets, peloids, and fossil debris also occur (Figure 3-5). The term 'pellets' is used to designate ellipsoidal to spherical cryptocrystalline grains of fine to medium sand size which are thought to be fecal pellets from mud-ingesting organisms. The term 'peloids' designates cryptocrystalline grains whose origin cannot be ascertained. Such peloids may be highly altered grains such as ooids, pellets, and bioclasts, or they may be diagenetic in origin. Shape is usually spherical to subspherical, although it may be highly irregular. Size typically ranges from silt to fine sand.

Average grain size lies within the medium sand range (Plate 1-3), although areas of coarse to very coarse sand size occur, and have correspondingly larger ooid nuclei (Plate 1-5). The predominant ooid nuclei are micritic ellipsoidal grains in the same size range as the ooids and are interpreted as fecal pellets. Less common ooid nuclei are fossil debris (notably mollusc fragments), ooids (to form multiple ooids), and rarely subangular quartz grains. Departure from ooid sphericity is notable in the larger ooid size fractions, and results from nucleation about larger, more irregularly shaped fossil nuclei. Moderately well sorted fossil-rich horizons containing mollusc and <u>Halimeda</u> fragments are present (Plate 1-4).

<u>Sedimentary Structures</u>: Horizontal to sub-horizontal laminae of medium and coarse grained ooid grainstone are observable in several cores. Lamina thickness varies from a few millimetres up to approximately a centimetre. Boundaries between laminae of different grain size are typically gradational. Some laminae appear notably disturbed, possibly the

Figure 3-5. Average facies compositions of the Q5 unit at Big Pine Key. See Appendix C for details.


BP-7
2
1
OOID
GRAINSTONE
MO-6
2
1
OOID GRAINSTONE
TO PACKSTONE
MO-2
2
1
OOID GRAINSTONE
TO PACKSTONE
BP-5
2
1
OOID GRAINSTONE
TO PACKSTONE
BP-3
2
1
PELOID - BIOCLAST
PACKSTONE TO GRAINSTONE
OOIDS
MATRIX
1
PRIMARY POROSITY
BIOCLASTS
PELOIDS
2
SECONDARY POROSITY

result of sediment mixing by burrowing organisms.

Cross-bedding is detected throughout most of the grainstone in borehole BP-7 and is observed in 3 related ways (Plates 1-1, 2):

1. Alternating cross-laminae of varying grain size are at angles up to 20-25° from horizontal.
2. Channel and vug porosity frequently follows the above.
3. Fracturing of the core is parallel to the cross-laminae and porosity.

Interpretation: Ooids require agitation for their formation and growth (see Bathurst, 1976, p. 302). However, it is uncertain whether the site of deposition represents the actual site of ooid formation. Harris' (1977) work on the Joulters Cays ooid sand shoal clearly shows the relationship between the site of ooid formation and the site of deposition. Ooids are formed in mobile, linear belts whereas accumulations are much more laterally extensive.

The high degree of sorting, lack of mud, relative absence of effects from burrowers, and cross-bedding suggest that this facies represents deposition in a high energy environment.

OOID GRAINSTONE TO PACKSTONE FACIES

Components: This facies is characterized by a high degree of vertical and lateral variability. Generally, the sediments consist of fine to medium sand size ooids with varying quantities of pellets, peloids, and fossil debris (Figure 3-5; Plates 2-1; 3-1). Local grain size variations occur with average grain size being coarse silt. In thin section the sediment is rarely either completely grainstone or packstone, although the portion of this facies within borehole BP-7 is mud-free. Commonly, areas of grainstone grade into packstone with no change other than mud content.

Sorting is generally moderate, becoming increasingly poorer with increasing mud content. Based on point count data, mud content typically falls in the 10-25% volume range.

Bioclast content of this facies is variable. Bioclast-rich intervals (boreholes BP-7 and MO-6) from several centimetres to several tens of centimetres occur (Plates 2-4; 3-2). These intervals are usually poorly sorted with a bimodal grain size distribution of very coarse sand to granule size Halimeda plates and mollusc fragments, and fine to medium sand size ooids, pellets, and peloids.

Very coarse sand to granule size Halimeda plates and mollusc shell fragments, both of which may be coated, occur ubiquitously. Large bivalve shells, echinoderm fragments (both up to 5 cm in size), and more rarely encrusting bryozoans (Schizoporella floridana) occur randomly (Plates 2-2, 3, 4).

Sedimentary Structures: Evidence of burrowing is common. Burrow infills tend to be coarser, somewhat better sorted, and less cemented than the surrounding sediment (Plate 2-5). Irregular areas of varying grain size or mud content also suggest burrowing. Mud-lined burrows are common (Plate 3-5).

Interpretation: Although ooids constitute up to 40% of the rock by volume (Figure 3-5, borehole BP-5), the moderate sorting, abundance of mud, and obvious effects of burrowing suggest that the facies represents deposition in a less energetic environment than the well sorted ooid grainstone (below wave base or sheltered behind some barrier).

PELOID-BIOCLAST PACKSTONE TO GRAINSTONE FACIES

Components: Generally, the facies is a poorly sorted, silt to medium sand size peloid-bioclast packstone to grainstone (Figure 3-5; Plates 4-1; 5-1, 2; 6-1; 8-1, 2, 3). Grains vary from coarse silt to very fine sand size peloids, to fine to medium sand size pellets, to granule size to several centimetre bioclasts (typically *Halimeda* or molluscs). Branching corals (*Porites* sp.) and head corals (*Montastrea annularis*) are also common, some several centimetres or more in size (Plates 4-5; 6-2, 4; 7-1, 2).

Much of the packstone so identified may in fact be an agglomeration of silt and very fine sand size peloids which, in the extreme, merge to give the appearance of a featureless muddy matrix (Plates 5-3, 4). In less extreme cases, a clotted 'structure grumuleuse' texture may be seen. This fabric is discussed in Chapter 6.

Sedimentary Structures: Burrows, with or without mud linings, are common. Burrowing is also inferred to account for some of the mixed packstone and grainstone areas.

Interpretation: Of the three facies recognized, the peloid-bioclast packstone to grainstone represents the lowest energy deposit. This deduction is based on the poor overall sorting, the large amounts of mud, and a wide assortment of relatively intact bioclasts, all suggesting deposition in a relatively quiet environment (below normal wave base or in sheltered locations).

STRATIGRAPHY OF BIG PINE KEY

INTRODUCTION

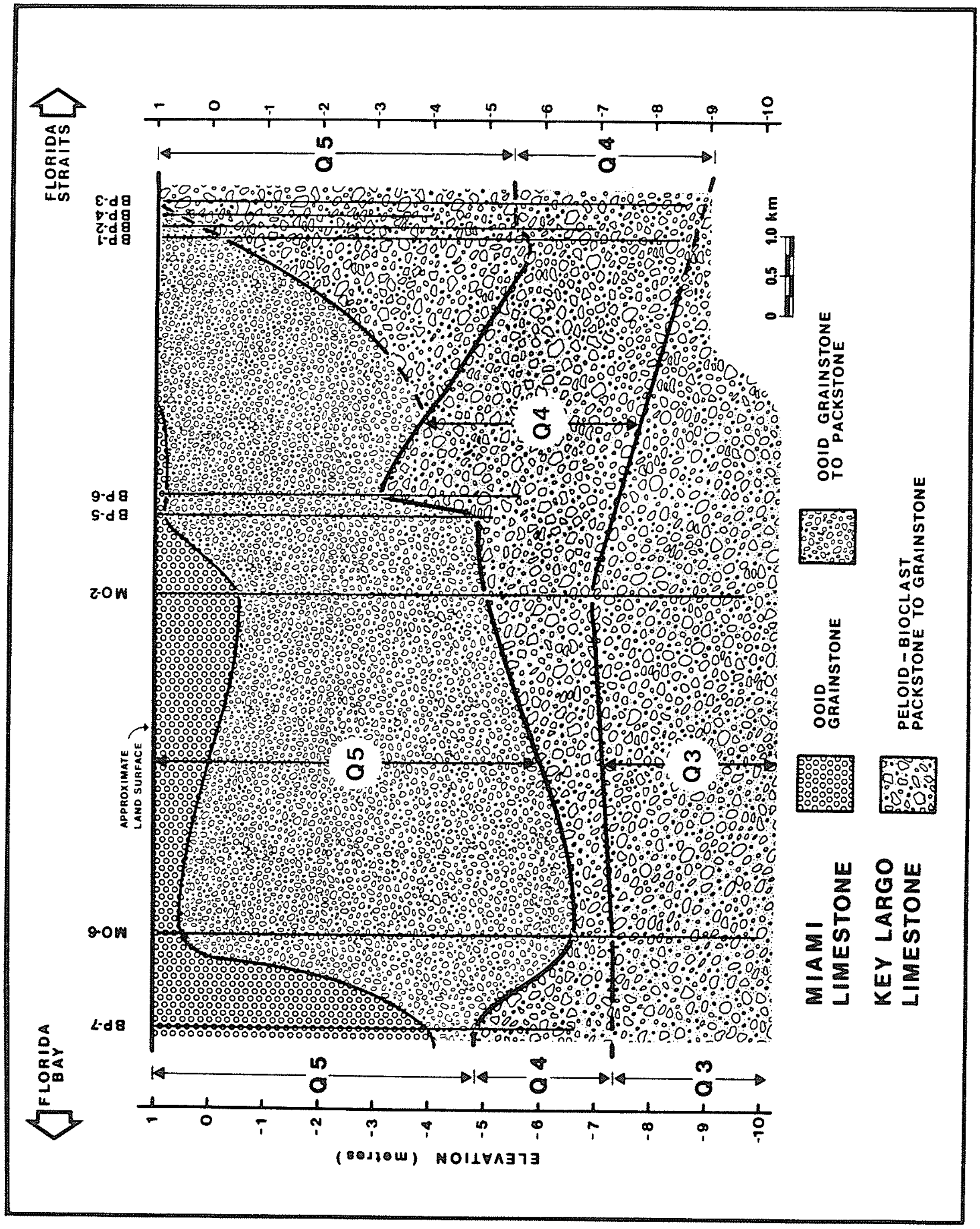
Nine boreholes on Big Pine Key, defining a section perpendicular to depositional strike, permit the recognition of the 3 major facies discussed in the previous section and of 3 stratigraphic units which correspond to the Q5, Q4, and Q3 units of Perkins (1977, Figure 3-6). The stratigraphic units are separated by distinct subaerial exposure surface zones (caliche profiles; discussed in Chapter 5).

On Big Pine Key, the Q3 and Q4 units consist of only 1 major facies - a peloid-bioclast packstone to grainstone facies, considered to be Key Largo Limestone (Figure 3-6). Within the Q5 unit, 3 major facies are recognized. From northwest to southeast, these are an ooid grainstone facies, a transitional facies, and a peloid-bioclast packstone to grainstone facies. The grainstone and transitional facies constitute the Miami Limestone, whereas the peloid-bioclast packstone to grainstone facies is Key Largo Limestone. Pre-Q5 unit topographic relief (Q4 unit surface) apparently varies randomly with no indication of seaward dip.

Q3 UNIT

Facies Variations: The main lithology of the Q3 unit on Big Pine Key is a peloid-bioclast packstone to grainstone. The lithologic and diagenetic

Figure 3-6. Facies distribution and stratigraphy of Big Pine Key. The units are separated by distinct discontinuity surfaces. Refer to Figure 3-1 for borehole locations.



FLORIDA BAY
FLORIDA STRAITS
ELEVATION (metres)
BP-7
MO-6
APPROXIMATE LAND SURFACE
MO-2
BP-5
BP-6
BP-1
BP-2
BP-4
BP-3
Q5
Q4
Q3
0 0.5 1.0 km
MIAMI LIMESTONE
KEY LARGO LIMESTONE
OOID GRAINSTONE
OOID GRAINSTONE TO PACKSTONE
PELOID-BIOCLAST PACKSTONE TO GRAINSTONE

similarity between the Q3 unit and the overlying Q4 unit renders the discontinuity surface which separates them less distinct than that which separates the lithologically and diagenetically dissimilar Q5 and Q4 units. This unit is observed in only two boreholes (MO-6 and MO-2) and, based on available control, has a minimum thickness of 2.7 m (borehole MO-2, Figure 3-6). The lower boundary of this unit is not penetrated.

From northwest to southeast, a noticable increase in abundance of large mollusc shell fragments (several centimetres) occurs (Plate 7-3). Halimeda-rich intervals are prominent in the northwest (borehole MO-6). Porites sp. fragments of several centimetres up to 15 cm are common throughout the facies (Plates 7-1, 2), however, evidence of Montastrea annularis is present only in borehole MO-6. Unidentified comminuted fossil debris of coarse silt to fine sand size is abundant. The nature of many bioclasts up to medium sand size is not known due to lack of microstructure (Plate 8-4). Typically, only the micrite envelope remains to identify a former bioclast precursor as being distinct from a sparry matrix. Encrusting coralline algae occur locally and fragments up to a couple of centimetres are common.

Interpretation: The limited control for this facies shows that coral debris is common under most of Big Pine Key. Most of the coral debris is Porites sp., and only one piece of Montastrea annularis was found. The abundance of the branching corals and mud suggests low energy conditions, possibly those of the immediate back-reef. This is supported by Perkins' (1977, plate 3) borehole at the southeast point of Big Pine Key which demonstrates that head corals are abundant in the Q3 unit at this location. Perhaps, the main reef existed here.

Q4 UNIT

Facies Variations: On Big Pine Key, this unit is capped by a distinct subaerial surface zone which allows straightforward separation from the overlying Q5 unit. A minimum Q4 unit thickness of 0.7 m is observed in borehole MO-6 (Figure 3-6). Maximum thickness of the unit is greater than 3 m based on evidence from borehole BP-3. The lower boundary of this unit is observed in only two boreholes (MO-6 and MO-2) and is demarcated by the Q3 unit subaerial surface zone.

The lithology of the Q4 unit is primarily a peloid-bioclast packstone to grainstone. However, because of its relative thinness, a subaerial overprint (pervasive micritization) exists within much of the available rock and delineation of original sedimentary characteristics is difficult. Also, because of the thinness, establishment of lateral continuity of the variations present within this facies is not certain. These two factors result in the recognition of only one major facies with two laterally limited variations.

From northwest to southeast there is a general reduction in mud and peloid content. Unidentifiable bioclastic debris of silt to very fine sand size occurs in some areas. Within this facies, bioclast-rich areas consist of *Halimeda* plates and/or peneroplid and miliolid foraminifera (Plate 6-3). These occur without recognizable pattern or trend. To the southeast there is an increase in the coarse sand to granule size fraction (or larger) of coral, coralline algae, encrusting foraminifera, echinoderm, and mollusc debris (Plates 6-1, 2, 4). At the southeast point, fossil fragments up to several centimetres in size are common. In the northwestern part of the island, Porites sp. fragments are rare but they increase in number and in size progressing to the southeast. In the southeast, cored head corals (Montastrea annularis and Diploria sp.) up to several tens of centimetres occur. Based on evidence in borehole BP-3 coral abundance increases with depth.

The most notable variations of the facies on Big Pine Key are as follows. In borehole BP-7, most of the available Q4 unit comprises a poorly sorted peloid-pellet-ooid packstone to grainstone (Plate 6-5). Pellets range in size from fine to medium sand whereas ooids range from medium to coarse sand size. As with this facies in general, coarse silt to very fine sand size peloids are abundant. Local wackestone with very fine sand size bioclastic debris also occurs.

The other facies variation occurs in borehole BP-1 and is at least 2 m in thickness. The lithology consists of a moderately sorted to poorly sorted peloid-bioclast grainstone with subordinate packstone. In the uppermost portion of the unit, bioclastic debris comprises most of the grains but peloids (and pellets) become more abundant with depth. Grain size ranges widely from coarse silt size peloids to granule size bioclastic debris, but average grain size is fine to medium sand. The majority of identifiable bioclastic debris consists of intact and fragmented foraminifera (peneroplids and miliolids largely; Plate 6-6). Articulated coralline algae and mollusc debris is scarce in the uppermost unit, but increases in abundance with depth. Local packstone occurs where peloids appear to merge, but the majority of interparticle space is filled with a relatively clear spar suggesting rock fabric is dominantly grainstone.

Interpretation: According to Perkins (1977) the site of the present-day lower Keys during Q4 time did not see the development of the ooid tidal bar belt as was the case during Q5 time. The lithology of the Q4 unit of Big Pine Key supports this. Perkins (1977, plate 1, figure 18) shows that the peloidal grainstones and packstones ('bryozoan facies' of Hoffmeister et al., 1967, p. 178) occupies most of the area under the lower Keys. The high content of bioclasts within the Q4 unit of Big Pine Key resembles the Key Largo Limestone lithology of the Q5 unit of this study. Based on this, the entire Q4 unit at Big Pine Key is considered to be Key Largo Limestone, rather than Miami Limestone (Figure 3-6). The sediments of the Q4 unit suggest a low energy depositional environment.

Interestingly, ooids in the Q4 unit sediments of borehole BP-7 suggest that they were forming during Q4 time. However, the mud-rich matrix within which the ooids occur suggest that the site of formation was not nearby (possibly further north in the Gulf of Mexico). This may also be due to the effects of sediment mixing by burrowing organisms.

The dominantly peloid-bioclast grainstone interval found in borehole BP-1 is interesting because to the northwest and southeast, mud content increases and the rocks are typically packstones. This occurence of grainstone suggests a tidal channel deposit. The channel may have connected the fore-reef waters with those of the back-reef. The characteristically strong currents in such channels may explain the apparent lack of mud.

Q5 UNIT

Ooid Grainstone Facies Variations: In cross-section, the ooid grainstone facies is thickest in the northwestern parts of the island and evidence from borehole BP-7 suggests a maximum thickness of 5 m (Figure 3-6). Southeastward, the facies quickly diminishes to 0.5 m in borehole MO-6 with a slight increase in borehole MO-2. The facies pinches out in the southwestern part of Big Pine Key. Within the facies, sorting progressively decreases in a southeastern direction and distinction from the underlying facies becomes more difficult. This is partly the result of the transitional nature of the facies boundary, and partly due to the near-surface diagenetic effects which obscure primary depositional detail.

Ooid Grainstone to Packstone Facies Variations: Volumetrically, this facies is the most important of the Q5 unit of Big Pine Key. A maximum thickness of 7 m is present in borehole MO-6 (Figure 3-6). Whereas sorting is generally moderate, there is a tendency toward decreased sorting with depth and to the southeast. The latter trend typically coincides with an increase in mud content. Boreholes in the northwestern part of Big Pine Key show an increase in bioclast content with depth (Plate 2-4).

Coated, ooid-like coral grains are extremely rare in the northwestern portion of the island, but coral grains up to a few millimetres in size occur to the southeast. Coarse sand size grains of serpulid worm tubes and encrusting foraminifera (*Homotrema rubrum*) also become more abundant.

A true packstone exists in borehole MO-6 from approximately -1 to -3 m depth, but lateral continuity is not observed elsewhere (Plates 3-3, 4). Within this packstone interval smaller areas of wackestone occur.

The lowermost few centimetres of this facies directly overlying the Q4 unit subaerial surface may contain clasts up to a few centimetres in size. These clasts are thought to have been derived from the Q4 unit surface. They are considerably darker in colour than the sediment in which they are found, and are usually well cemented and non-porous.

The facies grades vertically and to the northwest into the ooid grainstone facies, and southeast into the peloid-bioclast packstone to grainstone facies (Figure 3-6).

<u>Peloid-Bioclast Packstone to Grainstone Facies Variations</u>: Within the Q5 unit, this facies is limited to the southeast point of Big Pine Key (Figure 3-6). The facies grades into the southwesternmost Q5 Key Largo Limestone. The facies is thickest in borehole BP-3 where it comprises the total Q5 unit and is approximately 6.5 m thick.

Two general trends can be delineated in this facies. The most obvious is the transitional change from the Miami Limestone of the main portion of Big Pine Key to the Key Largo Limestone in the southeast. From borehole BP-1 to borehole BP-3 (Figure 3-6), a decrease in sorting accompanied by an increase in mud content occurs. The lithology at borehole BP-3 is a very poorly sorted peloid-bioclast packstone, in which individual peloids are rarely discernable (Figure 3-5, borehole BP-3). Fossil debris size may be as large as several centimetres, and an increase in the amount of coral (<u>Porites</u> sp. and <u>Montastrea annularis</u>), encrusting forams (<u>Homotrema rubrum</u>), mollusc, encrusting coralline algae, echinoderm, and serpulid worm tube debris occurs to the southeast

(Plates 4-2, 4, 5; 5-5). Corals and large mollusc shell fragments may show signs of boring. In this trend, the appearance of large (up to 1 cm) tubular-shelled mollusc (?) debris occurs. Granule size Halimeda plates are common and may be quite abundant locally (Plates 5-3, 4).

A less obvious trend involves a slight sorting improvement with depth from poor sorting to moderate sorting with a concomittant decrease in coarse bioclast and mud content (Plate 4-3). The trend is readily observed in the two southeasternmost wells.

Interpretation: Generally, the Q5 unit lithology of Big Pine Key is characterized by a southeastward decrease in sorting and ooid content and concomittant increases in mud and bioclasts (especially corals and coralline algae). The surface contact of the Miami and Key Largo Limestones as defined by Hoffmeister and Multer (1964, p. 60) is approximately coincident with the transition from the moderately sorted ooid grainstone to packstone to the peloid-bioclast packstone to grainstone facies (Figures 3-2, 6). As was seen in previous discussions, the transitional contact occurs on the surface between boreholes BP-3 and BP-4 (Figure 3-2). The northward dipping contact is in accord with observations of previous workers (see Hoffmeister and Multer, 1964, p. 60).

By the facies designation used in this study, the very well sorted ooid grainstone is restricted to the northern part of Big Pine Key and is only present in any significant amount in borehole BP-7. Most of the Miami Limestone which comprises Big Pine Key consists of the moderately sorted ooid grainstone to packstone which is typically highly burrowed and varies greatly in degree of sorting. The southeastern trend of increasing bioclast content, especially corals, approaches, but is still short of the 30% coral content cited by Stanley (1966, p. 1930) for the Key Largo Limestone of the Key Largo Waterway and the Windley Key quarry. There is, however, a dramatic increase in coral content with each successive borehole southward at the southeast point of Big Pine Key.

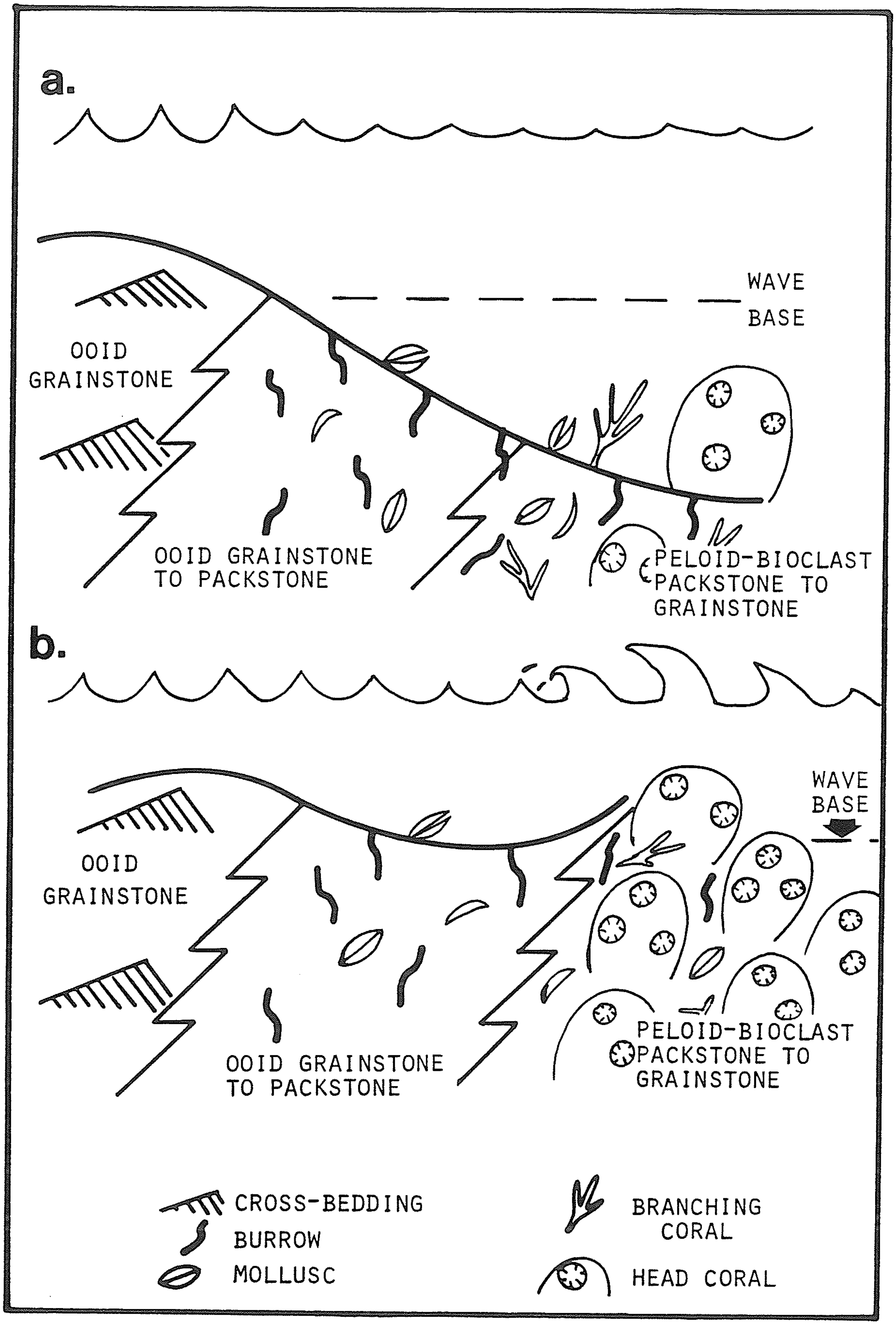
The facies patterns observed on Big Pine Key provide implications for wave energy and water depth (Figure 3-7). The ooid grainstones represent deposition in a high energy, above wave base setting. Seaward, oolitic sands are deposited together with mud. Further seaward, the peloid-bioclast packstone to grainstone may represent a low energy setting at the same depth or slightly deeper than the ooid grainstone to packstone , or it may represent deposits which resulted from the sheltering effects of a reef framework, which itself may be close to or above wave base.

Although the situation at Big Pine Key is not straightforward with regard to its depositional setting, it would be interesting to see if the other lower Keys show a growth history and stratigraphy similar to that of the Joulters Cays sand shoal (Harris, 1977). In this shoal, ooid grainstone (actually a loose sediment) occurs as a thin mobile fringe on the seaward side of the shoal. Earlier in the Holocene transgression, ooid grainstone was much more extensive. However, syndepositional relief due to the growing shoal gradually restricted circulation and considerably reduced wave energy over most of the shoal. Consequently, bioturbation by various organisms led to the mixing of the ooid grainstone with the underlying muddier sediments. These muddy sediments were deposited

Figure 3-7. Possible paleoenvironmental settings for the deposition of Big Pine Key Q5 unit sediments.

a: Ooid grainstone deposited in a high energy setting whereas ooid grainstone to packstone and peloid-bioclast packstone to grainstone deposited in a low energy setting.

b: Ooid grainstone deposited in a high energy setting but the ooid grainstone to packstone and the peloid-bioclast packstone to grainstone accumulated in a sheltered environment.


a.
WAVE
BASE
OOID
GRAINSTONE
OOID GRAINSTONE
TO PACKSTONE
PELOID-BIOCLAST
PACKSTONE TO
GRAINSTONE
b.
WAVE
BASE
OOID
GRAINSTONE
OOID GRAINSTONE
TO PACKSTONE
PELOID-BIOCLAST
PACKSTONE TO
GRAINSTONE
CROSS-BEDDING
BURROW
MOLLUSC
BRANCHING
CORAL
HEAD CORAL

earlier during the Holocene transgression when sea level was very shallow over the Pleistocene bedrock floor of the area and wave energy did not travel far. As a result, most of the upper portions of the shoal comprise an ooid packstone with ooid grainstone being significantly less abundant. Stratigraphically, the shoal clearly shows the increase in grain size and ooid content and decrease in mud content with increasing elevation in the shoal.

CHAPTER 4 - SUBMARINE DIAGENESIS

SUBMARINE CEMENTATION

INTRODUCTION: SUBMARINE AND BEACHROCK CEMENTATION

The literature of the last decade abounds with descriptions and interpretations of submarine and beachrock cementation of ancient and modern sediments (e.g. see Bricker, 1971; numerous references in Bathurst, 1976, and Milliman, 1974). Submarine cements are those which have been precipitated from seawater in a subtidal environment, whereas beachrock cements have been precipitated from seawater in a beach or intertidal environment.

The supersaturation of the ocean with respect to various calcium carbonate phases and the inorganic precipitation of high Mg-calcite and aragonite as cements have been studied intensely and in detail from physiochemical and biochemical viewpoints (refer to Bathurst, 1976, Chapter 6, for a comprehensive discussion of the subject). Mg-calcite has greater than 4 mole % $MgCO_3$ in solid solution (Chave et al., 1962, p. 33). The term 'calcite' is used in this work to designate calcite that has 0-4 mole % $MgCO_3$ in solid solution.

With respect to morphology and mineralogy, submarine and beachrock cementation is very similar microscopically, and the two environments cannot be discerned solely on the basis of this evidence (Bricker, 1971, p. 47). Cement morphology is variable and includes fibrous (acicular) and bladed crystals (e.g. Taylor and Illing, 1969, p. 94; James et al., 1976, p. 536) and microcrystalline crusts (e.g. Moore, 1971, p. 9). Milliman (1974, p. 274) also described a disk-like aggregate within Halimeda utricles and grapestones.

The nature of cementation and, indeed, whether it occurs at all is controlled by many physical and chemical factors (e.g. Bathurst, 1976). For example, cementation may be highly fabric selective and related to grain size (e.g. Shinn, 1969, p. 122) or pore size and type (e.g. James et al., 1976, p. 542).

The time required for submarine or beachrock cementation depends on many factors, and perhaps by inference it may be quite variable. In the Persian Gulf, Shinn (1969, p. 112) reported artifacts believed to have been cemented during the 20 years prior to his observation. Dravis (1979, p. 204) reported that cementation of the oolitic sands he studied in the Bahamas can take place within a few months or less.

SUBMARINE CEMENTATION OF BIG PINE KEY SEDIMENTS

Introduction: Submarine cement is present in the sediments of Big Pine Key, although it is only common in the Q5 unit sediments. Cement occurs both as interparticle and, much more commonly as intraparticle (intraskeletal) primary pore fillings. It is therefore volumetrically insignificant as a binding agent.

Submarine cement crystals vary from a few micrometres long by less than 1 μm wide to as large as 100-150 μm long by 5-10 μm wide. The majority of interparticle cement crystals are approximately 5 μm long, whereas most intraparticle (intraskeletal) crystals are 30-50 μm long. The largest submarine cement crystals observed occur in the intraparticle pores of corals.

Submarine cement occurs as isopachous fringes of fibrous crystals oriented perpendicular to the substrate (Plates 9-1, 2). The fringes vary from a few micrometres to 150 μm wide and single crystals occasionally constitute the entire fringe.

Intraparticle Cement: Scanning electron microscope (SEM) examination of Q5 unit cements shows that the submarine cements are tabular in nature, and crystal terminations vary from needle-like to scalenohedral to square (blunt or sharp corners; Plates 9-4 to 7). Commonly, crystals with different terminations occur within the same pore, but one type is usually predominant. Taylor and Illing (1971, fig. 18D) observed similar co-occurrences of differing terminations in Persian Gulf beachrock cements.

The nature of submarine cements occurring in intraparticle pores was observed within a coral specimen of the Q5 unit. In some pores, cement consists of aggregates of 3 to 5 acicular to tabular crystals (composite needles of Schroeder, 1973, p. 184) 40-70 μm long by 1-4 μm wide with scalenohedral terminations (Plates 9-5, 6). In other pores the cement occurs as individual crystals with either square or scalenohedral terminations (Plate 9-4). Petrographic examination shows syntaxial overgrowth of coral aragonite in some cases (Plate 9-3), and hence an aragonite mineralogy for these cements is inferred. This is further substantiated by staining with Feigl's solution and SEM energy dispersive analysis which shows above-background strontium content.

Substrate control is thought to result in the parallel and radiating fibrous aggregates observed in coral and other intraskeletal pores (Plates 9-2, 3, 4). Further suggestion of substrate control (i.e. syntaxial overgrowth) is provided by a marked orientation of some of the cement crystals (cf. Schroeder, 1973, fig. 3a). Occasionally, submarine cements grow from various internal sediments within intraparticle pores.

Schroeder (1973, p. 184) observed similar aragonite submarine cements in Pleistocene reef rocks from Bermuda and suggested that composite needles were the result of intergrowth or twinning or both.

Besides corals, submarine cement occurs in the intraskeletal pore spaces of foraminifera, pelecypods, gastropods, Halimeda (within utricles), and within microborings (discussed later).

Pore space may be partially or completely occluded by submarine cement. When partial, remaining pore space may be empty, or a sparry mosaic occludes the remainder of the pore (Plate 9-2). When completely occluded, the fringes having grown into each other define an irregular suture along their contact.

Intraparticle (intraskeletal) cements are common throughout the sediments and tend to increase in abundance as bioclast content increases. Many of the smaller pelecypods (ostracodes?) that contain intraparticle cement also have oolitic coats. The common occurrence of intraparticle submarine cement and scarcity of interparticle cement suggests that the protected micro-environment in which precipitation occurs can exist even within an apparently largely uncemented carbonate sand shoal. Taylor and Illing (1969, p. 80) made a similar conclusion concerning Persian Gulf sands. Staining of several samples of intraparticle cements suggests aragonitic mineralogy.

Interparticle Cement: Interparticle cement occurs as relatively isopachous fringes which may be very unevenly distributed on a macroscopic to microscopic scale. The small size of the fibrous crystals imparts a brown color to the fringe. The small crystal size precluded the use of Feigl's solution in order to determine mineralogy. An aragonitic mineralogy is inferred by analogy with intraparticle cements of Big Pine Key.

The cement is most clearly seen in grainstones (Plate 9-1). The nuclei of multiple ooids are commonly bound with such cement. Presumably, this interparticle cement allowed the grains to act as a single nucleus for the multiple oolitic coat. It is not known to what extent the mud-rich lithologies have been cemented in the submarine environment (i.e. within interparticle pores).

ENDOLITH MICROBORINGS AND MICRITIZATION IN THE SUBMARINE ENVIRONMENT

INTRODUCTION

Micritization due to boring endoliths is extremely common in ancient and modern shallow water carbonate sediments. Bathurst (1976, p. 388) described micritization as a three step process:

1. Boring and colonization by algae.
2. Death of the algae and vacation of the borings.
3. Precipitation (by processes unknown) of aragonite or Mg-calcite within the borings. Margolis and Rex (1971, p. 850) suggested that algal metabolism, boring activities, or bacterial action on decomposing algae produces a microchemical environment favourable to cement precipitation by raising the pH via the production of ammonia.

A dense array of such borings generates the micrite envelope which gradually centripetally replaces the grain. Complete micritization leads to the development of a micritic grain (peloid) with no recognizable features except perhaps its original geometry (Bathurst, 1976, p. 389).

Because the envelope is generally resistant to dissolution (organic and/or mineralogic interaction), it commonly survives destruction by dissolution whereas the grain may be totally dissolved (Winland, 1968, p. 1324). According to Bathurst (1976, p. 387), the micrite mineralogy is variable, but why this is so is uncertain. Substrate control is suggested by some studies (e.g. Kobluk and Risk, 1977, p. 1070) whereas others noted no relationship (e.g. Winland, 1968, p. 1323).

ENDOLITH MICROBORINGS AND MICRITIZATION OF OOIDS FROM BIG PINE KEY

The ooids of Big Pine Key appear to have undergone significant micritization in the marine environment. Evidence of microborings of various sizes and shapes are abundant within the ooid cortex. The evidence occurs either as aragonite molds of the microborings where surrounding cortex has been leached away (Plates 10-1 to 6), or as empty to partially submarine cement-filled spherical bores (Plate 10-2). These microborings range in size from 4-22 μm and are visible with SEM on ooid surfaces.

The aragonite molds typically consist of radiating to randomly oriented rods approximately 2-5 μm long by 0.1-0.2 μm in diameter (Plate 10-4). The aragonitic mineralogy of the molds was verified by Harris et al. (1979, p. 220) in a study of Pleistocene ooids from the Atlantic Coastal Ridge in south Florida. It was suggested that the radiating orientation of the rods is due to radiating crystals apparently growing outward into the boring from a collapsed algal sheath (Harris et al., 1979, p. 217). Some molds are partially composed of 2-7 μm rhombohedral spar thought to be a neomorphic replacement of the aragonitic mold (Plate 10-6).

The preferential solution of ooid cortex aragonite over the microboring infills (submarine precipitates) may reflect the effects of organic mucilagenous coatings associated with the borings (Margolis and Rex, 1971, p. 848), or it may reflect the preferential dissolution of finer grained cortex aragonite crystals (Harris et al., 1979, p. 220). The differential dissolution clearly demonstrates that most of the microborings follow the concentricity of the cortex layering. The microborings are also typically discontinuous and irregularly developed within a given layer (Plate 10-1). Occasionally, calcite spar cement may envelop the microborings within the altered ooid cortex (Plates 10-1, 15-7).

The microborings increase in abundance towards the periphery of the cortex. According to Margolis and Rex (1971, p. 847) the borings occur parallel to, but just below the grain surface. Harris et al. (1979, p. 217) also observed borings extending a few tens of micrometres into the grain. Margolis and Rex (1971, p. 850) suggested that preferential occupation by algae of certain laminae within the ooid cortex reflects a fluctuating environment. The tangentially oriented ooid cortex laminae are physiochemical precipitates deposited during periods unfavourable for algal growth. In contrast, the discontinuous non-oriented laminae reflect conditions where microboring was intensive.

Four categories or forms of microboring molds were encountered in the ooids from Big Pine Key:

1. 2 μm diameter spherical molds (rare).
2. 2-3 μm diameter filaments up to 35 μm long. Randomly oriented rods and sub-micrometre size equant crystals were observed in these molds (rare; Plate 10-3).
3. Spherical to irregular 10-65 μm molds. It is uncertain whether these are aggregates composed of smaller units or whether these are single units (abundant; Plates 10-5,6).
4. 7-10 μm diameter randomly oriented filaments up to 40 μm long (abundant; Plate 10-3).

The latter two types commonly occur in clusters where they may be overwhelmingly predominant over any other forms present. Although forms 3 and 4 suggest an algal origin based on information from cited literature, the origin of forms 1 and 2 is uncertain.

A discussion on the process which leads to the preservation of the ooid microborings is deferred to Chapter 6.

CHAPTER 5 - SUBAERIAL EXPOSURE SURFACE ZONE: PRODUCTS AND PROCESSES

INTRODUCTION

Discontinuity surfaces, also referred to as subaerial exposure surfaces, result when a carbonate rock or sediment is exposed subaerially. In the case of south Florida, subaerial exposure during Pleistocene time has been the result of eustatic sea level changes (Perkins, 1977, p. 137). The rock or sediment below the surface is typically altered and the exposure surface is commonly accompanied by a surficial calcareous crust or caliche profile. The term 'caliche' designates the secondary calcareous material that has accumulated as a Cca soil horizon or as beds beneath the soil (Harrison, 1977, p. 128).

The delineation of the stratigraphic units on Big Pine Key is based largely on the recognition of these subaerial exposure surface zones. Macroscopically, the exposure surface zone or caliche profile is recognized by colour or textural criteria. The exposure surface zone may be a dense, hard crust, which is typically laminated, or the surface zone may exist as areas of gray to red-brown colouration where individual grains may or may not be distinguishable (discussed later).

COMPONENTS AND MICROBORINGS OF THE SUBAERIAL EXPOSURE SURFACE ZONE

CEMENTS

Introduction: The upper vadose zone is most rapidly affected by prevailing meteoric conditions by virtue of its position; as depth increases the effects are progressively dampened (James, 1972, p. 830). This is the environment of the most chemically undersaturated fluids (rainwater) and also of the most highly supersaturated fluids (due to solution that is followed by rapid evaporation). There are extremes in temperature and humidity. The main effects of floral and faunal stabilization occur here; life and post-mortem processes act in modifying to various degrees the water chemistry. The high degrees of supersaturation attainable in this environment results in the precipitation of $CaCO_3$ in various morphologies which, in some cases, are very distinct from those precipitated in other diagenetic environments.

For a more complete discussion of possible chemical-kinetic and organic controls as they pertain to the subaerial surface environment, consult works by James (1972) and Harrison (1977).

Fibrous Calcite Cement: This cement occurs as fibrous crystal aggregates which range in length from 20-100 μm and are sub-micrometre in width. The aggregates occur as isopachous fringes (partial or complete), or as fan-like (radiating) botryoids which, in extreme cases, occlude pore space and leave irregular dark suture lines where the botryoids meet (Plate 12-1). The larger botryoidal aggregates show growth (cementation) episodes demarcated by concentric 'dust' lines.

Fibrous calcite cement occurs in the uppermost 1-2 cm of the surface zone and may line or occlude irregular vuggy or interparticle porosity. The cement may also co-occur in pores with a partial fringe of spar cement. Occurrence is sporadic. This cement was observed only within the Q5 surface zone.

Harrison (1977, p. 137) described 'bladed spar' from Barbados and included the 'flower spar' of James (1972, p. 826) as a variant. The fibrous spar of Big Pine Key is thought to be a further variant of 'bladed spar'.

The reasons for the morphology of such cements are uncertain. Conditions of extreme supersaturation and a high concentration of impurities in the solute calcium carbonate have been suggested as possible causes (see Harrison, 1977, p. 137).

Random Needle Fibres: These acicular crystals are commonly 1-2 μm wide by 20-40 μm long, but size varies from 0.3 μm by 10 μm to 20 μm by 40 μm (Plates 12-2; 16-7). The needles occur as a loosely packed random aggregate - "the arrangement is like that expected of a handful of toothpicks dropped into a container" (Supko, 1971, p. 143). In order to distinguish these crystals from those which occur in a tangential arrangement as a coating, James (1972, p. 825) coined the term 'random needle fibres'.

The aggregate may consist of a few crystals or thousands of crystals. There is a tendency for larger aggregates to be composed of smaller crystals. Optically, the crystals are predominantly length-slow in character, and, by analogy with similar crystals described by James (1972, p. 825), they consist of calcite.

Random needle fibres occur in primary interparticle pore space, but development in intraparticle, moldic, and vuggy pores is also common. In large vugs, the crystals are best developed at points of constriction or in sheltered embayments. In smaller pores, the aggregates may span the pore and, more rarely, individual crystals may span the pore. Commonly, the development of microspar at various points within the aggregate cements the delicate latticework.

The crystals are most common within 1-2 m of the subaerial surface, but crystals as far as 5 m below the surface have been observed (borehole MO-2). Within the Q5 unit, there is a marked tendency for the southeastern wells to have a more widespread and profuse development of the crystals, whereas the northwestern wells usually have insignificant quantities. Random needle fibres are rare in the Q4 unit and have not been observed in the Q3 unit.

Supko (1971, p. 144) originally forwarded a freshwater phreatic origin for the 'whisker crystals' he observed in a Bahamian calcarenite, based on the occurrence of the crystals within the present-day freshwater phreatic zone. James (1972, p. 830) suggested that intense evaporation and extreme supersaturation in the upper vadose zone, possibly along with water chemistry changes due to salt spray or soil processes, may be responsible for the extraordinary crystal morphology. Harrison (1977, p. 138), based on the very common confinement of the needle fibres to root voids in Barbados crusts, suggested that organic controls may be important. The high degree of supersaturation demanded for needle growth could be attained in the microenvironment adjacent to plant roots where the rate of removal of CO_2 from the pore waters may be enhanced over the rate of removal in other vadose pores. Ward (1970; cited in James, 1972, p. 831, and Harrison, 1977, p. 138) suggested that the crystals may have been precipitated along fungal hyphae. This was based on the similarity between the morphology of the crystals and hyphae.

In Chapter 7 it will be seen that most of the boreholes on the main portion of Big Pine Key are located in the present-day freshwater phreatic lenses. In contrast, boreholes of the southeast point of Big Pine Key are presently in the marine phreatic zone. That these crystals are not stable in the freshwater phreatic zone was suggested by Steinen (1974, p. 1015) in working with Pleistocene carbonates from Barbados. The observed distribution of random needle fibres suggests that the crystals are more stable in the marine environment than in the freshwater phreatic environment, even though they are precipitated in the vadose environment. The reasons for this are uncertain.

Micrite: Volumetrically, micrite is the most important component of caliche. A micrite cement and a micrite of uncertain origin are recognized. The cement occurs as 1-2 μm rhombohedral to irregular calcite crystals. More rarely developed are tangentially oriented, length-slow, needle-like crystals less than 1 μm wide by up to 3-4 μm long. These occur intimately with the more equant crystals. Their elongate nature can be observed in the ultra-thin feathered edges of thin sections.

In thin section, cement colour varies from medium grey (predominant) to red-brown to dark brown. The cement coats grains, intraskeletal pores, and organic filaments, and may define short stringers (Plates 12-6; 13-8; 14-1, 6). The cement is most easily distinguishable macroscopically or microscopically in grainstones, and becomes progressively less obvious with increasing mud content.

Whereas the micrite cement shows evidence of having been deposited as a coating, micrite of an uncertain origin also exists within the surface zone. Essentially identical to the micrite cement in most respects, the colouration of this enigmatic micrite is typically a medium grey-brown to red-brown. Gradation into microspar is common.

Areas of less intensive micritization may contain ghosts of former grains, inferring a neomorphic alteration process (see Chapter 6 introduction) such as that described by Kahle (1977, p. 421). In surface aggraded zones (i.e. an actual accumulation of micrite versus cannibalization of pre-existing rock or sediment) where no ghost evidence exists, the micrite may be a true cement precipitated in layers (Plate 12-3).

Harrison (1977, p. 137) suggested that the micrite in the caliche profile is a result of rapid precipitation from supersaturated solutions. Sequential episodes of this process causes the observed layering of the laminated surface crusts, or, in the case of coated grains, faint laminae. Rapid precipitation may be the result of intense evaporation in the upper vadose zone (Harrison, 1977, p. 137). The origin of the colour variations is deferred to later discussion.

In situ micritization, as distinct from cementation, may also be a major contributor to the micrite observed within the caliche profile. Harrison (1977, p. 134) and Kahle (1977, p. 424) described grains and calcite cement which have been micritized from Barbados and the lower Florida Keys, respectively. Kahle's (1977, p. 426) observations suggested that vadose micrite is largely the result of micritization and 'sparmicritization', the process whereby sparry calcite crystals are gradually altered to micrite by a concommitant dissolution-precipitation mechanism. The intimate relationship of sparmicritized rock with fungal and algal endoliths suggests that the processes of micritization and sparmicritization may be related to changes in water chemistry induced by the above.

AUTHIGENIC NON-CARBONATE MINERALS

Within the caliche of the Q5 unit, aggregates of a yellow microcrystalline mineral (up to 10 μm in size) occur as irregular patches of varying size and shape. At low magnification under crossed nicols, the aggregates go to extinction although this is apparently the result of small crystal size. Under high power there is a pinpoint extinction, similar to that observed in chert. The areas in which the aggregates occur are micrite-rich and typically within centimetres of the subaerial surface. When abundant, the aggregates impart a yellow-orange colour to the rock. Within the Q4 unit surface, a red-brown amorphous mineral commonly occurs as secondary pore linings along with sparry calcite usually occluding the remaining porosity. The above minerals were not analysed, but are thought to be ferric oxides or hydroxides.

PELOIDS AND CLOTTED TEXTURE

Peloids of dark brown to grey micrite commonly ranging in size from 30-60 μm are extremely common in the subaerial surface zone (Plate 12-4). The peloids vary in shape from circular to subanglar, and commonly merge with one another to form indistinct massive micrite areas. Commonly, these are identifiable if either pore space or a thin microspar rind is present in order to delineate them. These form what James (1972, p. 825) calls 'clotted' fabric.

The peloids are typically found in the interparticle (original sedimentary particle) spaces closely associated with micrite-coated organic filaments, coated grains, and subsurface micritic stringers. Based on the above, and because they gradually disappear as the rock progressively shows less micrite cement, these peloids are inferred to be diagenetic in origin. With depth below the surface there is a decrease in the number of peloids, although some have been observed to occur where no other surface related features are found. In such cases it is not possible to differentiate those peloids which are primary depositional (e.g. submarine micritized coral grains) from those which are related to subaerial diagenesis.

Harrison (1977, p. 133), James (1972, p. 823), and Harrison and Steinen (1978, p. 391-392) proposed that the peloids may be partially or totally micritized grains and the clotted texture is an integral part of the upper vadose (subaerial surface) zone. Although evidence for particles transitional between unaltered and completely micritized (peloid) was not observed in the samples from Big Pine Key, this may reflect the tendency for smaller grains to be preferentially micritized. This may be a function of grain type or some related phenomena. Just as significantly, the peloids may result from micrite precipitation around some unknown nucleus, and their irregular size, shape, and distribution may be a function of different types of nuclei.

In terms of neomorphism, some of the peloids may in fact be the remaining vestiges of a more micritic area much of which has recrystallized to microspar 'islanding' the micrite and thus producing 'peloids'. The opposite may occur where spar is locally altered to micrite by Kahle's (1977) sparmicritization process.

In summary, the origin of the peloids may be:

1. Primary depositional grains which have been altered post-depositionally. This is unlikely for the reasons stated above.

2. Micrite cement precipitated around tiny nuclei of unknown origin
3. Neomorphism (degrading) producing an 'island' effect (sparmicritization).
4. Neomorphism (aggrading) progressively converting micrite to microspar or spar, also producing an 'island' effect.

CALCIFIED FILAMENTS AND RELATED MICROBORINGS

Micrite-coated filaments (hereinafter termed calcified filaments) typically range from 10-20 μm in diameter and attain lengths up to 150 μm (Plates 13-3, 4, 5). They are straight to slightly arcuate, pinch and swell slightly in diameter along their length, and may branch. The calcified filaments consist of micrometre-size (1-2 μm) scalenohedral calcite, and the thickness of coating is approximately 2-3 μm (Plate 13-6). Within the surface zones of the units, large (40-80 μm diameter) calcified filaments occur which have numerous smaller (10-20 μm diameter) calcified filaments extending radially outward (Plates 13-1, 2).

Calcified filaments are found in the uppermost metre of the units and become progressively less numerous below the subaerial surface. They occur predominantly as chasmoliths, but evidence of grain boring exists (Plate 13-4). Micrite-rich areas with no relict texture may have abundant circular to elongate pores approximately the same size as the smaller calcified filaments (Plate 13-7). In areas of less severe alteration, calcified filaments may be observed within grain ghosts. The above suggests that at least minor grain cannibalization has occured. Calcified filaments commonly span interparticle and secondary void spaces, and minor amounts of micrite precipitation produces what Harrison (1977, pl. 9-1) described as 'wisps' which subdivide larger pores (Plate 14-3).

Previous workers have observed similar (10-20 μm) calcified filaments in various carbonate rocks. James (1972, p. 826) noted them in Barbados crusts and suggested they were the calcified sheaths of blue-green algae. Harrison (1977, pl. 8-6) observed irregular networks of similar filaments in a skeletal sand-rich caliche horizon of Barbados and suggested the network was the result of algae. Kahle (1977, p. 419) described various types of endolithic and chasmolithic filaments from the crusts of the lower Florida Keys and although taxonomic description was greatly impeded by insufficient morphological detail, some of those he described as being due to algae are similar to the calcified filaments observed in this study. Their development in a non-photic zone cannot be used to preclude an algal origin since some algae are adapted for low light environments (Kobluk and Kahle, 1978, p. 366).

Although the above studies strongly suggest an algal origin for the calcified filaments, work by Klappa (1979a) suggests that root hairs may also provide an explanation for these calcified filaments. He noted that green algal filaments are commonly 6-20 μm in diameter, vary in diameter along a single filament, branch ('false ramification'), and may be irregularly shaped. In contrast, root hairs are typically 5-17 μm in diameter, have a constant single filament diameter, rarely branch, and may be straight or curved (Klappa, 1979a, table 1). Ward (1970; cited in James, 1972, p. 826) made similar observations in Pleistocene calcarenites from Mexico and suggested they were derived from root hairs of angiosperm dune plants. At Big Pine Key the occurrence of the large 40-80 μm calcified filaments from which some of the smaller 10-20 μm calcified filaments branch strongly suggests a root hair origin. Their occurrence within tiny soil pockets ('soil' comprising a mat of calcified filaments, unidentified opaque and translucent material, and micrite) further supports a root hair origin for these particular calcified filaments.

The diagenetic effects of the organic activity now represented by calcified filaments are significant and are discussed in "Evolution of the Caliche Profiles at Big Pine Key" (this chapter).

Recent works (e.g. Kobluk and Risk, 1977; Dravis, 1979) have described synsedimentary binding of grains by chasmolithic algae. Kobluk and Risk (1977, p. 1077) observed such calcified algal filaments (some were extensions of endoliths, others were true chasmoliths) within the Miami Limestone and in the 83,000 year BP Pleistocene reef terrace of Barbados. Dravis (1979, p. 199) observed similar features in Recent oolitic hardgrounds in the Bahamas. The result is a constructive type of micrite envelope in lieu of the more common destructive type (discussed in Chapter 4). Calcified filaments in the intergranular spaces of constructive micrite envelopes may be infilled and cemented. Consequently, the grains are bound by the resulting mesh of filaments and cement (Kobluk and Risk, 1977, p. 1077). The above type of constructive micrite envelope was not observed in this study. Rather, the calcified filaments observed from Big Pine Key are related to many post-depositional or syndiagenetic effects such as occupation of oomoldic porosity and other secondary pores, association with irregular discoloured areas, and micrite cement coatings on grains.

FUNGAL MICROBORINGS

Straight, singular or branching microborings are abundant in the upper few centimetres of the subaerial surface zone of the Q5 unit (Plates 12-5, 6). Bore lengths are variable from 15-70 μm, and are typically 1-2 μm in diameter. They occur in a randomly oriented fashion and are most easily observed in large, relatively clear grains (e.g. mollusc shells) or coarse spar, although they can be discerned with some difficulty in darker grains (e.g. ooids) and heavily micritized portions of the crust. According to Klappa (1979a, table 1) in his study of calcified filaments in Mediterranean Quaternary calcretes, the borings of fungi and actinomycetes (filamentous or rod-shaped bacteria) fit the above description. Klappa (1979b) assigned significance to fungal hyphae as a component of subaerial laminated crusts interpreted as lichen stromatolites. As with the microborings in the subaerial surface zone of Big Pine Key, Klappa (1979b, p. 393) observed a rapid density decrease away from the surface until only isolated hyphae were recorded at a maximum depth of 12 mm. However, care should be exercised in assigning all such borings to the subaerial surface zone. For example, Perkins and Halsey (1971, p. 845) found fungal borings in grains from the intertidal zone down to water depths of several hundred metres. Furthermore, the borers were quite particle selective, possibly a function of organic content (Perkins and Halsey, 1971, p. 852). Thus, abundant microborings of this nature in a grain may not be indicative of environment of deposition or diagenesis. However, microborings within the dense micritized portions of the crust provide incontestable evidence for a subaerial environment for the microboring process.

According to Kahle (1977, p. 419) structures from endolithic and chasmolithic fungi are the most abundant biogenic structures in crusts of the lower Florida Keys. No certain chasmolithic fungal evidence was observed by the present author in the Big Pine Key subaerial surface zone.

SUBDIVISION OF THE SUBAERIAL EXPOSURE SURFACE ZONE

INTRODUCTION

As mentioned at the beginning of this chapter, 'caliche' refers to secondary calcareous material that has accumulated as a Cca soil horizon or as hard beds beneath the soil (Harrison, 1977, p. 128). In order to integrate the various caliche components and features of Big Pine Key, it is useful to divide the caliche profiles into a laminated crust and a micrite cement zone (Figure 5-1; Plates 11-1 to 6). The former is always visible macroscopically. The latter is not. Closely related to the laminated crust are what Harrison (1977, p. 130) called 'subsurface' micritic stringers'.

LAMINATED CRUST

Petrography: The laminated crusts consist of irregular, pinching and swelling laminae which vary from 0.5 to 1.0 mm in thickness (Plates 11-1, 2, 3, 5, 6). The laminae are defined by colour banding which typically alternates with some combination of dark brown, tan, or white laminae. Areas of predominantly light coloured or dark coloured bands also exist. Where surface relief is discernable, individual laminae may thicken in depressions and thin over highs (Plate 11-1). Thick, indistinct laminae

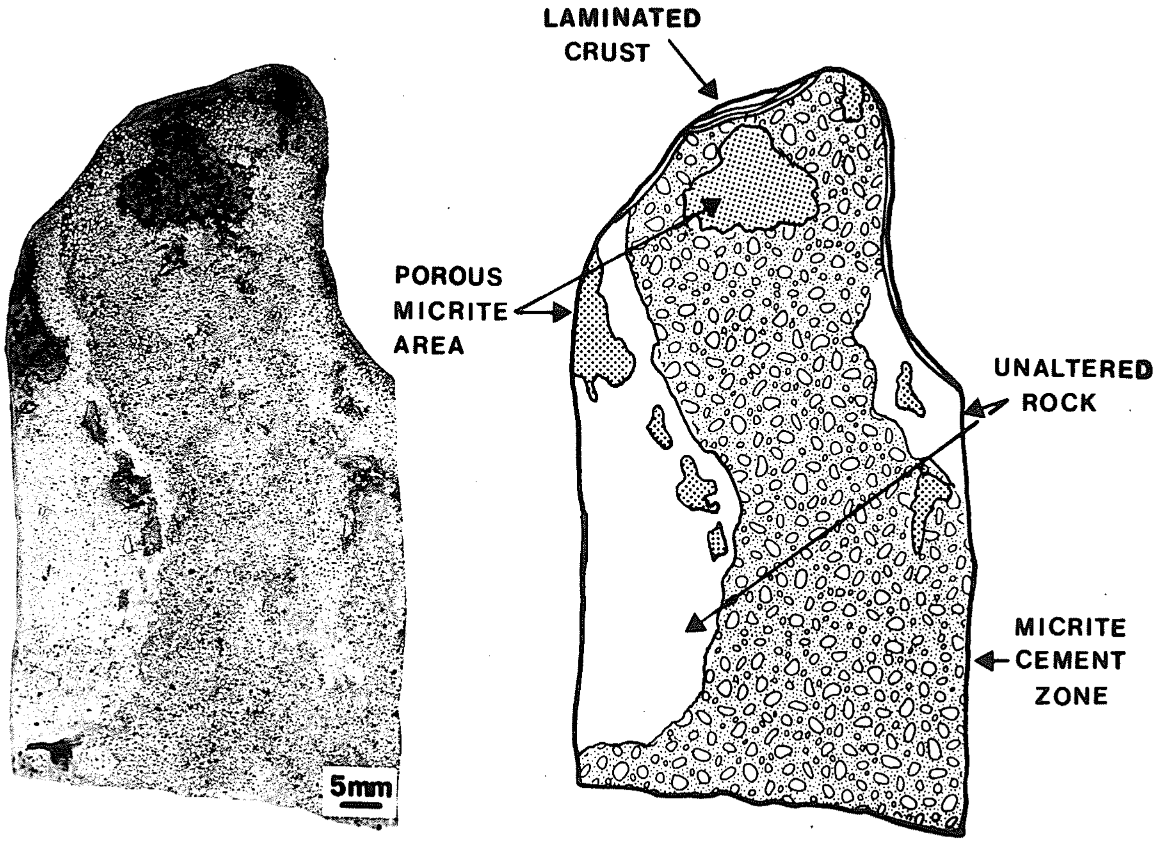
Figure 5-1. Core sample MO-6 subaerial exposure surface zone (caliche).

Laminated Crust: Transitional to case-hardened crust. Laminae on sub-millimetre scale. Colour varies from brown to grey.

Micrite Cement Zone: Characterized by abundant coated grains and calcified filaments. High interparticle and oomoldic porosity. Light brown interparticle matrix. Spar cement postdates micrite.

Porous Micrite Area: Coated calcified filaments form large networks. Primary textures largely obliterated although some grain ghosts occur. Colouration varies from red-brown to tan. Yellow iron (?) compound abundant.

Unaltered Rock: Well sorted ooid grainstone is spar cemented. High interparticle porosity.


5mm
LAMINATED CRUST
POROUS MICRITE AREA
UNALTERED ROCK
MICRITE CEMENT ZONE

up to 1 cm wide occur and may laterally pass into several smaller, more distinct laminae within a few centimetres distance. Total crust thickness varies from less than 1 mm (superficial crust) to greater than 8 cm.

The boundary between alternating laminae and between crust and underlying rock is commonly sharp although gradational boundaries are present. Black peloids generally less than 1 mm in size may be present within the crust. Typically, no relict sedimentary textures are discernable.

Tubular pores ranging from 0.1-0.3 mm occur within the crust, some of which may contain the remains of the roots (?) which once occupied them (Plate 11-3). Rarely, pores up to 3 mm in size occur. These pores are not confined to any particular horizon, although it is evident that some laminae are considerably more porous than others. Moldic porosity after mollusc shell fragments and peloids is common.

In thin section, the distinction between laminae is not as apparent as in hand specimen (Plate 12-3). The laminae are extremely diffuse and consist of irregular and convolute discontinuous horizons approximately 30-150 μm wide. These laminae consist of medium grey-brown to red-brown micrite (1-2 μm, equant crystals) to slightly cloudy microspar (10 μm crystals). Microspar also commonly occurs with micrite as randomly distributed, irregularly shaped aggregrates within the micritic portions of the crust. These aggregates vary in size from 0.15-0.30 mm. Occasionally, these are so abundant as to loosely define a microspar lamina. No relict textures are present.

Microspar and more rarely calcite cement form 10-20 μm isopachous fringes which commonly line vugs. Rarely, clear spar (20-50 μm) replaces bioclasts or fills in irregular secondary voids. Microspar may fill, to various degrees, fractures which occur in the crust. A clotted peloidal texture commonly is present. In some cases, laminae of peloids occur in a red-brown micrite matrix and the laminae may grade into featureless micrite.

Porosity in the laminated crusts typically comprises irregularly shaped vugs of various dimensions (typically 30-100 μm) and irregular, elongate voids (approximately 1 mm by 0.1-0.2 mm) which may be oriented approximately parallel to the laminae. These are possibly related to root paths, or, in terms of the lichen stromatolite hypothesis, may be shrinkage cracks (Klappa, 1979b, p. 398). Smaller, 10-15 μm, circular pores are abundant in certain laminae (Plate 12-3). These pores suggest micrite cement precipitation on calcified filaments to the extent that most of the voids between the filaments are occluded.

Further evidence of calcified filaments occurs as the previously described 'wisps' which commonly subdivide pore spaces. Whereas evidence of calcified filaments is abundant in the subaerial surface zones of the Q5, Q4, and Q3 units, fungal microborings are observed only in the Q5 subaerial surface zone.

Subsurface micritic stringers commonly originate from the lowermost lamina of the crust and invade the underlying rock for several centimetres (Plate 11-4). Where a network of stringers exists, a breccia-like fabric may be imparted to the rock.

The boundary between stringer and host rock may be sharp or diffuse and occurs as a dark micritic zone that slowly passes into a clotted texture which eventually dissipates in the rock. Micritic coatings, 'wisp' networks, and irregular microvuggy porosity may be associated with the stringers.

Subsurface micritic stringers commonly form the boundary between areas of differing diagenetic intensity (Plate 14-4). The inference is that they may have acted as local permeability seals.

Discussion

Introduction: Laminated crusts on Pleistocene and older carbonates have been described in detail in the literature (e.g. Multer and Hoffmeister, 1968; James, 1972; Walkden, 1974; Read, 1974; Harrison, 1977; Klappa, 1979b; Walls et al., 1975; Kahle, 1977; Krumbein, 1968). Such crusts may occur within the soil profile (e.g. the calcrete profile of Read, 1974) or they may occur on exposed bedrock surfaces (e.g. Multer and Hoffmeister, 1968). The crusts vary in degree of induration and there is some correlation among lamina distinctiveness, lamina thickness, and porosity. Crusts up to 16 cm thick have been observed in the lower Florida Keys (Kahle, 1977, p. 416) although the usual thickness is a few centimetres. When found in a soil profile the upper contact with the soil is usually abrupt (Harrison, 1977, p. 130; Read, 1974, p. 258). The lower contact may or may not be abrupt, depending somewhat on the nature of the original sediment and degree of diagenesis which occurred prior to crust formation. Harrison (1977, p. 141) and Read (1974, p. 263) both allude to a gradual porosity reduction of the original sediment, creating an essentially impermeable surface upon which the laminar crust may then develop.

Crusts may form:

1. At some shallow depth beneath the surface but within the sediment or rock (cannibalization of host).
2. At the surface upon the sediment or rock.

Where crusts appear to be forming at the expense of the host, Harrison and Steinen (1978, p. 391) pointed out that the laminated fabric is at best only poorly developed. Regardless of where the crust forms, the mechanism of crust formation and build-up necessitates an aggradational origin.

Origin and Types of Crusts: Multer and Hoffmeister (1968, p. 185) found that there were two types of laminated crusts in the Florida Keys: a dense laminated crust, and a porous laminated crust. The dense crust was thought to have originated under a thin soil or on exposed bedrock receiving periodic drainage from nearby soil, whereas the porous laminated crust was thought to have originated under a thick tropical forest cover where plant roots, etc., would give structural support to the substrate and prevent infilling or compaction. In the former case a more rapid solution-precipitation process could take place while in the latter, the dampening effect of the soil cover on the embryonic crust would prevent the same rapid fluctuations. Walkden (1974, p. 1240-1242) used a similar type of crust differentiation on Carboniferous limestones in England.

The laminated crusts of Big Pine Key do show the characteristics of the laminated crusts described by Hoffmeister and Multer (1968). However, the intimacy of those laminae which can be considered 'dense' as opposed to the 'porous' ones renders the clear separation of the Big Pine Key laminated crusts impractical and at best artificial (Plate 11-3).

The transitional nature of the mostly 'dense' crusts to mostly 'porous' with depth on the order of a few millimetres suggests that the laminated crusts of Big Pine Key be dealt with as a spectrum.

Besides repeated, rapid precipitation from supersaturated solutions, other phenomena have been invoked to help explain the origins of laminated surface crusts. Recently, a direct biological origin for these crusts was proposed by Klappa (1979b) who suggested that the crusts may in fact be 'lichen stromatolites' produced by saxicolous (endolithic and epilithic) lichens.

Origin of Colour Banding: Although it is generally agreed upon that the crusts are in fact an aggrading phenomenon (references previously cited), the origin of the color and textural variation is not straightforward. Multer and Hoffmeister (1968, p. 188) suggested that the colour variations are due to the rise and fall of bacterial populations with wet and dry periods. During wet periods, bacterial activity increases and the products of microbial decomposition are washed together forming the dark organic-rich layers, whereas the drier periods see a decline of bacterial activity and the accumulation of predominantly fine grained non-organic detritus.

The colour banding may be due to a combination of differential staining due to local organic material concentrations or various iron compounds (Klappa, 1979b, p. 398; Multer and Hoffmeister, 1968, p. 188; James, 1972, p. 831; Harrison, 1977, p. 141), but Harrison further suggested that differences in crystal size and/or morphology between successively precipitated layers may also be partly responsible.

MICRITE CEMENT ZONE

Petrography: The micrite cement zone is macroscopically recognized as irregularly shaped grey to red-brown patches which vary from millimetre-size to several centimetres or more (Figure 5-1; Plates 11-5, 6). Where preservation of the original depositional texture is high, the anomalous colouration is limited to the interparticle matrix. In such cases, the grains commonly appear white and stand out sharply from the matrix. The boundary between unaltered rock and the micrite cement zone or between differing intensities of micrite cement zone development may be sharp and outlined by a dense micritic stringer (Plate 14-4), or the boundary may gradational.

The micrite cement zone is typically characterized by the presence of two components, coated grains and coated calcified filaments, and the diagenetic effects related to the presence of the filaments. The degree of grain and filament coating and related diagenesis largely governs whether the micrite cement zone is visible macroscopically.

Micrite cement grain coatings are usually approximately isopachous, and vary in thickness from micrometre size up to 0.15 mm (Plates 12-6; 14-1 to 6). Most commonly they are in the 15-60 μm range. The range is less than this in any given sample. Coatings are rarely laminated, but when they are, the laminae are approximately 10-20 μm, the laminae being defined by differences in colour (Plate 14-1). Where laminae are too diffuse to be identified, colour may gradually change across the coating from light brown on the periphery to medium grey-brown towards the interior.

There is no preferential gravitational thickening of the grain coatings and there is a marked tendency towards an even distribution. Commonly, the coatings encompass more than one grain and more rarely pass into a short, diffuse, unlaminated, grey, micritic stringer. The inner boundary of the coating against the grain is usually sharp whereas the outer boundary may have a diffuse contact with overlying sparry cement, if any is present. In cases where only incipient micrite cementation occurs, cement may show meniscus 'bridges' between grains which are not coated on all sides.

Unlaminated, micrite-coated calcified filaments of root hair or algal origin (previously discussed) are of variable thickness (Plate 13-8). In some cases where there is a high density of filaments, grain boring, and cannibalization, micrite filament coatings may occlude much of the pore space between the filaments (Plate 13-7). The pores which remain define molds of the original filaments. The remaining porous massive micritic area has no relict sedimentological textures. As with the coated grains, the nature of the outer boundary with overlying sparry cement is variable.

Where spar cementation of the rock is uneven, micrite cement and calcified filaments occur in those areas which are poorly spar-cemented. This suggests that spar cementation predates the precipitation of micrite cement and growth of the filaments. However, evidence of micrite cement and calcified filaments actually overlying spar is rare.

In other cases, spar can be seen to occlude pore spaces between coated grains and coated calcified filaments, thus postdating micrite cementation and filament growth. Evidence of micrite cementation and filament growth before and after spar cementation indicates the importance of local factors in determining the diagenetic history of sediments in the subaerial surface zone. Random needle fibres vary greatly in abundance in this zone and commonly are intimately associated with coated calcified filaments ('wisps').

Porosity varies according to the amount of micrite cement which fills interparticle space as coatings on grains and calcified filaments. Porosity also varies according to the amount of leaching of the grains themselves which results in moldic porosity which, in places, may be quite pronounced. In such cases, only the micrite cement rims and intraskeletal pore linings may be left behind (Plate 14-5). The pores defined by the various networks of calcified filaments ('wisps') as they span interparticle space are highly irregular (Plate 14-3).

Within the surface zone of the Q4 unit, centimetre-size vugs commonly show a much higher degree of micritization in an annulus around the vug. This may reflect either preferential access of solutions to this zone, or, based on the presence of some vugs containing abundant calcified filaments in 'soil' patches, may reflect alteration accompanying root hair growth. In this case, the vugs may outline original root pathways.

Discussion: Micrite-coated grains appear to be a common constituent of Pleistocene and older caliche (e.g. Harrison, 1977, p. 133; James, 1972, p. 822; Walkden, 1974, p. 1243). Grains have also been observed to be partly replaced by the micrite which envelopes them (see James, 1972, p. 822).

In a study based on Quarternary caliche from Australia, Read (1974,

p. 263) suggested that the absence of gravitational thickening of the micrite grain coatings was due to a combination of low rainfall and high evaporation. Apparently, this prevented the localization of water films on the underside of grains. Read (1974, p. 263) further suggested that on steep slopes gradual rotation of grains due to creep prevents localized micrite accumulations on the grains.

The network of 'wisps' which span interparticle and secondary pore spaces were interpreted by Harrison (1977, pl. 9-2) as being due to precipitation of micrite cement around roots which have been subsequently leached out. The 'wisps' of Big Pine Key caliche further suggest that they may originate by micrite precipitation on calcified filaments (Plate 13-8).

EVOLUTION OF THE CALICHE PROFILES AT BIG PINE KEY

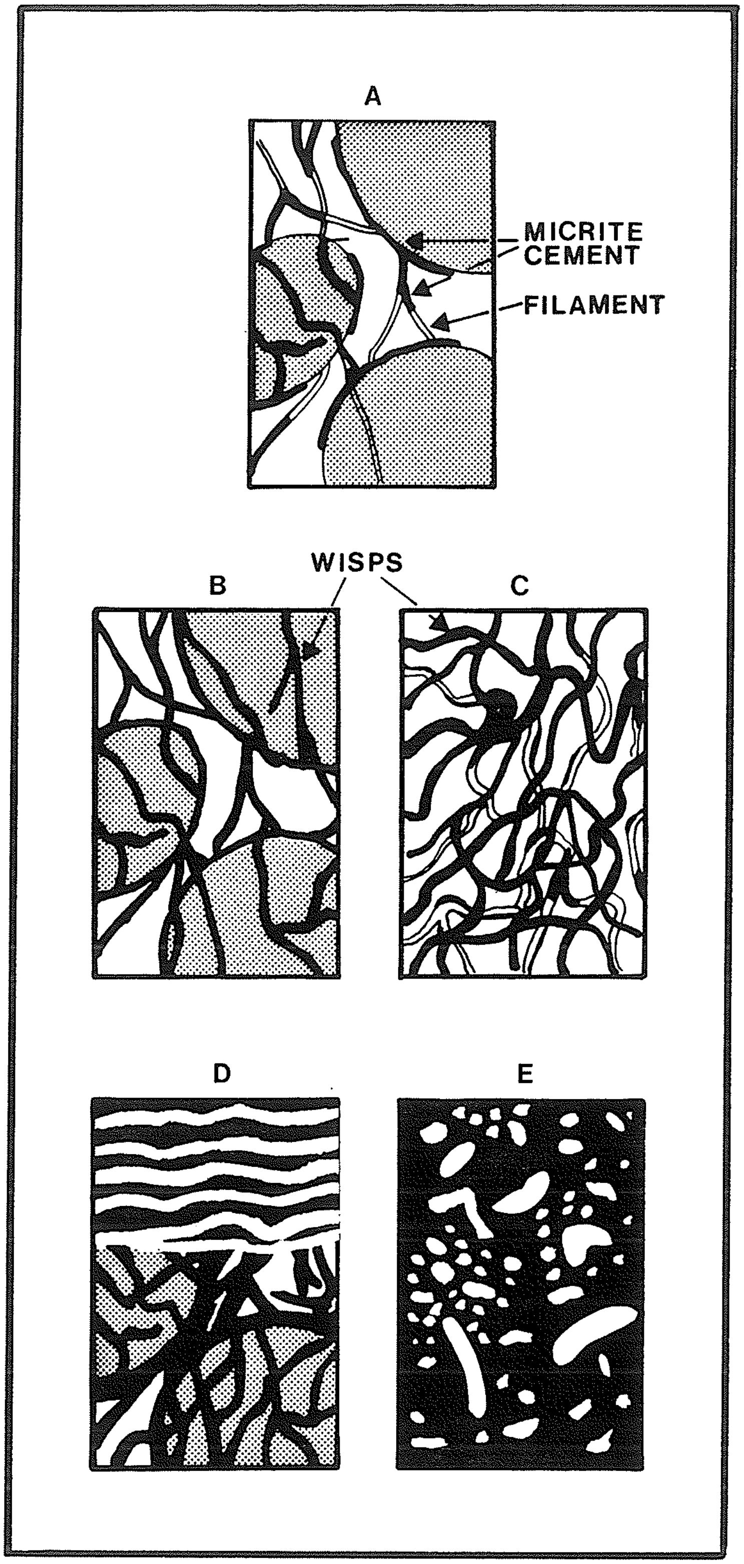
INTRODUCTION

Although both the laminated crust and the micrite cement zone of the caliche profile may be very distinct from one another, they are the end members of a spectrum of alteration that accompanies subaerial exposure of carbonate sediments. The evolution hypothesis of Big Pine Key subaerial surfaces is based mainly on the information furnished by the Holocene (Q5 unit) surface. The rather moderate core recovery, in some instances, of the older caliche profiles plus possible erosion of the surface zone prior to Q5 deposition leads to a rather limited suite of observable features.

SEDIMENT ALTERATION

The evolution, in a simplified manner, is depicted in Figure 5-2.

Figure 5-2. Caliche profile evolution. Diagrammatic sequence A to E illustrates sediment alteration resulting from invasion by organic filaments and micrite precipitation during subaerial exposure. See text for details.


A
MICRITE
CEMENT
FILAMENT
WISPS
B
C
D
E

Unlithified sediments or rock are subaerially exposed. Colonization by algae and/or plants occurs under favourable climatic conditions. The uppermost portion of the exposed unit is penetrated by numerous filaments from these sources, at first mainly in interparticle spaces (A). Previously cemented areas do not allow any large scale invasion by the filaments, except perhaps with some sparmicritization occuring near the surface. Calcification of the filaments may occur while living or after death (Klappa, 1979a, p. 963). Incipient micrite grain coatings begin to form. More filament growth (and calcification) occurs in select areas (B). Grain alteration (boring, cannibalization) occurs in those areas of more profuse filament growth until only grain ghosts or no traces of precursor grains exist (C). A network of calcified filaments remain. Pore space is continuously reduced as micrite cementation continues. At the surface and perhaps at select horizons in the subsurface in areas where pore space has been largely occluded, subsequent micrite precipitation occurs above or around these areas (D). Thus, at the surface, laminated crusts form and in the subsurface micritic stringers develop. Elsewhere, where porosity between filaments has been occluded, remaining pores are molds of the original filaments (E).

Neither laminated crust nor micrite cement zone is always present in the subaerial surface zone and they may occur independently of one another. When they do occur together, however, the micrite cement zone invariably underlies the laminated crust.

TIME RELATIONSHIPS

Harrison (1977, p. 145) pointed out that there is no correlation between caliche profile thickness and age. In addition, sediments of equivalent age do not necessarily show profiles of comparable thickness. This clearly illustrates the importance of the microenvironment in the development of caliche. Whether the environmental differences are due to the presence or absence of a soil, small scale local relief, or even microclimatological differences, to name a few factors, the profile may be extremely variable both in time and space.

Caliche may be significantly younger than the substrate upon which it is developed (Harrison, 1977, p. 145). For example, Multer and Hoffmeister (1968, p. 189) reported bulk radiocarbon ages of the present day crust in the Keys to be in the range of 900 to 4400 years.

More recent work by Robbin and Stipp (1979, p. 179) clearly illustrates the importance of the relationship between caliche and substrate. They showed that laminated crust from Key Largo (Key Largo Limestone) began to form approximately 6000 years ago and accumulated at a rate of 1 cm/2000 years, whereas laminated crust from Big Pine Key (Miami Limestone) began to form approximately 8000 years ago and accumulated at a rate of 1 cm/4000 years. These dates provide the maximum age due to the incorporation of older material in the lowermost laminae of the crusts that were analysed. Robbin and Stipp (1979, p. 180) suggested that the large difference in age between the onset of crust formation and sediment age reflects a long period of emergence prior to initiation of plant growth and crust development, these occurring when sufficient moisture became available.

SOURCE OF CALCIUM CARBONATE

The source of calcium carbonate which is precipitated in the caliche profile may be from several sources. These include the soil cover where meteoric waters percolating downward may acquire Ca^{++} by dissolution of calcite or aragonite, or by cation exchange with local clays (Harrison, 1977; p.144; James, 1972, p.829). Dissolution of the substrate proper may provide a source of $CaCO_3$ (Kahle, 1977, p. 426; Harrison, 1977, p. 144; James, 1972, p. 829). The rate at which the source can provide $CaCO_3$ may be quite high if the substrate retains original unstable mineralogy (e.g. Barbados limestones, Miami and Key Largo Limestones). Other $CaCO_3$ sources of probable lesser importance include atmospheric dust, salt spray, and storms (see James, 1972, p. 829).

PHYSICAL CONTROLS OF NEAR-SURFACE DIAGENESIS

According to Harrison (1977, p. 143) the formation of caliche is dependent on the existence of a favorable moisture balance. The moisture balance is governed by evapotranspiration, precipitation, and soil drainage properties. The requirement is that a moisture deficit exists, the effect being that any available meteoric water is held close to the rock-soil or rock-air interface and any precipitation is confined to there. Too little meteoric water leads to superficial case-hardened surfaces whereas an excess may result in net solution and may induce a karst-like terrain.

Although caliche does not characterize any particular climatic zone (James, 1972, p. 829), previous workers have generally agreed that caliche development is most pronounced in areas that are dominated by a semiarid to arid climate (Reeves, 1970; Krumbein 1968; Read, 1974; Harrison, 1977). The single most important aspect of climate is that it must be characterized by moderately short periods of rainfall followed by extensive periods of evaporation.

A characteristic which is superposed upon the above mentioned feature in many of the areas in which caliche development has been described is the marked seasonality of the climate. Generally a wet (storm-prone) season is followed by a dry season, as is the case throughout south Florida (discussed in Chapter 7). For a more detailed discussion of the physical controls which influence near-surface diagenesis, consult works by Harrison (1977) and James (1972).

CHAPTER 6 - SUBSURFACE DIAGENESIS: PRODUCTS AND PROCESSES

INTRODUCTION

The major components of Holocene carbonate sediments are dominantly aragonite and Mg-calcite with subordinate calcite. Aragonite and Mg-calcite are metastable in sea water, but they become unstable in the realm of meteoric water. Only calcite is stable at near-surface temperature and pressure conditions (Chave et al., 1962, p. 33). As a disequilibrium assemblage, the system approaches equilibrium in the subaerial environment in a series of chemical, isotopic, mineralogical, and fabric changes which constitute freshwater diagenesis.

Along with mineralogical equilibration of the sediment, the precipitation of calcite cement causes lithification and obliteration of primary porosity. Cement is taken "to include all passively precipitated, space-filling carbonate crystals which grow attached to a free surface" (Bathurst, 1976, p. 416). Pray and Choquette (1970, p. 209) indicated that modern carbonate sediments typically have initial porosities which vary from 40-70%. Cementation results in a porosity reduction to typical ancient carbonate rock porosity values of a few percent. The small number of compaction features within ancient carbonate rocks indicates that reduction of porosity via compaction is a minor factor (Bathurst, 1976, p. 415).

Neomorphism, as defined by Folk (1965, p. 21), refers to "all transformations between one mineral and itself or a polymorph". The term includes mineralogical inversion and recrystallization. Thus, phenomena such as carbonate mud (aragonite or Mg-calcite mineralogy) altering to micrite (see Folk, 1965, p. 36), aragonite altering to calcite (calcitization), and aggrading or degrading recrystallization (with a constant mineralogy) are included. The term excludes cementation, as is explicit in the above definition.

The processes of neomorphism and cementation appear superficially quite distinct from one another. However, as Bathurst (1976, p. 416) pointed out, the separation is artificial. In fact, there is a spectrum with large pores on one end and the intercrystalline boundary 'pore' on the other.

GRAIN ALTERATION

BIOCLASTS

Introduction: Bioclastic grains of the Big Pine Key units vary greatly in the degree and nature of alteration, from relatively unaltered to completely recrystallized or even totally dissolved grains. The degree of alteration is commonly organism-dependent to some degree. For example, some species of foraminifera show a marked tendency to remain cryptocrystalline, whereas molluscs and *Halimeda* grains are typically altered. This may reflect original skeletal mineralogy. The degree of microstructure preservation of an Mg-calcite test is high because of incongruent dissolution, whereas the nature of the aragonite to calcite transition is less likely to provide the same textural fidelity (Land, 1967, p. 927). Within a sample, a single species may show the spectrum of variation. It is extremely common to find mollusc shell fragments and *Halimeda* plates which have been totally recrystallized adjacent to those which are absolutely pristine (Plate 17-2).

Molluscs: Mollusc shell fragments or complete valves may have only some of the skeletal microstructure preserved, inferring a neomorphic replacement (wet calcitization; see Bathurst, 1976, p. 347-350; Plates 16-3, 4; 17-1). Other samples consist entirely of a sparry mosaic, with an extremely wide crystal size range from 10-200 μm, that contains no relict microstructure. The inference is that the bioclast is a skelmoldic pore cement. In grainstones, the moldic pore may be preserved by the micrite envelope (usually 1-5 μm thick; see Chapter 4), and in packstones the mud matrix may preserve the mold, with or without the micrite envelope.

Some mollusc fragments show alteration suggestive of partial neomorphism (calcitization) and partial dissolution with either partial or complete subsequent pore filling. The neomorphism is indicated by the preservation (to various degrees) of skeletal microstructure. The dissolution-pore filling process is suggested by a featureless sparry mosaic which may also line incompletely-filled molds with terminated crystals. Commonly, some of the larger mollusc shells show preferential neomorphism or dissolution of selected laminae (Plate 16-3). A coarser prismatic spar may replace the prismatic layer in some shells while the laminated layers remain largely unaltered. This may reflect mineralogical differences as pointed out by Milliman (1974, p. 108), who stated: "... in those shells with mixed mineralogies, the outer layers (generally prismatic or foliated) are calcitic; the inner layers, generally nacreous or laminated, are aragonitic." Syntaxial cement overgrowths on the calcite of the outer layers are relatively common, and terminated crystals line pores in some places.

Halimeda: Halimeda plates are commonly replaced by a 60-100 μm neomorphic spar mosaic which possesses a marked brown pleochroism under plane polarized light (Plate 17-2). The strongest pleochroism corresponds to extinction under crossed nicols. Hudson (1962, p. 498) observed a similar feature in recrystallized lamellibranch shells from Middle Jurassic strata of western Scotland. He used the term 'psuedo-pleochroic' to describe this marked pleochroism and attributed the feature to the presence of conchiolin inclusions less than 1 μm in size, which were frequently aligned according to microstructure of the shells. The cause of the pseudo-pleochroism in recrystallized Halimeda is unknown. By analogy with Middle Jurassic shells, oriented organic inclusions may be responsible.

Halimeda utricles may be infilled by fibrous submarine cement (described in Chapter 4) and/or 10-20 μm blocky spar cement crystals. The blocky calcite cement invariably overlies the submarine cement (Plate 9-7). In the mud-rich sediments, moldic porosity after Halimeda commonly exists. In this case, only the arrangement of the infilled utricles may remain. Halimeda may also show partial neomorphism with plate centres preferentially altered while the plate peripheries remain cryptocrystalline.

Echinoderms: Echinoderm grains are commonly overgrown with a syntaxial rim cement (Plate 17-3). In some cases it is uncertain whether the rim is a cement or neomorphic overgrowth generated by the aggradational recrystallization of surrounding micrite.

Corals: Coral alteration is variable. In most cases, corals of the Q5 unit retain aragonitic mineralogy and microstructure. In one surface sample of Diploria sp., the internal portion consists of a coarse (0.1 - 0.3 mm) spar with minor ghost textures abutting apparently unaltered skeletal material towards the periphery (Plate 17-5). The boundary in some places is not demarcated, but elsewhere, a 10-20 μm wide, indistinct, dark line separates the spar from the unaltered aragonite. The process of coral calcitization was described by James (1974, p. 792), who interpreted the contact between the spar and the apparently unaltered coral aragonite as being the zone of 'chalky aragonite'. The zone is characterized by preferential dissolution parallel to the long axes of the aragonite needles, which separates and etches them, and thus renders the zone more porous.

In the Q4 and Q3 units, corals are typically replaced by a cloudy calcite mosaic with obvious curved intercrystalline boundaries. Crystals range from 30-300 μm in size, although variation within individual coral fragments is less than the above range. Within the larger crystal mosaics, irregular intercrystalline pores ranging from 60-300 μm across may occur.

Foraminifera: Light microscopy does not reveal evidence of recrystallization of peneroplid, miliolid, and encrusting (Homotrema rubrum) foraminifera. However, the original Mg-calcite mineralogy is inferred to have inverted to calcite by incongruent dissolution with attendant microstructure preservation (see Land, 1967, p. 927).

OOIDS

Introduction: Ooids occur abundantly in the Q5 unit, but are found below this only in the Q4 unit in borehole BP-7. Taken together for this discussion, the ooids of Big Pine Key represent a complete spectrum of variation involving the degree of alteration. Some ooids appear to be absolutely pristine except for microborings which may have been produced during ooid growth (discussed in Chapter 4). Others are completely dissolved and the oomoldic nature of the void is inferred from surrounding ooids or from the few remaining outermost laminae. Still other ooids, including the microborings, are completely recrystallized and may be difficult to identify in plane polarized light.

The ooids of the Q5 unit are usually less altered than those of the Q4 unit. Generally, alteration of the Q5 unit ooids has left the nuclei (usually pellets) apparently unaffected (Plates 15-1, 6). Partially or completely recrystallized nuclei or dissolved nuclei surrounded by sparry cortex are more common in the Q4 unit ooids. Commonly, within both units, cortex alteration is more pronounced toward the grain periphery. Examination of the Q5 unit suggests that cortex alteration increases towards the Q4 unit surface. Within a single thin section, several degrees of alteration may be present.

Laminae Alteration: Within both the Q5 and Q4 units, although much more common in the former, ooid cortex laminae may be selectively dissolved, producing a grain in which there are moldic pore 'shells' (Plates 15-1, 3, 4, 6, 7). These 'shells' vary from approximately the thickness of an individual cortex layer (3-7 μm) up to several tens of micrometres, and alternate with aragonitic material. In other cases, 'shells' of spar alternate with cortex aragonite. Occasionally, both types of 'shells' occur in the same grain (Plate 15-1). Friedman (1964, p. 801) described similar features in Pleistocene ooids from Bahamas.

Altered cortex laminae are typically extremely irregular and discontinuous concentrically and, in the case of sparry laminae, a single crystal usually spans the width of a lamina. These crystals are usually equant, although elongate crystals faithfully following the concentricity of the cortex laminations occur. Spar crystals commonly traverse the remaining cortex material into the next altered lamina. Continuation of the spar crystal from the outermost altered lamina to interparticle space is rare, perhaps because of the outermost micrite envelope. In the more highly altered ooids of the Q4 unit, spar continuing from the cortex into the completely or partially replaced nuclei is common. The above features suggest that the spar may be recrystallized cement and is therefore neomorphic.

Alteration of Aragonitic Microboring Molds: Commonly, aragonitic microboring molds remain with or without minor cortex material when the ooid cortex has been leached or infilled by spar (discussed in Chapter 4). In more extreme cases, the concentricity of the cortex laminae may be delineated on the basis of these remaining microborings together with traces of original cortex or other (organic?) material (Plate 15-2). In some of the most extreme cases in the Q4 unit, the ooid form is barely recognizable, and evidence of microborings is no longer present.

Harris et al. (1979, p. 220) outlined the steps involved in preservation of these microborings. The first three steps are outlined in Bathurst (1976), and were previously listed in Chapter 4. These occur within the marine environment. The subsequent steps are related to freshwater diagenesis. These are:

1. Interparticle cementation of ooids and selective dissolution of ooid aragonite leaving the more resistant microboring molds behind. This step is taking place in Holocene ooids of the meteoric water environment of Joulters Cays and the Pleistocene ooids of Big Pine Key.
2. Occlusion of intra-ooid secondary porosity from step 1 by calcite cement. This is also seen in the Pleistocene ooids of Big Pine Key.
3. Dissolution of aragonitic microboring molds and replacement by calcite. This step is inferred to have occurred in the Q5 unit where nearly complete replacement of the ooid cortex shows a lesser abundance of microborings than in less altered ooids (compare Plate 10-1 with Plate 15-2). This may also have occurred in the Q4 unit where the ooid has little internal textural preservation suggesting replacement of the microboring molds with a calcite indistinguishable from calcite of step 2, although it may be argued that the outline of the boring would likely be preserved.

Alteration of Aragonite Rods: SEM examination of ooids shows the nature of cortex alteration. Cortex aragonite rods or needles are variable in size, from 0.2-2 μm long by 0.1-0.3 μm in diameter (Plates 15-5, 8). Some equidimensional crystals were also observed. The rods have blunt, rounded terminations, and their diameters increase and decrease slightly along their length. The cause of this may be the result of differential solution along the length of the rods or related enhancement of primary irregularities. Consequently, the 'unaltered' ooid laminae are not as unaltered as they appear to be via light microscopy.

Winland (1971, p. 80), on the basis of Holocene Bahamian ooids, suggested that such irregular crystals may have recrystallized from much more uniformly sized, tangentially oriented, 1 μm long, aragonite rods. The recrystallization apparently occurs while in the marine environment and is driven by the greater surface area-to-volume ratio of the unaltered aragonite. Furthermore, the preferred orientation of the cortex ultrastructure is destroyed. If the above is true for the Big Pine Key ooids, i.e. the irregular rods are the result of recrystallization rather than dissolution, the amount of cortex ultrastructure obliteration thought to occur by Winland (1971, p. 80) is not substantiated.

The loose mesh of aragonite rods is periodically interrupted by 1-7 μm euhedral rhombs, interpreted as a cement growing in void spaces between rods or small dissolution pores of the cortex (Plate 15-8). The possible dissolution of the aragonite rods may account for the diffuse nature of the pseudo-uniaxial isogyre of the ooids. Where the effects of boring are not readily seen, this is thought due to the increased air space traversed by light rays. This explanation was used by Loreau and

Purser (1973, p. 324) to partially account for the low birefringence of Holocene ooids as compared to standard aragonite.

COMPARISON OF GRAIN ALTERATION IN THE Q UNITS

In general, the bioclasts and ooids in the Q5 unit are less altered than those in the Q4 and Q3 units. There is a marked tendency to have more skelmoldic porosity in the Q4 and Q3 units and there are also more molds filled by calcite cement (reduced skelmoldic porosity) with no relict textural preservation. The reason for the contrast between Q5 and the Q4 and Q3 units is that diagenesis has been active longer in the older units.

NEOMORPHISM OF MICRITE

Close examination of mud-rich areas demonstrates that the micrite is intimately associated with a microcrystalline, relatively clear spar which typically varies from 5-10 μm in size (Plate 17-4).

A clotted texture ('structure grumeleuse') results where micrite appears as 'islands', usually rounded, with the fine spar as a matrix. Where grains are surrounded by micrite, the spar may constitute a 10-20 μm fringe which separates the grain from the micrite (Plate 17-4). The fringe is variably developed and usually non-isopachous. Related to the above, gradational contacts may exist between areas of spar and areas of micrite or between spar and peloids.

Microfabrics in mud-rich areas suggest very strongly that micrite has been altered by aggrading neomorphism to microspar. Folk (1965, p.

40) referred to the resultant fringes of microspar as 'an aureole of neomorphism'. As mentioned in Chapter 3, the irregular distribution of the micrite and spar (e.g. the grainstone to packstone lithologies) is possibly explained as being partly due to aggrading neomorphism of micrite. The amount of micrite that is presently observed may in fact be considerably less than that present at the time of deposition.

CEMENTATION

CALCITE SPAR

Cement fabrics of the Q5, Q4, and Q3 units of Big Pine Key appear to be similar. The crystal size of interparticle spar is variable and there is a strong inverse relationship between size and mud content. Thus, cement crystal size in the well sorted ooid grainstone facies is usually in the 20-60 μm range, with a progressive decrease in crystal size to 5-20 μm as mud content increases.

Spar to particle contacts are generally sharp with the exception of those adjacent to peloids and micrite. Occasionally, spar crystals as large as 150 μm occur as pore linings or fillings. Within mud-rich sediments, such as those of the Q5 unit at the southeast point of Big Pine Key, crystals lining pores are usually less than 10 μm in size, if developed at all.

Intraparticle (intraskeletal) spar is usually significantly coarser (60-120 μm) than adjacent interparticle cement (Plate 9-2). The intraparticle spar occludes, to various degrees, intraparticle pore space. Planar intercrystalline boundaries and terminated crystals are common, both suggestive of cement rather than neomorphic spar (see Bathurst, 1976, p. 417, cement criteria no. 12).

The spar crystals are generally equant in shape. However, in grainstones, bladed spar (20-50 μm long by 10-20 μm wide) with square or acute terminations is oriented perpendicular to mollusc and foraminifera debris and more rarely peloids (Plate 16-1). This suggests a pore-filling cement. Primary and secondary pores may be lined by these elongate cement crystals, but equant scalenohedral and rhombohedral crystals are also common.

Within the Q5 unit and especially in mud-rich lithologies, spar ranging from 5-20 μm is intimately associated with random needle fibres. In some cases, evidence for post-needle fibre cement precipitation occurs where spar fills much of the void space between the random needle fibres.

The typical pore-filling mosaic with its crystal size increasing away from the substrate of the sparry mosaic is rare (Plate 16-2), and when it occurs it is usually only crudely developed (Bathurst, 1976, p. 419, cement criteria no. 13).

MICRITE CEMENT

Micrite cement occurs as a medium brown micrite which lines pores and smooths over microtopography generated by terminated crystals. Meniscus textures are common and pore lining is often irregular (Plate 16-2). This cement is extremely rare and was observed in only two thin sections (5 m below Q5 surface in borehole BP-7 and 1.3 m below the Q4 surface in borehole BP-2).

CEMENTATION TRENDS

In general, there are no obvious trends in cementation as a function of depth. An exception is borehole BP-7 where cementation increases slightly with depth. There is a significant increase of crystal size in small areas (up to a few millimetres across) of Q5 sediments within 0.5 m of the Q4 surface. The occurrence of the large crystals is erratic, even on a thin section scale. The increase in crystal size is most pronounced in bioclast-rich areas where interparticle spaces are large. Cement crystals may reach up to 3-4 mm and some crystals poikilotopically (see Friedman, 1965, p. 651) enclose several grains (Plate 17-3). Within these coarse crystalline areas, planar grain boundaries and terminated crystals lining pores are more prominent than elsewhere. These localized areas are present in some of the boreholes located within the present-day phreatic lenses (see Chapter 7).

The lowermost 10-20 cm of Q5 core from boreholes within the present day meteoric phreatic lenses also tend to be more highly cemented than overlying rock.

COMPARISON OF CEMENTATION IN THE Q UNITS

Within the Q4 unit and to a lesser extent within the Q3 unit, the occurrence of relatively coarse spar exhibiting planar intercrystalline boundaries and a pore filling fabric, both suggestive of cement, is slightly more common than in the Q5 unit. These features are best developed in the primary interparticle pores of those sediments which contain large bioclasts (*Halimeda* and molluscs) and in skelmoldic cavities and other secondary pores.

Generally, the core of the Q4 and Q3 units is more dense and not as friable as that of the Q5 unit, reflecting the overall higher degree of cementation in these older, more diagenetically 'mature' units. However, porosity study shows that in this respect, the units are comparable (discussed in Chapter 8).

CEMENTATION HETEROGENEITIES AND ASSOCIATED POROSITY DEVELOPMENT: IMPLICATIONS FOR ENVIRONMENT OF DIAGENESIS

INTRODUCTION

Vadose and phreatic zone diagenesis is marked by distinct characteristics which enable petrographic distinction. Dunham (1971) modelled the vadose and phreatic zones and compared the results with various Pleistocene grainstones. Land (1970) documented vadose and phreatic diagenesis in the middle Pleistocene Belmont Formation of Bermuda and interpreted such fabrics in light of an inferred sea level which was thought to be 1 to 2 m above present day sea level. Halley and Harris (1979) documented the meteoric (vadose and phreatic) cementation of the Holocene oolite of Joulters Cays, Bahamas. Thorstenson et al. (1972) simulated vadose and phreatic cementation in the laboratory and found that the cement textures produced were very similar to those of Bermuda. Numerous other papers which deal with meteoric water cementation can be found in Bricker (1971, Part III).

OBSERVATIONS

Grainstones: Both cemented areas and relatively uncemented areas within grainstones vary in size from a few grains to more than several centimetres (Plate 16-5). The geometry of the areas is usually irregular, although in some cases there is physical control such as cross-bedding. Cemented areas may or may not have all of the original primary interparticle space occluded. The uncemented or poorly cemented areas are extremely porous (primary interparticle porosity).

Smaller scale (microscopic) cementation heterogeneities also exist. A prominent grain contact cement - the 'meniscus' cement of Dunham (1971) - occurs in some grainstones of the study area (Plates 16-2, 6, 7). The preferential precipitation of calcite cement at grain contacts commonly develops 'rounded' crystals. One or two crystals span the meniscus 'bridge'. Some grainstones show a fairly uniform fringe over most of their peripheries, but thickening usually occurs where the fringes of two grains intersect. Pores with meniscus textures may also contain cements with crystal terminations.

Pore rounding disappears at depth in the sediments in which it is best developed (ooid grainstones of borehole BP-7) and is not observed in the Q4 and Q3 units. This reflects the slightly higher degree of cementation with depth in borehole BP-7 and overall higher degree of cementation of the older units. Consequently, terminated crystals are more common.

Closely linked to the large scale cementation heterogeneities of the grainstones of Big Pine Key is the development of vug and channel porosity. This is the most prominent macroscopic diagenetic feature of these rocks (see Plates 1, 2). These pores are similar in size and geometry to the macroscopic cementation heterogeneities.

Porosity is developed to various degrees in the rocks and no trends are apparent. When highly porous, the core may be extremely crumbly, especially in the Q5 unit with its lower degree of cementation relative to the Q4 and Q3 units. The borehole recovery diagram (Figure 3-3), to a first approximation, can be correlated with intense development of vug and channel porosity (crumbly = core loss). The degree of macroscopically visible interconnection of the pores varies from nil to a complex network.

In most cases, the development of cemented, uncemented, and porous areas appears to be random. However, in the Q5 unit there are several instances where the development is related to a primary sedimentary fabric. The degree of cementation or porosity development is a function of grain size variation associated with cross-bedding and burrowing (Plates 1-1, 2; 2-5). The coarser grained laminae of the cross-bedding and coarser grained burrow infills tend to be less well cemented and/or have large vug or channel pores developed in them, relative to surrounding sediments. Commonly, the vug or channel pore is enclosed by an area of poorly cemented sediment. However, mud-lined burrows occasionally define a channel pore.

Mud-rich Sediments: The development of secondary vug and channel porosity in mud-rich sediments equals or exceeds that in grainstones (Plates 4, 6, 7). Because of the muddy nature of the sediment, grain contact (meniscus) cements are not obvious.

In thin section, the cementation heterogeneities observed in the grainstones have a microspar-micrite heterogeneity counterpart in the mud-rich rocks. Instead of well cemented areas, there are areas which have most interparticle space filled with microspar-micrite and equivalent areas where the larger grains remain with only a small amount of microspar cement (?) fringing at grain contact positions. The original muddy matrix of these sediments has been leached away, producing the appearance of an uncemented grainstone. As with the grainstones, the mud-rich sediments develop the same very poorly cemented, highly porous zones adjacent to vug and channel pores. Primary controls such as burrowing may influence their development to some extent, but little evidence is found to substantiate this.

In the Q5 unit, the nature of the mud-rich sediments within the present-day freshwater lenses and those outside the lenses are not different in terms of cementation or any other diagenetic effects. Unfortunately, the limited number of boreholes and the extremely variable lithology both laterally and vertically do not allow quantitative comparison.

INTERPRETATION AND DISCUSSION

The grain contact or meniscus cement of Dunham (1971) is characteristic of pore space in the vadose zone that contains air and water. This cement texture is caused by capillarity forces which concentrate the water at the grain contacts and result in calcite precipitation occurring near grain contacts. This phenomenon leads to pore rounding and the development of blunted crystals which are the result of 'starvation' against the air-water interface (Dunham, 1971, p. 298).

Although the terms 'vadose' and 'phreatic' are used in a 'macro' sense, these terms can also be applied on a microscale. The cementation heterogeneities of the vadose zone are the result of microvadose and microphreatic systems, which, because of very localized microclimatic and primary sediment inhomogeneity, may be ephemeral. This possibly explains the close coexistence of the blunted pore rounded crystals and terminated crystals with growth not limited by the air-water interface.

The same capillarity forces encourage the retention of water more efficiently by smaller grains. Thus, in a sediment with areas of differing grain size, finer grained material is expected to be better cemented (Dunham, 1971, p. 298; Land, 1970, p. 181). This size influence explains the differentially cemented burrows and burrow infills, and the difference in cementation of the various laminae of the cross-bedded ooid grainstone.

The differential cementation in turn leads to preferential pathways for fluids percolating through the sediments. Hence, any dissolution which occurs will do so preferentially in the more permeable, coarser grained layers. The development of channel pores and vugs in the cross-bedded grainstone and burrow infillings testifies to this. Dunham (1971, p. 298) suggested that the much enhanced transmissability of rocks in which finer grained areas are better cemented than the coarser grained areas, with resultant dissolution of the latter, is due to vadose zone diagnesis.

Whereas grain size may be used as an explanation for dissolution or relative lack of cementation where burrows and cross-bedding exist, the majority of these large vug and channel pores and areas of little cementation show no evidence of being related to either burrowing or cross-bedding. The development of vug and channel porosity appears to be random.

In contrast to the vadose zone, the meteoric phreatic zone contains only one fluid - water. Consequently, cementation tends to be even, without the grain contact cement or patchiness that is associated with vadose cementation. Terminated crystals may be randomly distributed about the grain surfaces, although they are not necessarily isopachous (Halley and Harris, 1979, p. 976). As a result of the more random nature of meteoric phreatic cementation, the Joulters Cays oolite in the meteoric phreatic zone is more friable than that in the vadose zone where grain binding meniscus cements occur (Halley and Harris, 1979, p. 975).

The terminated crystals which tend to become more abundant and larger with depth in the ooid grainstone may reflect several phenomena. First, they may record the rising water table during the last several thousand years (discussed in Chapter 7). Consequently, possible diagenesis in the phreatic zone has been active longer in the lowermost sediments of the Q5 unit. Secondly, because of the highly permeable nature of the grainstone, rainfall which occurred throughout most of the diagenetic history, when the water table was much lower than at present, may have quickly filtered through the rock and accumulated in the lowermost sediments.

The development of vug and channel porosity and the various cementation heterogeneities suggests that the vadose environment has been the main environment of diagenesis in the Q5 unit sediments. Effects attributable to freshwater phreatic diagenesis (e.g. large cement crystals) are minor and occur locally in sediments directly overlying the relatively impermeable Q4 unit surface. This may reflect a perched water table.

The similarity of the diagenetic fabrics (vug and channel pores, differential cementation) in the Q4 and Q3 sediments with those of the Q5 unit, suggests, by analogy, that their principle diagenetic environment was the vadose zone. One coral sample in the Q4 unit shows the 'cross cutting mosaic' texture of Pingitore (1976, p. 992) where single crystals of the calcite mosaic include both the coral skeleton and the intra-skeletal pore spaces (Plate 17-6). The occurrence of this texture in the Q4 unit suggests that, at least, locally, phreatic zone diagenesis may have also been significant. This feature was inferred by Pingitore (1976, p. 992) as being indicative of coral diagenesis under phreatic conditions. Theoretically, the phreatic zone permits development of a void stage between dissolution and reprecipitation and ultimately results in a lesser degree of textural preservation than had the process taken place in the vadose zone.

CHAPTER 7 - PRESENT-DAY FRESHWATER PHREATIC LENSES OF BIG PINE KEY AND POST-DEPOSITIONAL SEA LEVEL FLUCTUATIONS

PRESENT-DAY FRESHWATER PHREATIC LENSES OF BIG PINE KEY

The climate of the Florida Keys is subtropical. It is characterized by a marked wet season from May to October, and a marked dry season from November to April. Approximately 75% of the Keys' precipitation falls between June and October (Multer and Hoffmeister, 1968, p. 184). The seasonality of precipitation on Big Pine Key is manifested in the size and geometry of the present-day freshwater phreatic lenses.

Using water chemistry data from Hanson (1980, table 6), the geometries of the freshwater phreatic lenses of Big Pine Key were reconstructed for representative months of the wet and dry seasons (Figures 7-1, 2). The methodology of construction is outlined in Appendix G. Based on available data, two lenses with a minimum wet season separation of only 0.5 km (data for September 14, 1976) are recognized. The lenses are separated by a low-lying area of less than 1 m elevation (Hanson, 1980, p. 19).

The volume and geometry of the lenses are influenced by the interplay of a number of factors. Volume is dependent of the recharge-discharge balance. The island's only source of recharge is rainfall, while evapotranspiration, lateral and vertical losses to sea water, and artificial effects such as well pumpage are the main routes for discharge. Maximum and minimum lens water volume may be expected at the peak of the wet and dry seasons, respectively. During the wet season the northern

Figure 7-1. Contours on base of freshwater phreatic lenses of Big Pine Key for representative months of the wet and dry seasons. Data from Hanson (1980).

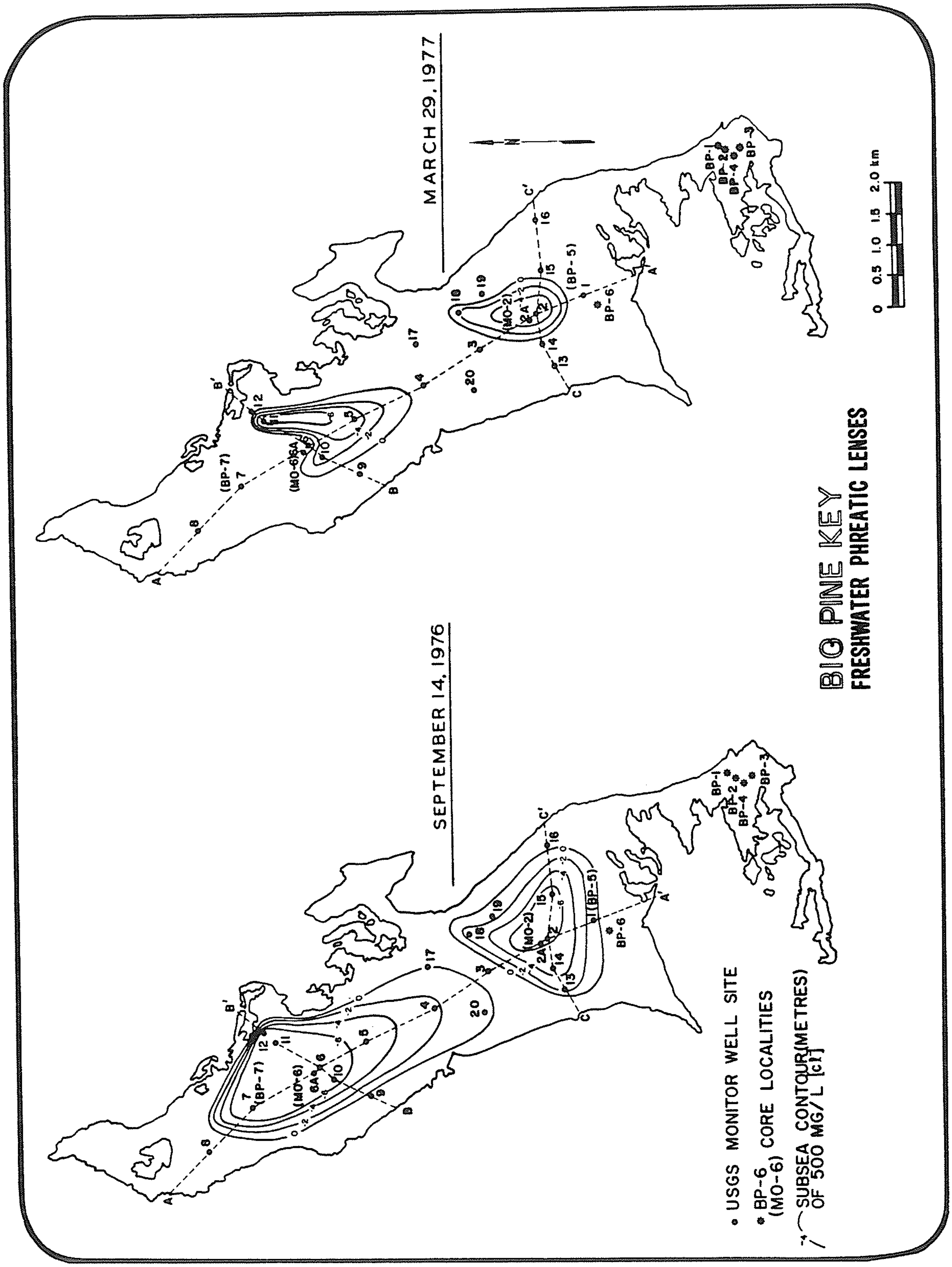
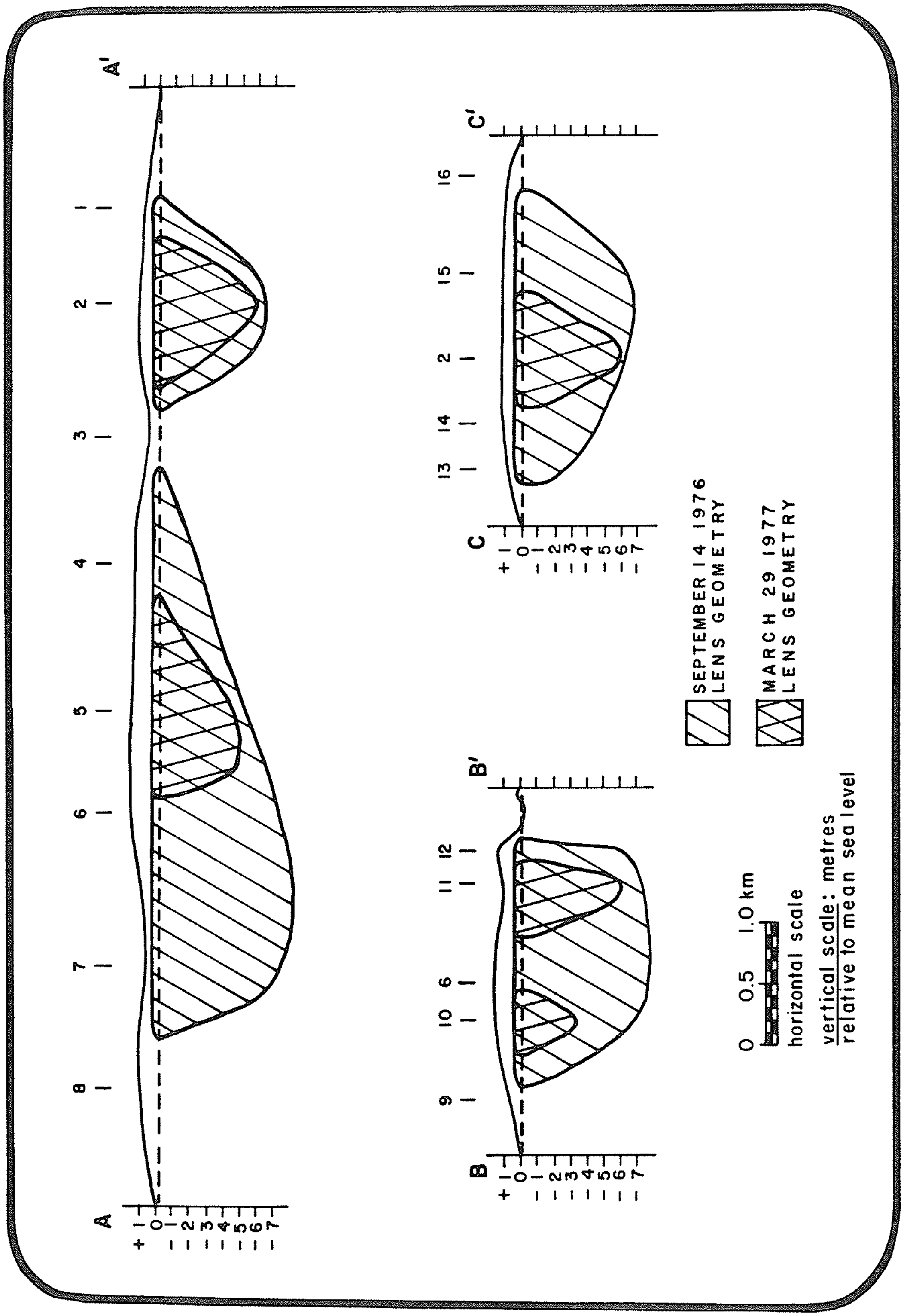

SEPTEMBER 14, 1976
MARCH 29, 1977
BIG PINE KEY
FRESHWATER PHREATIC LENSES
USGS MONITOR WELL SITE
BP-6 (MO-6) CORE LOCALITIES
SUBSEA CONTOUR (METRES) OF 500 MG/L [Cl]
0 0.5 1.0 1.5 2.0 km

Figure 7-2. Cross-sections of lenses shown in Figure 7-1.



A
A'
B
B'
C
C'
SEPTEMBER 14 1976
LENS GEOMETRY
MARCH 29 1977
LENS GEOMETRY
0 0.5 1.0 km
horizontal scale
vertical scale: metres
relative to mean sea level

lens is significantly larger than the southern lens, but the difference during the dry season is not great.

Besides the obvious seasonal fluctuations, daily fluctuations act on a somewhat reduced scale to alter lens geometry. According to Hanson (1980, p. 12) the water table fluctuations of Big Pine Key are primarily influenced by rainfall during the wet season and tides during the dry season. Fluctuations due to evapotranspiration are minimal relative to these.

PLEISTOCENE SEA LEVEL FLUCTUATIONS

In a coastal situation where fresh water meets salt water, an interface develops between the two fluids as a result of their contrasting densities. Under ideal simplified (artificial) conditions, salt water will occur at a depth of 40 times the height of the water table (Ghyben-Herzberg relation; Todd, 1959, p. 278). Accordingly, the water table lies above sea level and slopes down toward the ocean. It follows that a considerable thickness of freshwater would be necessary to raise the water table as little as 1 m in a coastal situation (requires 41 m lens thickness). Therefore, as a simplification for the following discussion, the water table height will be regarded as being the same as sea level at any given time, an assumption also made by Steinen (1973, p. 87) in working with the paleohydrology of Barbados.

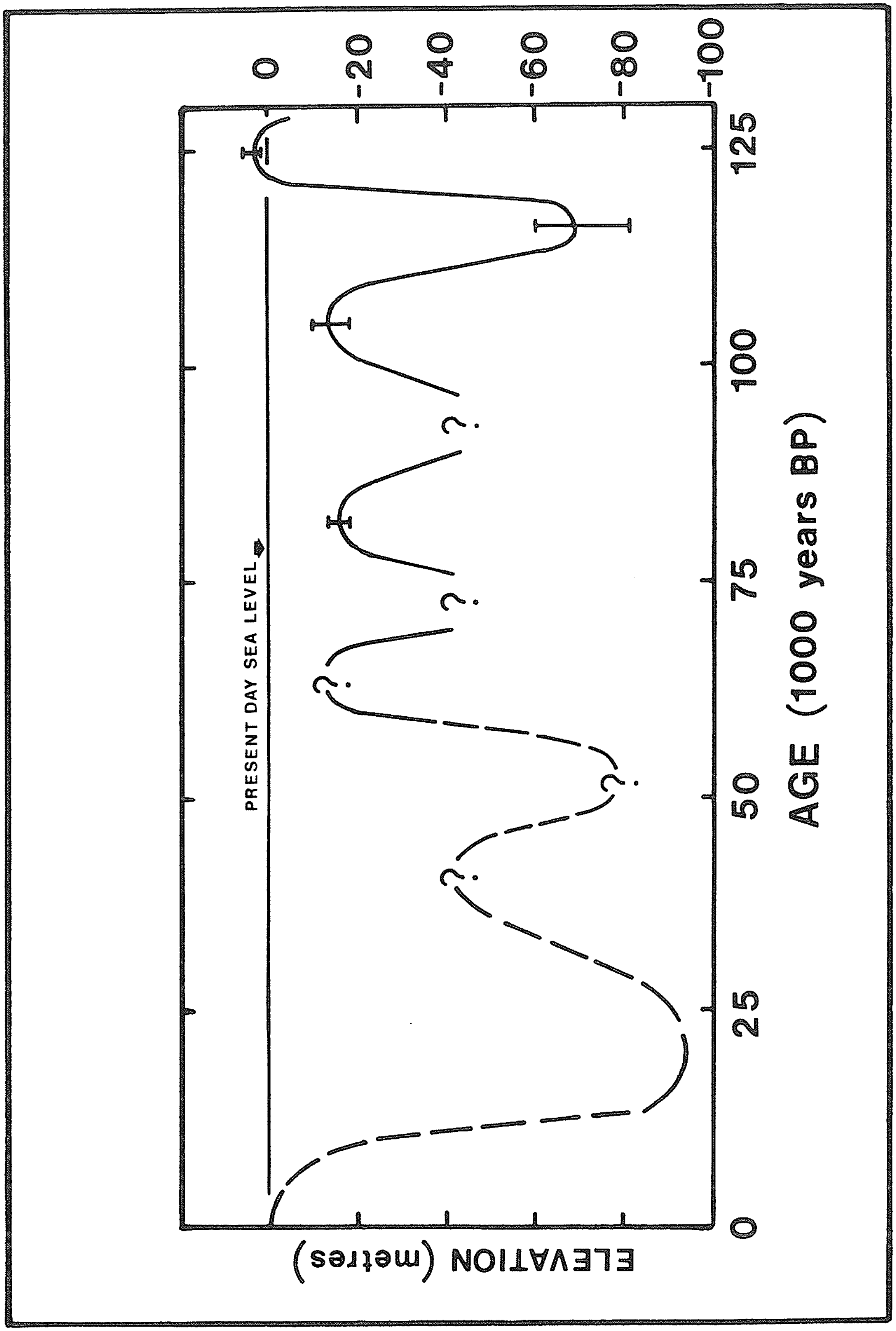
Relative sea level changes are the result of interaction of a eustatic component and a tectonic component (Redfield, 1967, p. 688). The Florida platform has been stable during Pleistocene (Perkins, 1977, p. 137) and Holocene time (Scholl et al., 1968, p. 564). Therefore, the sea level changes need no tectonic correction and any changes are eustatic.

Based largely on data from coral terraces and time-equivalent buried paleoexposure surfaces from Barbados, Steinen et al. (1973) constructed a curve of sea level fluctuations from 125,000 years BP (time of Q5 deposition) to present (Figure 7-3). Figure 7-3 shows four short-duration (20,000 years) climatic cycles post-dating Q5 deposition. The transgression leading into Holocene time is interpreted as a deglacial hemicycle (Steinen et al., 1973, p. 69).

From Figure 3-6, the elevation of the Q4 unit surface can be approximated at -5 m relative to present-day sea level. The sea level curves in Figure 7-3 show that the highest post-depositional sea level stand at 105,000 years BP had a maximum elevation of -10 m relative to present-day sea level. If sea level was equal to the water table height, the Q5 unit sediments remained within the vadose zone. Assuming the conditions necessary for the Ghyben-Herzberg relation, this high stand of sea level was still insufficient to develop a freshwater lens in the lowermost Q5 unit sediments of Big Pine Key. To develop a lens standing 5 m above maximum sea level 105,000 years BP, it would require a freshwater lens extending 200 m below sea level at that time. There is no evidence that climatic or drainage conditions were conducive to a lens of such enormous proportions.

Watts' (1975, p. 344) pollen studies from south-central Florida showed that a dry climate existed in the region between 37,000 and 13,000 years BP. The dry, unforested environment of the Pleistocene and early Holocene (boundary taken as 13,000 years BP) gradually changed to a

Figure 7-3. Late Pleistocene eustatic sea level history of the last 130,000 years. Bars indicate range of values calculated from Barbados data (modified from Steinen et al., 1973 and Steinen, 1973).


PRESENT DAY SEA LEVEL
ELEVATION (metres)
0
-20
-40
-60
-80
-100
AGE (1000 years BP)
0
25
50
75
100
125
?
?
?
?
?

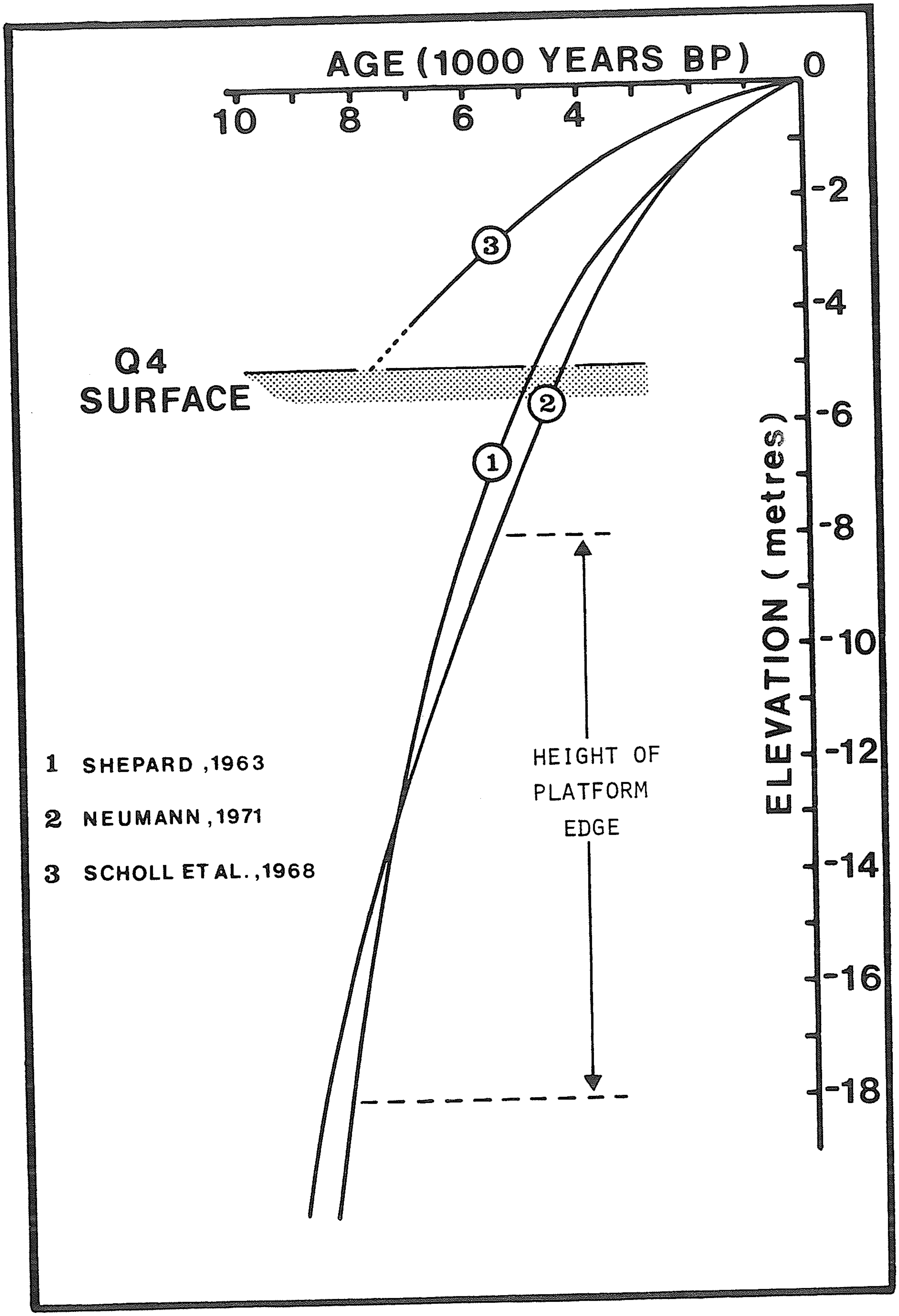
humid, warm climate at approximately 5000 years BP and the modern flora and climate were established at this time. Robbin and Stipp (1979, p. 180) implied that this change to a more humid climate initiated the development of the subaerial laminated crusts of the Keys (discussed in Chapter 5). Watts' (1975) pollen study extended the dry climate back to only 37,000 years (sampling limit), but the Holocene development of the crusts suggests that an extremely dry climate existed since 125,000 years BP. The above inference is cautiously made because of the possibility of the crusts having formed and disappeared as a result of climatic change. Furthermore, it is uncertain whether the degree of diagenesis attained by the Q5 sediments could result from such an arid climate.

INCURSION OF THE HOLOCENE FRESHWATER PHREATIC LENSES

The present-day freshwater phreatic lenses are interpreted as the first lenses to occupy Q5 unit sediments. The time of the initial incursion of the lenses into the Q5 unit can be estimated using sea level curves from a number of sources. Shepard (1963, in Perkins, 1977, fig. 20-b) constructed a generalized curve from stable areas of the world (curve 1, Figure 7-4). Neumann (1971) using peat from boreholes in Bermuda, established a sea level curve (curve 2, Figure 7-4) which shows a rapid rise from 9200 to 4000 years BP at a rate of approximately 36.6 cm/100 years. From 4000 years BP to present the rate slowed to approximately 9.8 cm/100 years. The curve is very similar to that of Shepard (1963).

Redfield (1967) using peat radiocarbon ages from Bermuda, south Florida, North Carolina, and Louisiana found that between 12,000 and

Figure 7-4. Holocene sea level rise and flooding of the Florida carbonate platform. Elevation of Big Pine Key Q4 unit surface is estimated at -5 m. Height of the Florida carbonate platform edge is estimated at -8 to -18 m (Enos, 1977). See text for details.


AGE (1000 YEARS BP)
10
8
6
4
0
-2
-4
-6
-8
-10
-12
-14
-16
-18
ELEVATION (metres)
Q4
SURFACE
HEIGHT OF
PLATFORM
EDGE
1
2
3
1 SHEPARD ,1963
2 NEUMANN ,1971
3 SCHOLL ET AL.,1968

4000 years BP sea level rose at a rate of approximately 33.5 cm/100 years. From 4000 years BP to present, the rate was 7.6 cm/100 years, rates which are similar to those for Bermuda (Neumann, 1971).

In contrast to the above, Scholl et al. (1968), using radiocarbon dated sediment samples from various Florida Keys, Rodriquez Bank, coastal mangrove swamps, and the Everglades of southwestern Florida, developed a curve for eustatic sea level during the last 6500-7000 years which is significantly different from those previously discussed (curve 3, Figure 7-4). From approximately 7000 to 3500 years BP, the sea level rise was approximately 8.3 cm/100 years. After 3500 years BP, the rate slowed to 3.5 cm/100 years. Scholl et al. (1968, p. 564) were confident in their curve after 5000 years BP, but from 5000 to 7000 years BP there is an estimated error of ±1 m of former eustatic levels.

Using -8 to -18 m as the approximate elevation of the Florida platform edge as determined by the depth of the shallow shelf break from Enos (1977, p. 9) and using curves 1 and 2 in Figure 7-4, sea level would have reached the platform edge as long ago as 9000 or as recently as 5000 years BP. Since then, the Florida platform has been below sea level.

Because of the similar climatic regime since 5000 years BP (Watts, 1975, p. 344), it can be assumed that the sizes of the freshwater phreatic lenses were similar to those which exist today. If sea level approximates the water table, then the freshwater phreatic lenses migrated into the lowermost Q5 unit sediments at Big Pine Key approximately 4000-5000 years ago. Extrapolating Scholl et al.'s (1968) curve, the time is estimated at slightly earlier - 7000-8000 years BP. Since then, rising sea level has moved the water table upward in the Q5 unit sediments to its present-day elevation.

The initial contact of the sediments with the freshwater lenses becomes younger upward in the unit. From the above, the Q5 unit sediments of Big Pine Key have been in the vadose zone for 115,000-120,000 years. As the sediments have been in the freshwater phreatic zone for only 9000 years or less, this explains the large vadose zone diagenetic imprint and the relative paucity of effects attributable to freshwater phreatic diagenesis.

Unfortunately, the inferred perched lens effects (discussed in Chapter 6) visible in some of the lowermost Q5 sediments are not amenable to a time interpretation as these sediments may have ponded water over the relatively impermeable Q4 unit surface since sea level first dropped below the Q4 unit surface.

CHAPTER 8 - EVOLUTION OF MINERALOGY AND POROSITY

MINERALOGIC EVOLUTION

INTRODUCTION

This section places the observations made from the Big Pine Key carbonates into the framework generated by such workers as Land et al. (1967) and Gavish and Friedman (1969). The detailed chemistry of diagenesis and related isotopic changes are not discussed, but some information is drawn from studies based on the above. Bathurst (1976) discussed the various facets of diagenesis and reviewed major contributions to this field and the interested reader is directed to this work.

The study of Pleistocene carbonates from Big Pine Key has demonstrated that grain and matrix dissolution and neomorphism occur. Cementation of primary and secondary void space is common and overall, the Q4 and Q3 sediments are more diagenetically advanced than the Q5 sediments. Detailed observations reveal that, besides heterogeneities in the primary sediments of a unit, single grain types may show a spectrum of alteration on a microscopic scale.

PROGRESSIVE DIAGENESIS

Land et al. (1967) and Gavish and Friedman (1969), working with Pleistocene calcarenites from Bermuda and Israel, respectively, described the process of freshwater diagenesis as occurring in several grades or stages. For simplicity the breakdown of the process as per Land et al. (1967) will be used as the main framework.

Five diagenetic grades were recognized in the Bermudan calcarenites.

Each sequential grade represents a higher degree of diagenesis. The grades are:

I. Unconsolidated calcarenite: The mineralogy is dependent on those inorganic and biologic constituents of which it is composed. The sediment is unaltered.

II. Allochthonous meniscus cement: This grade is characterized by a grain contact (meniscus) calcite cement between grains with no detectable mineralogical alteration.

III. Mg-calcite inversion: The primary sediment mineralogy is composed of aragonite and calcite. Any Mg-calcite has completely inverted to calcite without detectable fabric change.

IV. Aragonite alteration: Dissolution of aragonite grains provides the first autochthonous cement. The moldic porosity thus formed approximately balances the continual occlusion of primary porosity by calcite cement. Not all aragonite is leached. _In situ_ calcitization (neomorphism) also occurs.

V. Stabilization: A monomineralic limestone remains comprising solely calcite with approximately 20% porosity.

The absence of Mg-calcite, alteration of aragonite, and cementation of the Q5 unit sediments suggests that, in general, the sediments have attained grade IV of Land et al.'s (1967) diagenetic evolution scheme. The sediments of the Q4 and Q3 units are essentially mineralogically stabilized (Figures 8-1, 2) to calcite, therefore, these sediments have attained grade V. The remaining porosity in the Bermudan calcarenites agrees well with the figures for Big Pine Key carbonates of similar grade or stage (Table 8-1).

Figure 8-1. X-ray diffraction analyses of core from boreholes MO-6 and BP-3. The unconnected dot at the top of the borehole MO-6 analysis represents sediment which is calcite-rich due to micrite coatings around grains. A small (20-30 cm) bioclast-rich interval above the Q4 surface is thought to account for the anomalously high aragonite content at -5.5 m in borehole BP-3. See text for details.

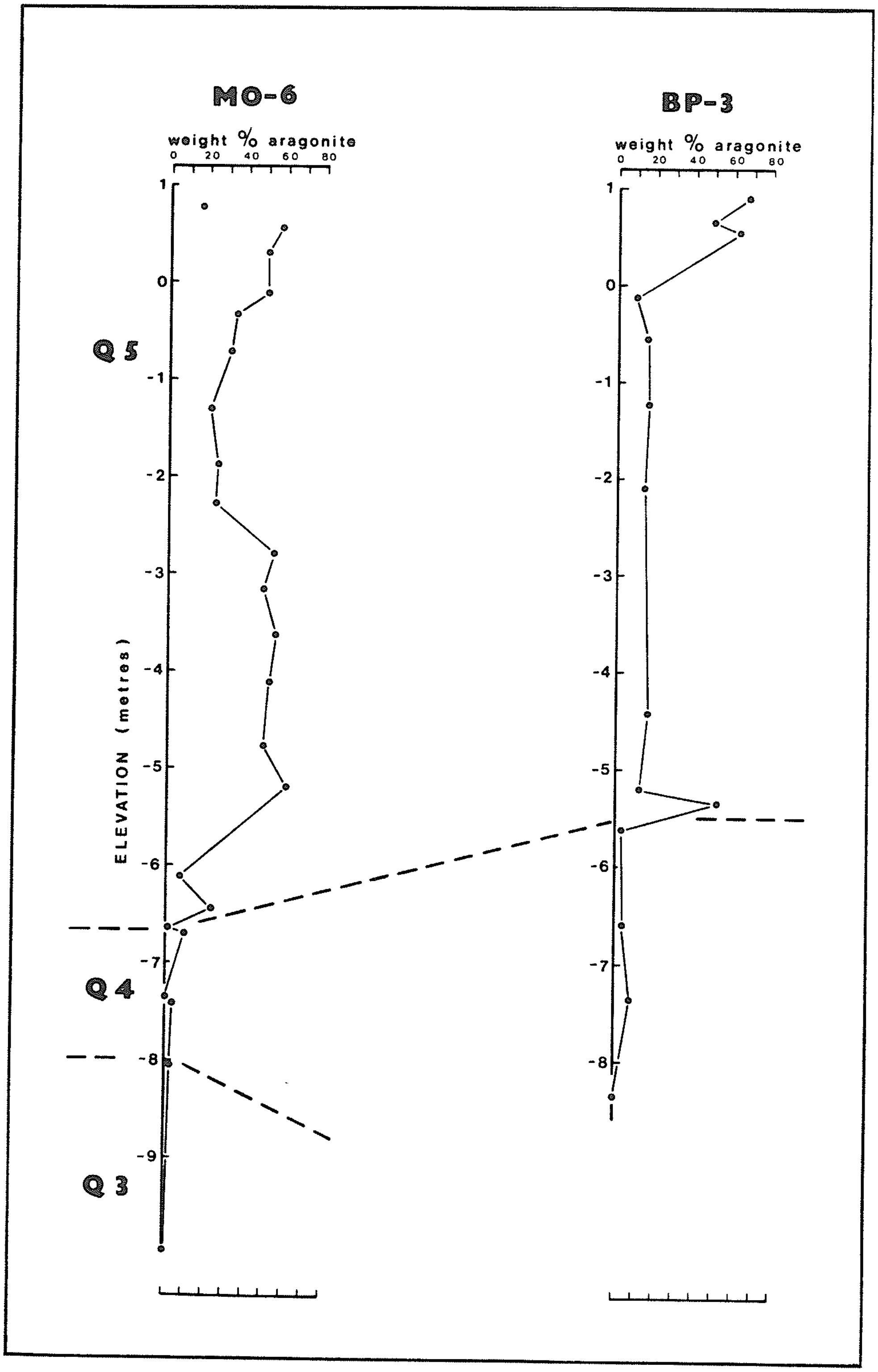

MO-6
BP-3
weight % aragonite
0 20 40 60 80
weight % aragonite
0 20 40 60 80
1
0
-1
-2
-3
-4
-5
-6
-7
-8
-9
ELEVATION (metres)
Q 5
Q 4
Q 3

Figure 8-2. Mineralogy of Big Pine Key sediments showing the likely path of diagenetic evolution for all carbonate sediments. Large dots represent average values for the units indicated. Range of values (stippled bar) determined by one standard deviation of the mean. Stippled area at bottom is an approximate outline of the field for Holocene Bahama Bank skeletal sands (Friedman, 1964). Letters within stippled area represent south Florida Holocene sediments (Enos, 1977): A, Inner shelf (Florida Bay) wackestones and mudstones; B, Inner shelf margin packstones and wackestones; C, Outer shelf margin grainstones and packstones.

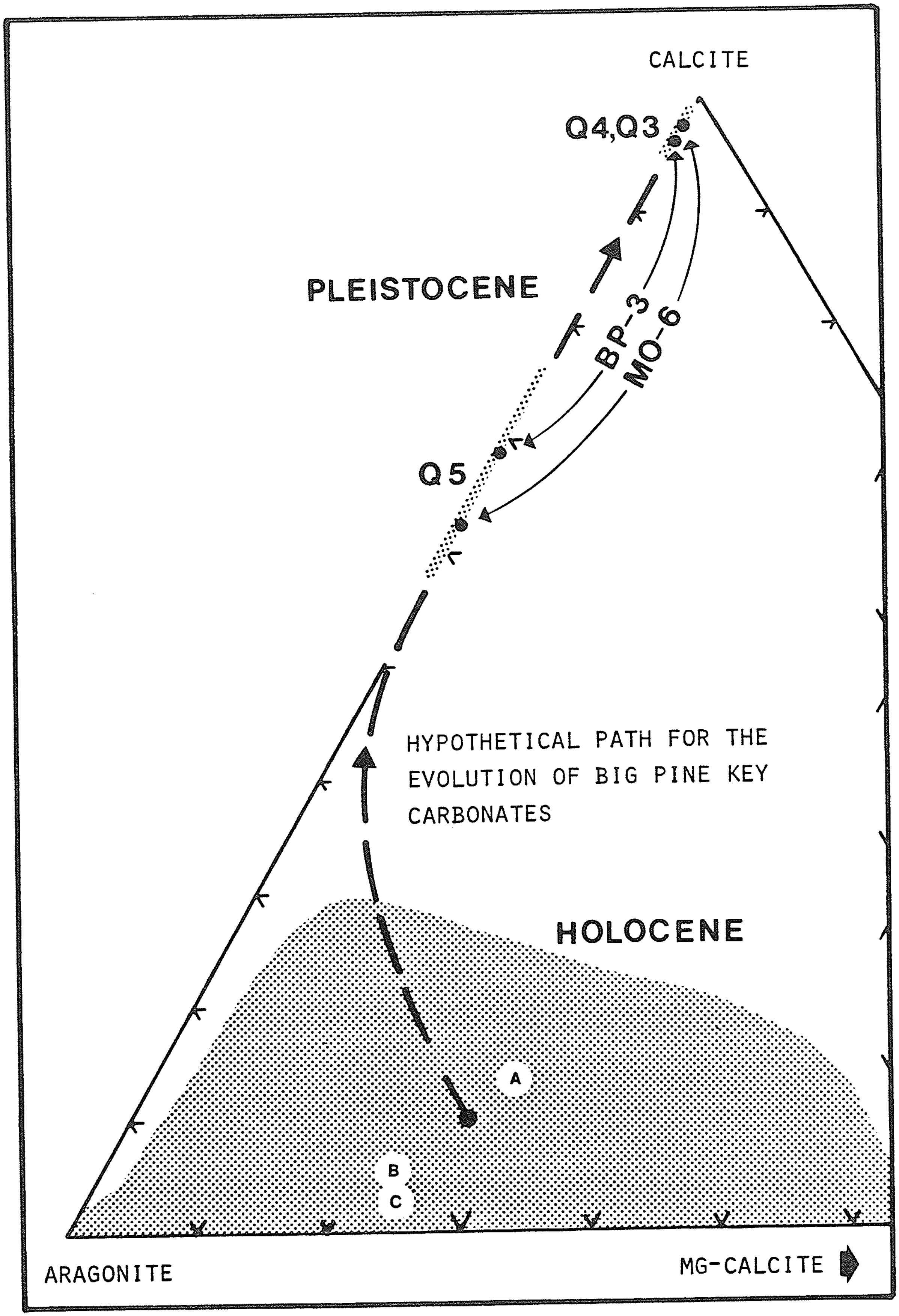

CALCITE
Q4,Q3
PLEISTOCENE
BP-3
MO-6
Q5
HYPOTHETICAL PATH FOR THE
EVOLUTION OF BIG PINE KEY
CARBONATES
HOLOCENE
A
B
C
ARAGONITE
MG-CALCITE

Table 8-1. Porosity data; listed such that total porosity overlies primary and secondary porosity, respectively. Letter or number codes which precede porosity data refer to thin section number. All values are listed in order of increasing depth. No depth correlation exists between boreholes.

Borehole Number

	BP-7	MD-6	MD-2	BP-5	BP-6	BP-1	BP-3
Q5 UNIT	C. 18.7 9.2/9.5	D. 9.7 1.9/7.9	D. 11.4 6.2/5.2	B. 13.4 9.4/4.0	I. 12.3 2.7/9.6	7. 18.2 11.1/7.1	D. 25.0 2.3/22.7
	D. 18.1 8.6/9.5	E. 12.6 4.0/8.5	F. 3.0 1.2/1.8	C. 7.4 3.4/4.0		C. 21.9 5.5/16.4	E. 20.1 0/20.1
	E. 18.1 10.3/7.9	H. 8.9 1.0/7.9	G. 9.6 3.2/6.4	E. 9.1 2.5/6.6		D. 16.1 7.1/9.0	F. 32.6 0/32.6
	F. 12.9 10.4/2.5	K. 11.7 3.7/8.0		F. 4.1 2.2/1.9		E. 11.0 3.0/8.0	
	G. 8.8 8.2/0.6	N. 5.6 0.7/4.9				F. 14.6 2.1/12.5	
						G. 19.6 5.5/14.1	
						H. 19.7 1.0/18.7	
Borehole Average	15.3 9.3/6.0	9.7 2.3/7.4	8.0 3.5/4.5	8.5 4.4/4.1		17.3 5.0/12.3	25.9 0.8/25.1
Q4 UNIT	I. 2.3 0/2.3	P. 8.5 0/8.5	J. 39.8 0/39.8			J. 16.9 0/16.9	G. 19.0 1.3/17.7
	J. 8.6 0/8.6	79.0.6 0/0.6	K. 14.2 1.5/12.7			K. 23.7 0/23.7	H. 7.8 0/7.8
						L. 8.0 0/8.0	J. 31.8 0/31.8
						M. 19.3 3.8/15.5	L. 39.7 0/39.7
Borehole Average	4.3 0/4.3	4.6 0/4.6	27.0 0.8/26.2			17.0 1.0/16.0	24.6 0.3/24.3
Q3 UNIT		V. 11.0 0/11.0	O. 18.0 0/18.0				
		R. 14.6 1.8/12.8	P. 23.7 0/23.7				
		S. 9.2 0/9.2					
Borehole Average		11.6 0.6/11.0	20.9 0/20.9				

Assuming approximately comparable primary sediment mineralogy, it is significant to note that mud-rich sediments (borehole BP-3, Figure 8-1) of the Q5 unit show a higher degree of mineralogical stabilization than do those sediments which are largely grainstones (borehole MO-6, Figure 8-1). This is based on the higher overall calcite content of the mud-rich sediments compared to the grainstones. Further evidence of preferred stabilization of mud-rich sediments is observed within borehole MO-6 (Figure 8-1). The interval of relatively low aragonite content between 0 and -2.5 m is largely packstone. Related to the low aragonite content within this interval, intense development of vug and channel porosity has presumably resulted from dissolution of much of the aragonitic mud.

SOURCE OF CEMENT AND TIME REQUIRED FOR DIAGENESIS

In Bermuda the first cements were allochthonous and were possibly derived from dissolution of overlying carbonate accompanying the evolution of the paleosols. Autochthonous cements were derived from the dissolution of aragonite in the same sediments in which they were precipitated and became important in a later stage of diagenesis. It can be reasonably inferred that the initial cements were related to leaching at the subaerial surface during the early stages of subaerial exposure. Simultaneously or possibly later, autochthonous cements derived from dissolution of bioclasts, ooids, and mud must have been significant.

The attempt to put a time scale on the various grades of diagenesis previously outlined is an oversimplification because of the numerous complexities involved. These include availability of water, chemistry, influence of biologic material, etc.

> "One must not assume that freshwater alteration of any given formation proceeded uniformly with time. Formations are not diagenetic units, even though there is a gross positive correlation between age and alteration. Apparently, vadose freshwater percolating through rocks like those of Bermuda does not move on a large scale as an advancing front but tends to flow in restricted channelways, the control of which may be obvious, such as bedding, or may seem random. The rocks in the channelways become more highly altered than those which surround the channel". (Land et al., 1967, p. 1002)

Work by Gavish and Friedman (1969) on Pleistocene sediments from Israel suggested that the inversion of Mg-calcite to calcite is rapid relative to the alteration of aragonite. Mg-calcite inversion took place in less than 7000-10,000 years whereas aragonite disappeared in less than 80,000-100,000 years.

POROSITY EVOLUTION

INTRODUCTION

A quantitative porosity study was conducted in order to provide answers to the following questions:

1. Is there a contrast in porosity type between the relatively young Q5 unit and the older, more diagenetically 'mature' Q4 and Q3 units?
2. How do apparently less permeable mud-rich sediments alter relative to grainstones during early diagenesis?

Two general categories of porosity were recognized: (1) primary porosity in which interparticle porosity was dominant, and (2) secondary porosity comprising mostly small vugs with minor moldic porosity (commonly after ooids). The porosity dealt with in this section is a 'matrix' type exclusive of the large, less easily quantifiable vug and channel pores.

POROSITY ANALYSIS AND CONCLUSIONS

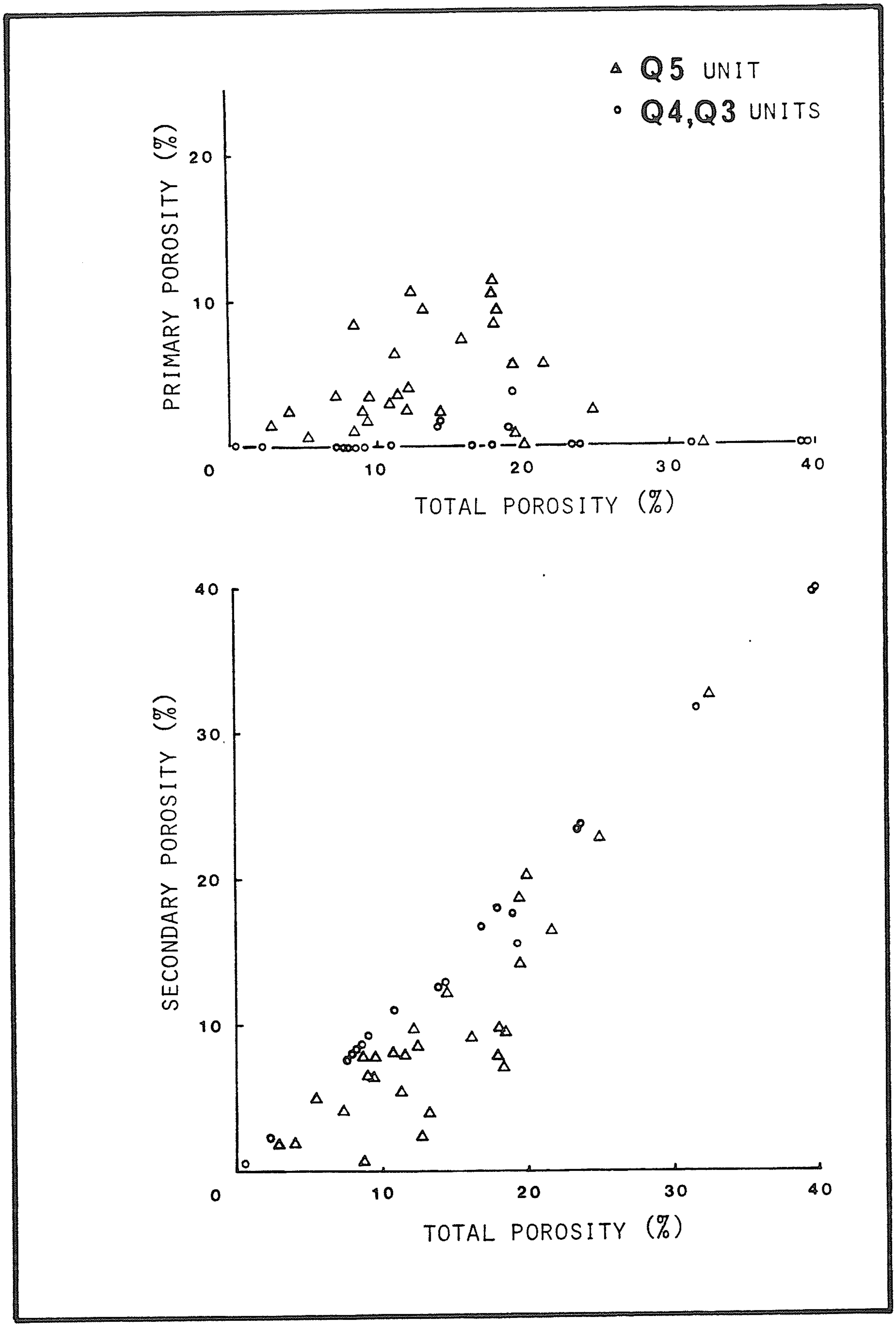
The variable lithologies of the Q5 and older units do not permit a straightforward time-porosity evolution study like those in which sediment type is relatively constant. Previously, such studies were involved mainly with grainstones (e.g. Harrison, 1975). Despite the complexities of the Big Pine Key samples, some interesting points emerge.

Comparison of the Q units (Table 8-1) demonstrates that total porosity values in all units are extremely variable. A simple 't' test comparing the total porosities of the Q5 and the Q4 and Q3 units suggests that no significant difference at the 90% confidence level exists between the two populations from which the samples were drawn. In addition, variations in total porosity are apparently without trend either vertically (within a borehole Q unit) or laterally (within a Q unit).

Examination of primary and secondary porosity in the units (Figure 8-3) demonstrates that primary porosity is almost non-existant in the older Q4 and Q3 units (also see Table 8-1). The absence of primary porosity in the older units corresponds closely to the trend described by several workers (e.g. Harrison, 1975) where the stabilization of sediment mineralogy from aragonite to calcite is accompanied by a distinct change in the porosity fabric (from primary to secondary), thus retaining porosity.

Within the Q5 unit, a striking difference exists if porosities are considered in terms of primary and secondary porosity. The highest

Figure 8-3. Total porosity as a function of primary and secondary porosity in the Q units of Big Pine Key. Total analyses for the Q5, Q4, and Q3 units are 28, 14, and 5, respectively.


Q5 UNIT
Q4,Q3 UNITS
PRIMARY POROSITY (%)
20
10
0
10
20
30
40
TOTAL POROSITY (%)
SECONDARY POROSITY (%)
40
30
20
10
0
10
20
30
40
TOTAL POROSITY (%)

primary porosities occur in the well sorted ooid grainstone (borehole BP-7, Q5 unit). Most of the other samples of the Q5 unit have secondary porosity dominant. The most mud-rich lithology also has the highest secondary and total porosity of the unit (borehole BP-3).

The two sediment types, grainstone and packstone, have presumably spent approximately the same amount of time in the vadose zone. During the 100,000 years plus of subaerial exposure, grainstones were cemented and their primary porosity was partially occluded. During the same time span, mud-rich sediments developed a pronounced secondary 'matrix' porosity, possibly the result of dissolution of largely aragonitic mud (modern shallow water lime muds comprise 60-95% aragonite, 5-40% Mg-calcite, and 0-10% calcite; Steinen, 1978, p. 1140).

From the above, the mud-rich sediments of the Q5 unit of Big Pine Key apparently attained secondary pore fabrics (Table 8-1) earlier than did grainstones under equivalent conditions. The reasons for this are uncertain. Intuitively, particle size may have been an important factor. Aragonitic mud appears to be more easily dissolved than aragonitic grains.

CHAPTER 9 SUMMARY AND CONCLUSIONS

FACIES AND STRATIGRAPHY

Nine boreholes in Big Pine Key, aligned approximately perpendicular to depositional strike, permit the recognition of three facies and three stratigraphic units. The stratigraphic units consist of various marine carbonate sediments and are separated by distinct subaerial exposure zones. The stratigraphic units correspond to Perkins' (1977) Q5, Q4, and Q3 units. These units record sea level high stands at 134,000, 180,000 and 236,000 years BP, respectively (Mitterer, 1975).

The Q5 unit comprises three intergradational facies which are defined on the basis of their principle components. In a southeastern direction, an ooid grainstone grades into an ooid grainstone to packstone. These two facies constitute the Miami Limestone of Hoffmeister et al. (1967). The southeast point of Big Pine Key is composed of a peloid-bioclast packstone to grainstone facies, and is the southernmost expression of exposed Key Largo Limestone. The southeasterly trend of decreasing sorting and ooid content and increasing micrite and bioclast content (especially corals) suggests that there are two possible paleoenvironmental settings. In both cases the ooid grainstone represents deposition in a high energy environment, accounting for the cross-bedding and scarcity of burrows. The ooid grainstone to packstone and peloid-bioclast packstone to grainstone facies suggest deposition in a low energy environment either below wave base, or in an environment in which a reef framework existed seaward of the site of deposition and provided shelter.

The Q4 and Q3 units are composed of a peloid-bioclast packstone to grainstone facies which is coral-rich and highly variable. The presence of ooids in the northwestern part of the Q4 unit of Big Pine Key suggests that previously undocumented (incipient?) ooid shoals were forming at that time.

SUBMARINE DIAGENESIS

Submarine cement is present in the sediments of Big Pine Key, although it is only common in the sediments of the Q5 unit. Cement occurs both as interparticle and, much more frequently, as intraparticle pore fillings. Fibrous interparticle submarine cement is rare and local in development. In contrast, intraparticle aragonite submarine cementation is common and is best developed in the intraparticle pores of corals within the Q5 unit, although it is also found in foraminifera, molluscs, *Halimeda*, and microborings. The presence of intraparticle cement and the relative absence of interparticle submarine cement suggests that the protected microenvironment of intraparticle pore space in which precipitation occurs can exist within an apparently largely uncemented carbonate sand shoal.

Grain micritization due to boring marine endoliths is common in all units and is most clearly observed in ooids. Their marine origin is deduced from their concentrations in various layers of the ooid cortex and their aragonitic mineralogy. Dissolution of the ooid cortex leaves aragonitic microboring molds relatively unaltered. Four categories of molds are recognized in the ooids, of which only two are common. Spherical to irregular 10-65 μm diameter molds and 7-10 μm diameter filamentous molds up to 40 μm long suggest an algal origin for the great majority of grain microborings.

SUBAERIAL EXPOSURE SURFACE ZONE

The subaerial exposure surface zones or caliche profiles of the Q5, Q4, and Q3 units can be separated into two types of features: a laminated crust and a micrite cement zone. In reality these are part of a spectrum of alteration fabrics which result from the subaerial exposure of carbonate sediments. Important components of caliche include micrite cement, calcified filaments, random needle fibres, peloids and clotted fabrics, and fungal microborings. Of lesser importance are fibrous calcite cement and authigenic iron oxides and/or hydroxides. The calcified filaments are thought to originate from root hairs, although an algal origin is possible.

The evolution of the caliche profile begins with the obliteration of primary sediment porosity by micrite precipitation around grains and by invading filaments which later become calcified. In extreme cases, alteration accompanying a profuse development of root hairs or algal filaments produces a porous micrite-rich area with no relict texture. Continued occlusion of porosity results in a relatively impermeable foundation upon which the laminated crust can then develop.

SUBSURFACE DIAGENESIS

Bioclast and ooid alteration is variable. In general, grains in the Q5 unit are less altered than those of the Q4 and Q3 units. This contrast is due to the longer diagenetic history of these older units. Typically, the degree of cementation is higher in the older units, resulting in a less friable rock.

Micrite may be altered neomorphically to microspar or leached from the sediments. The close interrelationship of grainstone to packstone in the Big Pine Key sediments is thought to be at least partially due to the alteration of this micrite, which leaves areas resembling grainstones.

Ooid alteration occurs on several scales. The irregular shape of cortex aragonite rods suggests differential dissolution. Commonly, entire cortex layers are completely leached and the moldic pore thus produced may be partially or completely calcite infilled. The aragonitic microboring molds typically are relatively unaltered.

Cementation is heterogeneous on both a macroscopic and on a microscopic scale. Areas within grainstones which are cemented may be found immediately adjacent to those which are uncemented. The development of vug and channel porosity is closely related to these areas of poor cementation. Occasionally, the cementation heterogeneities may be due to grain size differences related to cross-bedding or burrowing. Microscopically, a prominent meniscus cement with pore rounding is sometimes present.

In mud-rich sediments, macroscopic vug and channel porosity is also associated with areas where micrite was leached. The development of features such as vug and channel porosity, the various cementation heterogeneities (patchy distribution, meniscus cements), and pore rounding strongly suggests that the vadose zone has been the principle influence in the diagenesis of the Q5 unit sediments. The similarity of diagenetic fabrics in the Q4 and Q3 units suggests that the vadose environment has also been a major influence in the diagenesis of these sediments.

Assuming that the water table height in a coastal situation approximates sea level, and using previously established sea level curves for the last 130,000 years BP, the sediments of the Q5 unit have been in the vadose zone for 115,000 - 120,000 years. The incursion of the present-day freshwater phreatic lenses may have occurred as recently as several thousand years ago. The paucity of effects attributable to freshwater phreatic diagenesis and the diagenetic similarity of boreholes within and outside the present-day freshwater phreatic lenses may be explained by this comparatively short period available for freshwater phreatic diagenesis. Effects suggestive of perched lenses are minor and local in occurrence.

EVOLUTION OF MINERALOGY AND POROSITY

The presence of significant quantities of aragonite, the absence of Mg-calcite, and the degree of cementation suggest that diagenetic grade IV of Land et al. (1967) has been attained in the Q5 unit sediments. The mineralogically stabilized sediments of the Q4 and Q3 units suggest diagenetic grade V. The units of Big Pine Key clearly demonstrate the path of progressive diagenesis via mineralogical stabilization.

In terms of total matrix porosity, the Q units are not significantly different. However, secondary porosity is much more extensive in the older units than in the Q5 unit. This is in accordance with the observation of other workers (e.g. Harrison, 1975) that during early, subaerial diagenesis, the amount of porosity remains constant but the porosity fabric is changed significantly from a primary pore fabric to a secondary pore fabric.

The Q5 unit of Big Pine Key also demonstrates that the nature of the sediment itself also has significant control on porosity evolution. Mud-rich sediments are as porous as grainstones, but porosity is almost entirely secondary. In the same time span that grainstones undergo primary interparticle porosity occlusion during early diagenesis, mud-rich sediments apparently develop secondary pore fabrics, presumably the result of dissolution of aragonite mud. Accompanying this early secondary pore fabric is a slightly higher degree of mineralogical stabilization of the mud-rich sediments.

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PLATES

PLATE 1

OOID GRAINSTONE FACIES - Q5 UNIT

1-1. Cross-bedding in ooid grainstone. Alternating laminae of varying grain size at angles of 20-25° and related vug and channel porosity. Scale as shown. (Core, BP-7).

1-2. Cross-bedded ooid grainstone shows preferential development of vug and channel porosity in coarser grained laminae. Scale bar 1 cm. (Core, BP-7-71).

1-3. Typical medium sand size ooid grainstone. Scale bar 2 mm. (Thin section, negative print, BP-7-F).

1-4. Moderately sorted bioclast-rich horizon in well sorted ooid grainstone. Note scarcity of oolitic coatings on bioclasts. Scale bar 1 mm. (Thin section, microphotograph, BP-7-D).

1-5. Coarse sand size ooid grainstone showing departure from sphericity in larger grains. Scale bar 2 mm. (Thin section, negative print, MO-6-C).

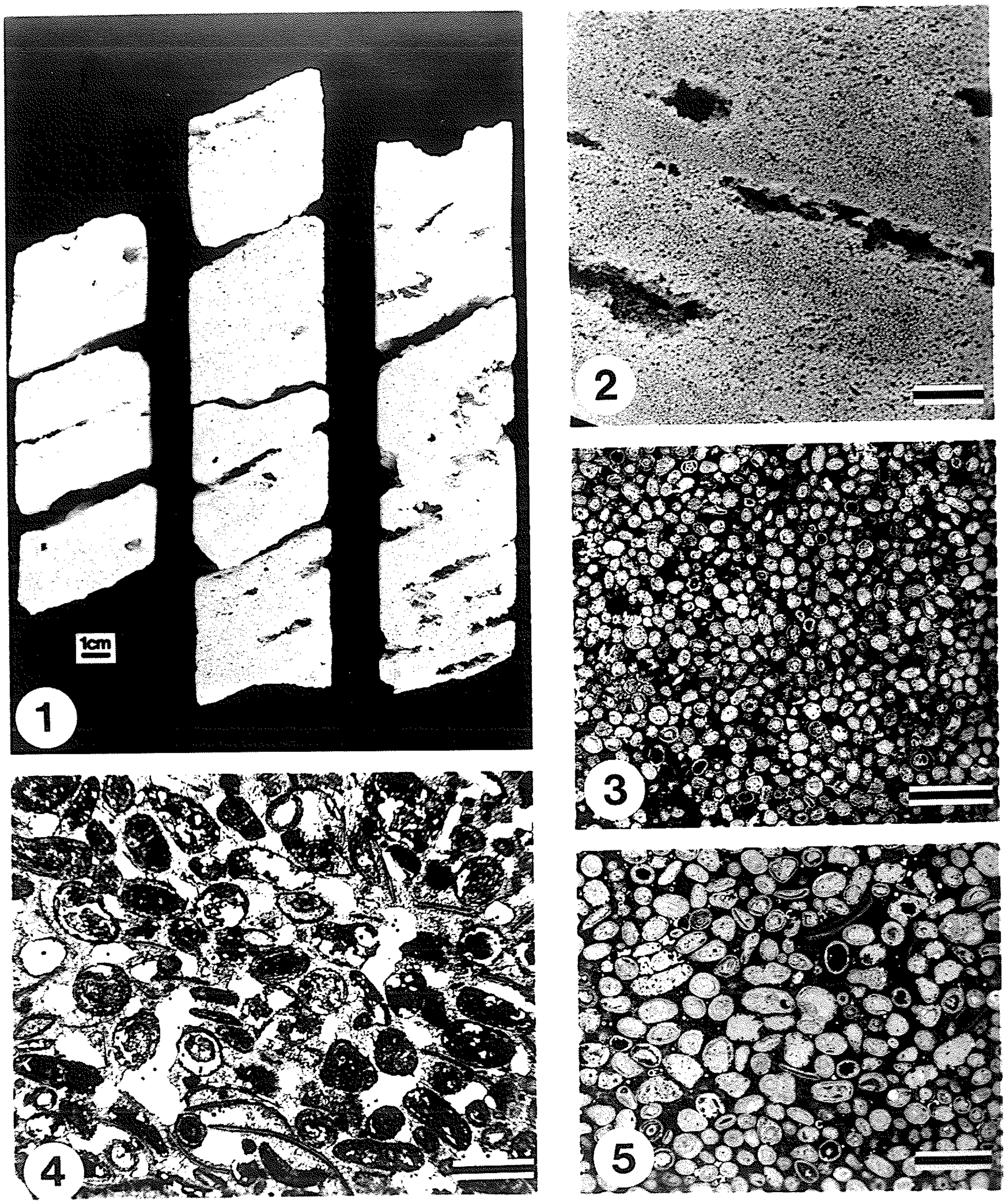
1cm
1
2
3
4
5

PLATE 2

OOID GRAINSTONE TO PACKSTONE FACIES - Q5 UNIT
(Core, scale bar 2 cm)

2-1. Abundant *Halimeda* (arrows) in ooid grainstone. (MO-6-48).

2-2. Typical ooid grainstone to packstone with large echinoderms (E) and burrow (B). (MO-2-53).

2-3. Large pelecypod fragments (arrows) in ooid grainstone to packstone. (MO-6-14).

2-4. Small bioclast-rich (dominantly *Halimeda*) intervals (arrows) from lower facies. Note small encrusting bryozoan, *Schizoporella floridana* (S). (BP-7-76, 75).

2-5. Highly burrowed (B) ooid grainstone to packstone. (BP-5-39).

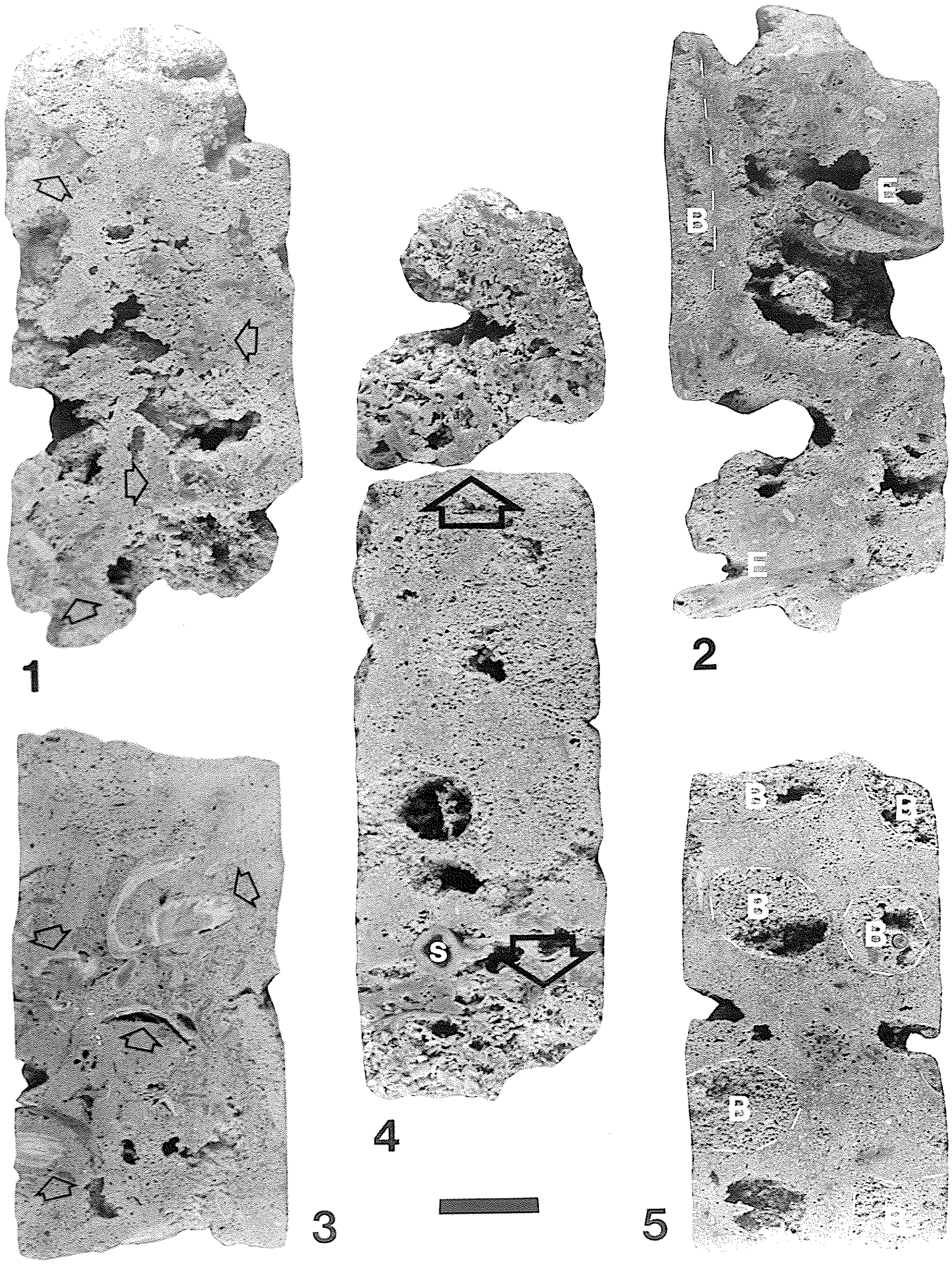
1
2
3
4
5
B
E
E
S
B
B
B
B
B

PLATE 3

OOID GRAINSTONE TO PACKSTONE FACIES - Q5 UNIT

3-1. *Halimeda*-rich ooid grainstone. Scale bar 2 mm. (Thin section, negative print, MO-6-L).

3-2. Increase in bioclast content (especially *Halimeda*) occurs with depth in northwest Big Pine Key. Note gastropod fragment (G). Bimodal grain size distribution ranges from very coarse sand size to granule size bioclasts and fine to medium sand size ooids, pellets, and peloids. Scale bar 3 mm. (Thin section, negative print, MO-6-M).

3-3. Packstone interval in borehole MO-6. Micrite appears as massive grey areas. Plate 3-4 is detail of outlined area. Scale bar 3 mm. (Thin section, negative print, MO-6-H).

3-4. Detail of Plate 3-3. Scale bar 1 mm. (Thin section, microphotograph, MO-6-H).

3-5. Burrow in ooid-bioclast-peloid packstone to grainstone. Infilling is a moderately sorted ooid-pellet grainstone. Mud lining suggests burrower may have been the shrimp *Callianassa*. Scale bar 3 mm. (Thin section, negative print, BP-7-L).

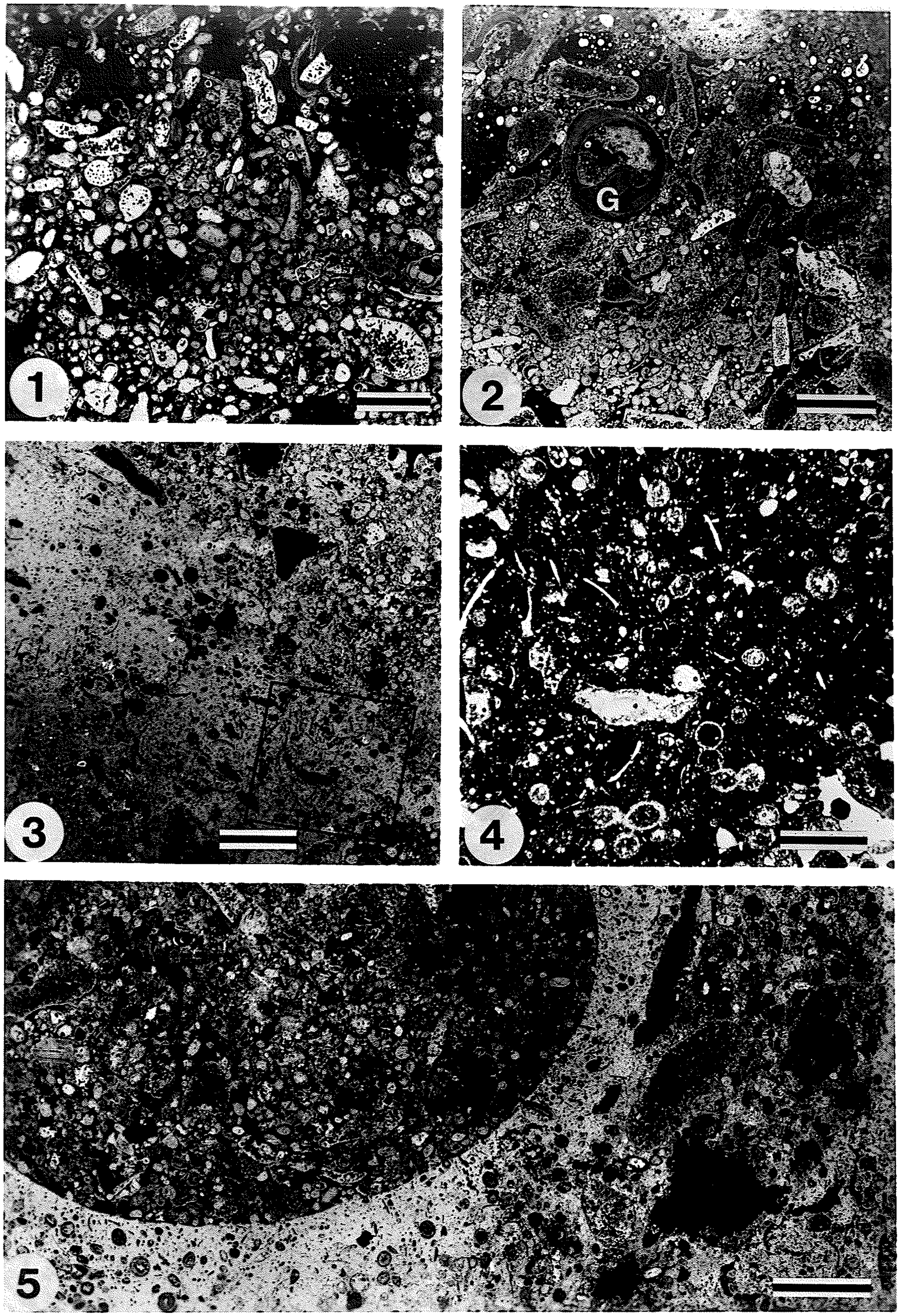
1
2
G
3
4
5

PLATE 4

PELOID-BIOCLAST PACKSTONE TO GRAINSTONE FACIES - Q5 UNIT
(Core, scale bar 2 cm)

4-1. Typical poorly sorted peloid-bioclast packstone to grainstone. Note large pelecypod (P), gastropod (G) fragments. (BP-4-12).

4-2. Large bored pelecypod (arrow) in typical poorly sorted peloid-bioclast (*Halimeda*-rich) packstone to grainstone. (BP-3-37).

4-3. Moderately sorted peloid-ooid-bioclast grainstone. This lithology becomes more poorly sorted higher in facies (compare with Plate 4-1). (BP-4-60).

4-4. Large encrusting bryozoan, *Schizoporella floridana* and, worm (?) tubes (arrows) in peloid-bioclast packstone to grainstone. (BP-4-23).

4-5. Large *Porites* sp. (P) and *Montastrea annularis* (M) fragments in peloid-bioclast packstone to grainstone. (BP-3-13).

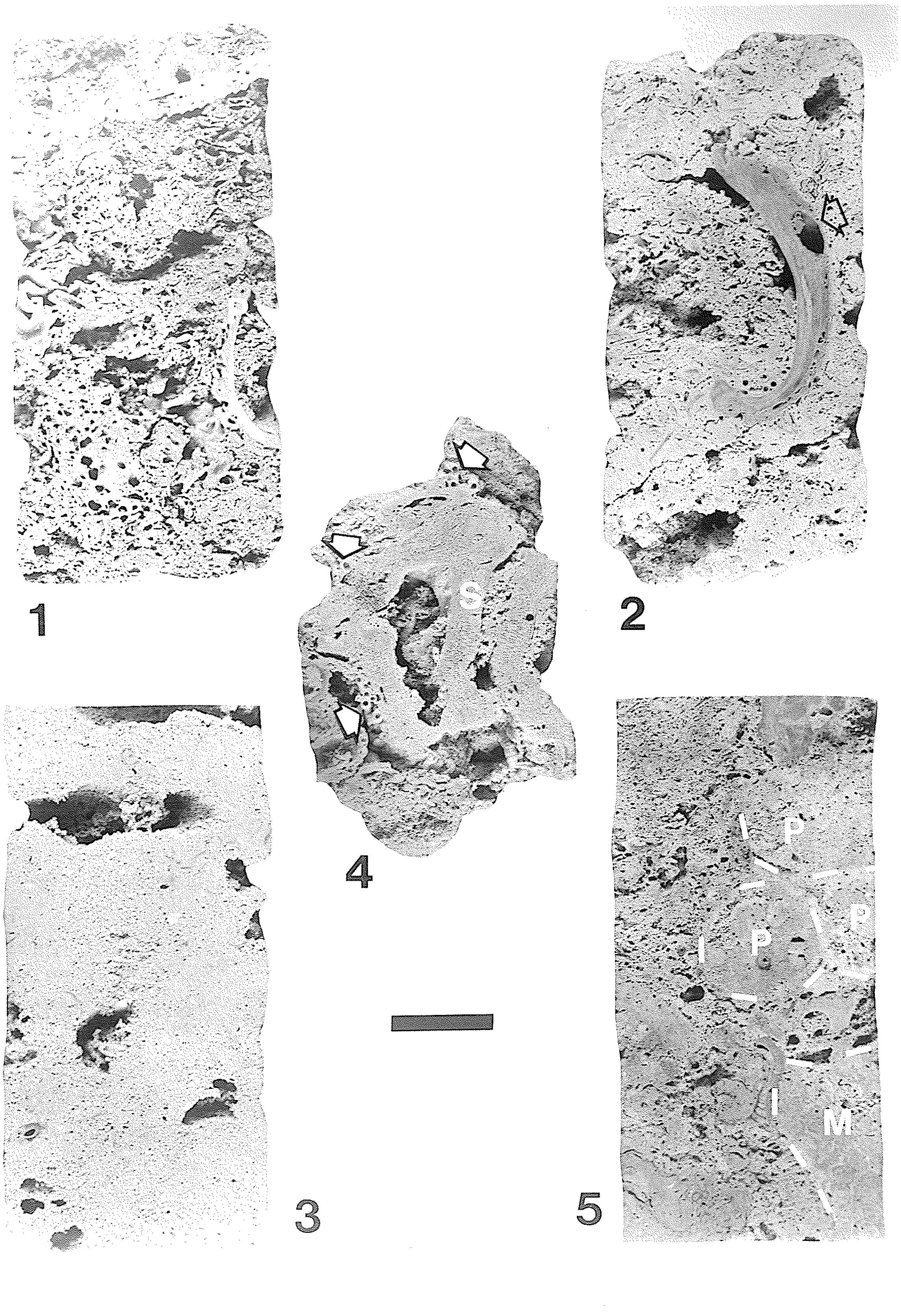
1
2
S
4
3
5
P
P
P
M

PLATE 5

PELOID-BIOCLAST PACKSTONE TO GRAINSTONE FACIES - Q5 UNIT

5-1. Typical peloid-bioclast packstone to grainstone. Note abundant Halimeda. Scale bar 3 mm. (Thin section, negative print, BP-1-G).

5-2. Detail of Plate 5-1. Scale bar 0.5 mm. (Thin section, microphotograph, BP-1-G).

5-3. Abundant Halimeda in poorly sorted peloid-bioclast packstone to grainstone. Scale bar 3 mm. (Thin section, negative print, BP-3-F).

5-4. Detail of Plate 5-3. Peloids appear to merge and form an indistinct, micritic matrix. Scale bar 1 mm. (Thin section, microphotograph, BP-3-F).

5-5. Bioclast-rich packstone to grainstone. Note fragments of Millepora sp. (M), Homotrema rubrum (H), peneroplid foraminifera (F), coralline algae (C), and pelecypods (P). Scale bar 2 mm. (Thin section, negative print, BP-3-B).

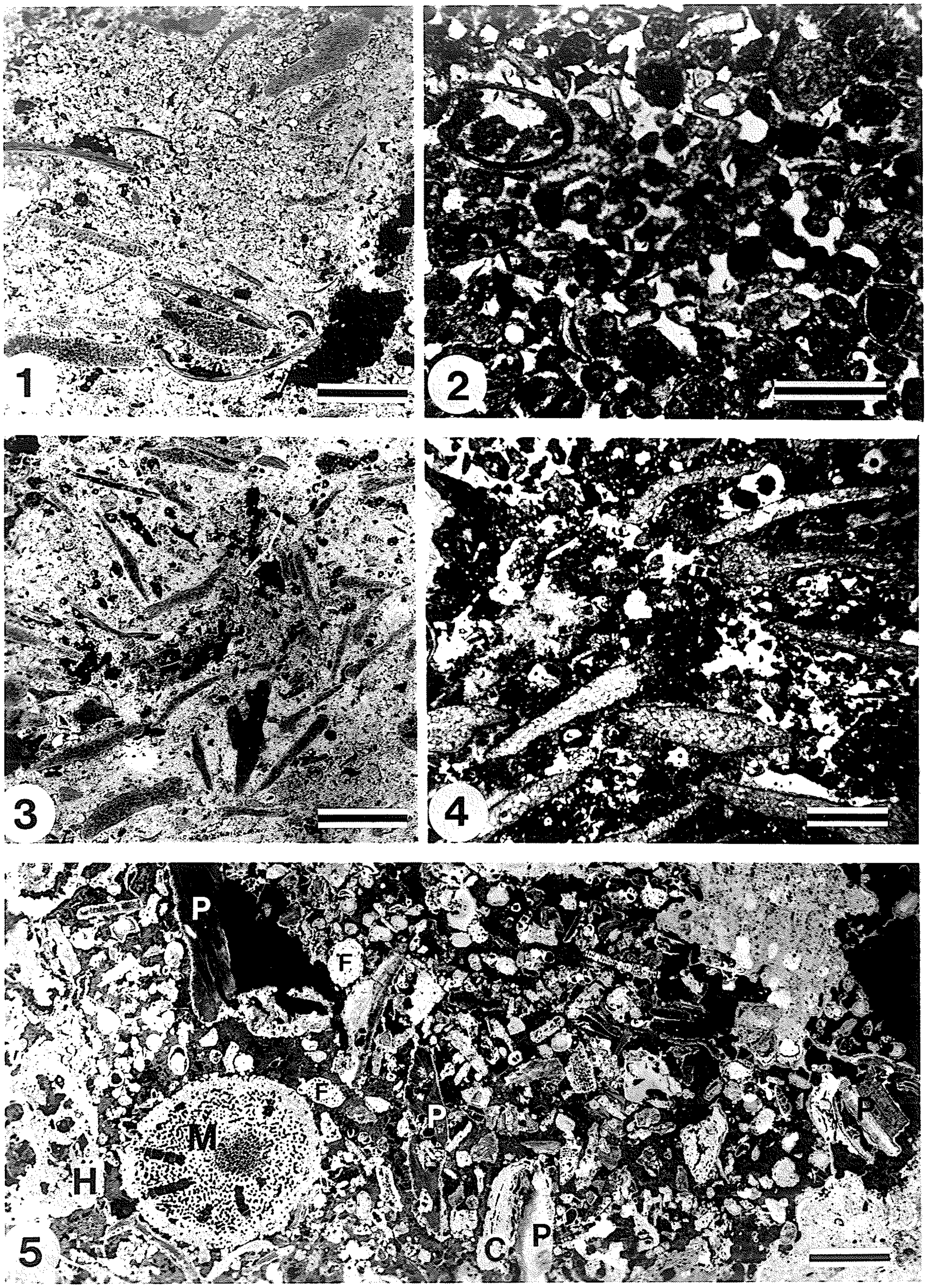
1
2
3
4
5
P
F
F
P
P
M
H
C
P

PLATE 6

PELOID-BIOCLAST PACKSTONE TO GRAINSTONE FACIES - Q4 UNIT

6-1. Typical peloid-bioclast packstone to grainstone. Note numerous mollusc shell fragments (arrows). Scale bar 5 mm. (Core, BP-3-57).

6-2. *Montastrea annularis* from peloid-bioclast packstone to grainstone facies. Scale bar 1 cm. (Core, BP-3-78).

6-3. Peneroplid foraminifera (F) in typical poorly sorted peloid-bioclast packstone to grainstone. Scale bar 0.5 mm. (Thin section, microphotograph, BP-6-L).

6-4. Large bioclasts in peloid-bioclast packstone to grainstone. Porous network in lower right of photo is a section through a *Spondylus* sp. shell(S). Upper left of photo shows *Montastrea annularis* (M) with mud-infilled corallites (light grey areas). Scale bar 5 mm. (Thin section, negative print, BP-3-K).

6-5. Peloid-ooid grainstone to packstone. Scale bar 0.5 mm. (Thin section, microphotograph, BP-7-J).

6-6. Moderately sorted peloid-bioclast grainstone. Most grains are fragmented and intact peneroplid and miliolid foraminifera. Scale bar 0.5 mm. (Thin section, microphotograph, BP-1-J).

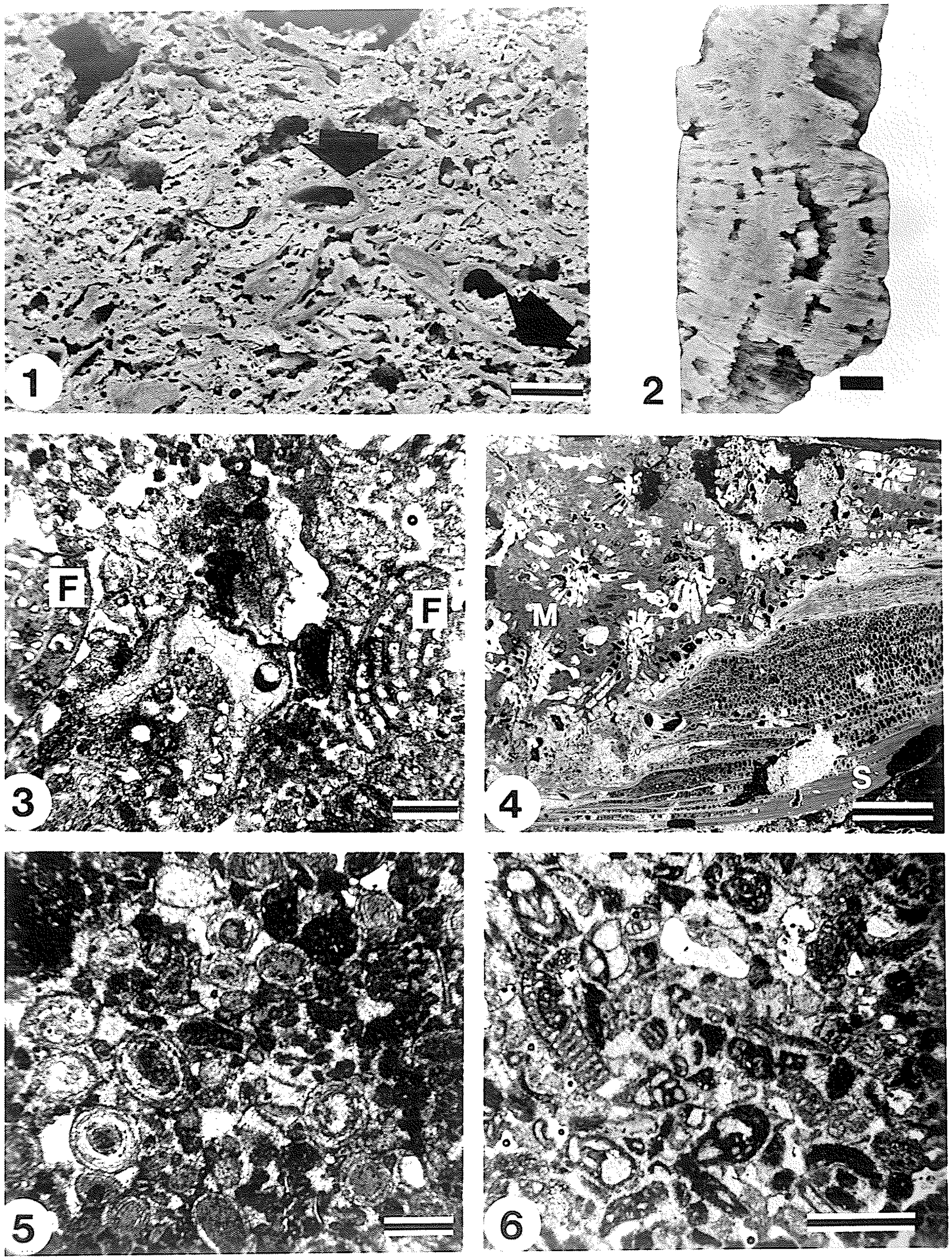
1
2
F
F
3
M
S
4
5
6

PLATE 7

PELOID-BIOCLAST PACKSTONE TO GRAINSTONE FACIES - Q3 UNIT (Core, scale bar 2 cm).

7-1. *Porites* sp. (P) and abundant *Halimeda* (arrows) in well cemented packstone. (see Plate 8-4). (MO-6-86).

7-2. *Porites* sp. (P) in well cemented packstone. (MO-6-81).

7-3. Abundant mollusc fragments (arrows) in well cemented packstone. (MO-2-78).

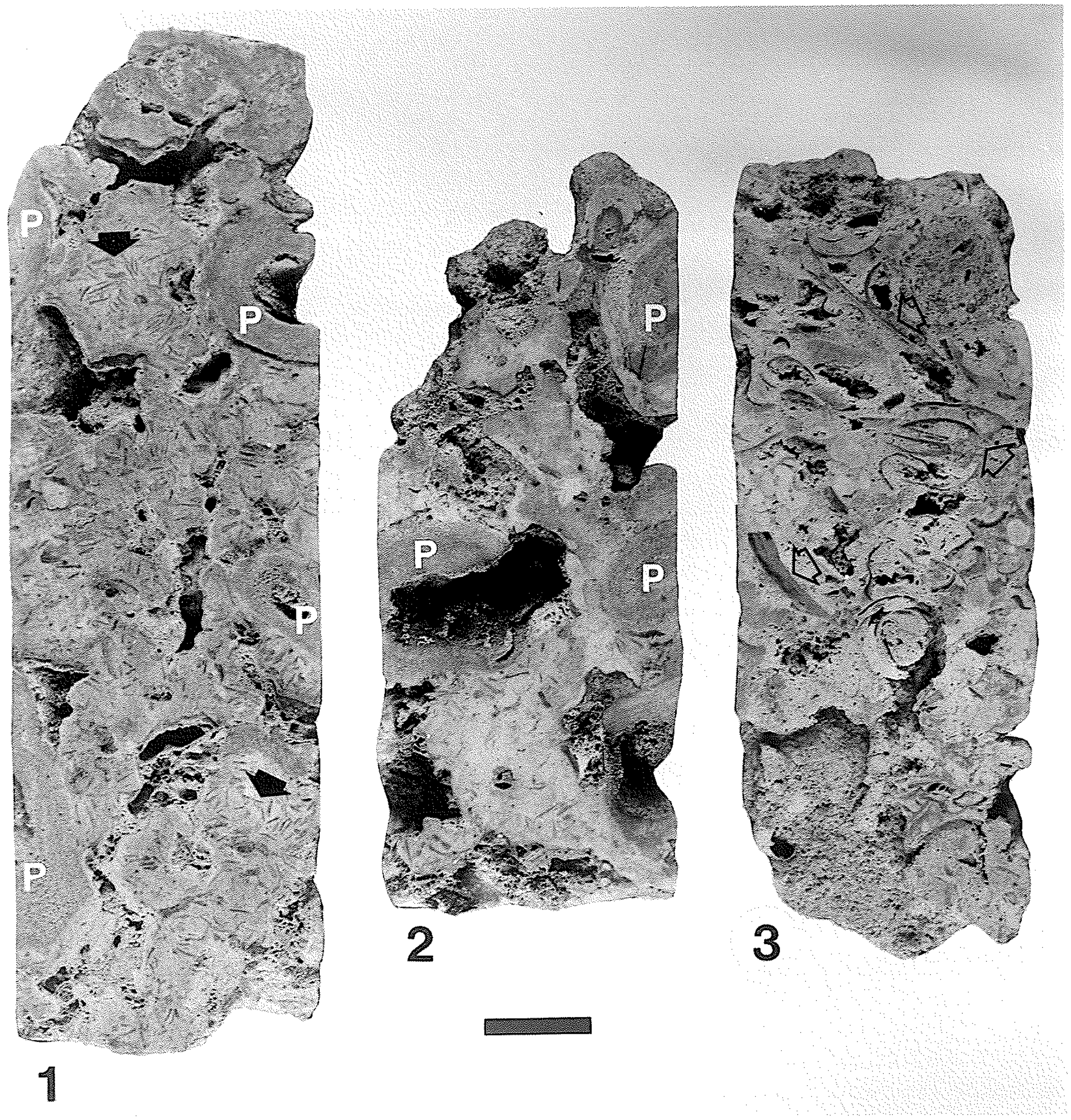
P
P
P
P
P
P
P
P
1
2
3

PLATE 8

PELOID-BIOCLAST PACKSTONE TO GRAINSTONE FACIES - Q3 UNIT

8-1. Typical peloid-bioclast packstone to grainstone. Note large amount of secondary (moldic) porosity (black). Scale bar 2 mm. (Thin section, negative print, MO-2-P).

8-2. Detail of Plate 8-1. Scale bar 0.5 mm. (Thin section, microphotograph, MO-2-P).

8-3. *Halimeda* (H) and foraminifera-rich (F) peloid-bioclast grainstone. Scale bar 2 mm. (Thin section, negative print, MO-6-R).

8-4. Well cemented packstone matrix of coralline packstone (see Plate 7-1). Precursor grains are not identifiable. Scale bar 0.2 mm. (Thin section, microphotograph, crossed nicols, MO-6-S).

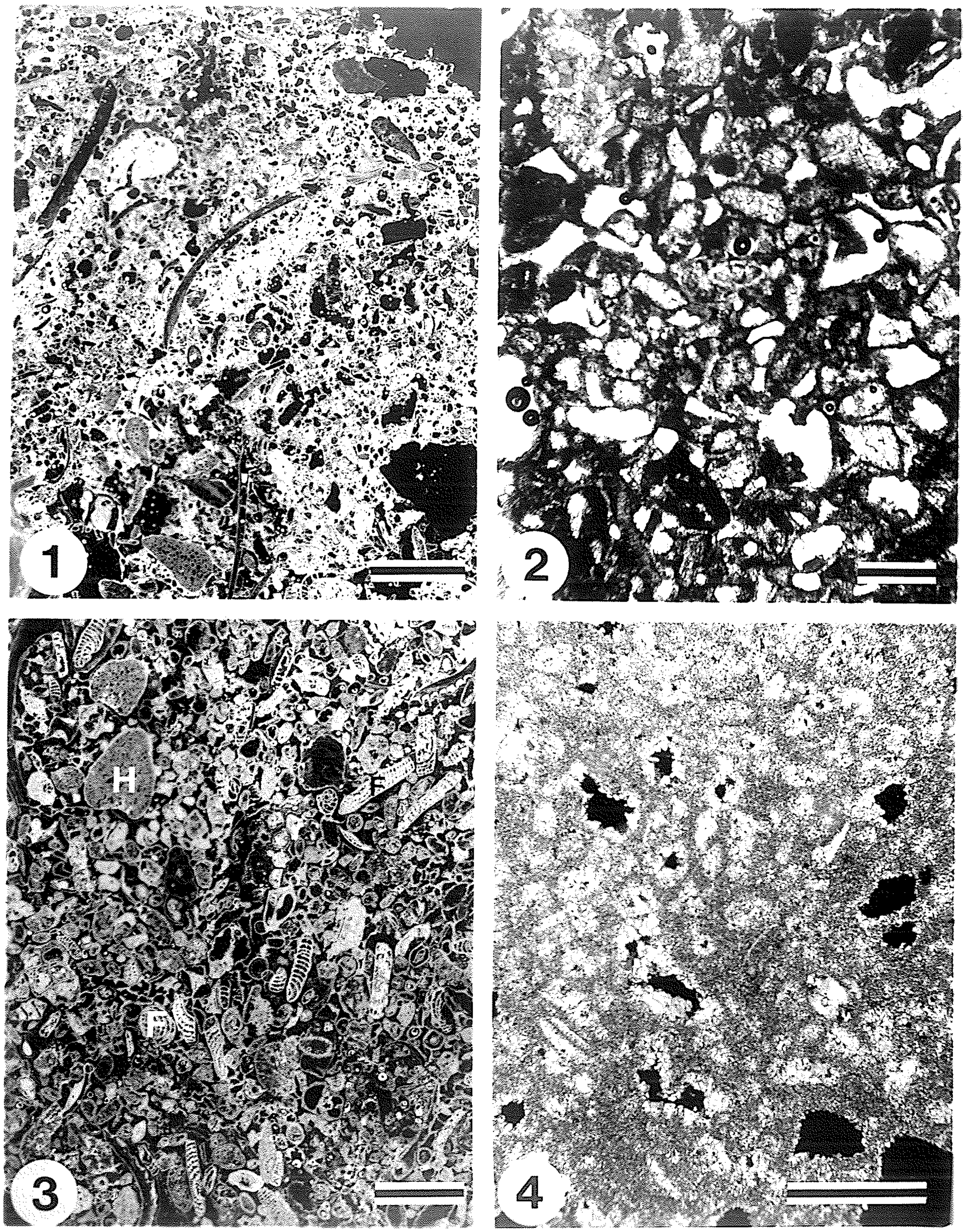
1
2
H
F
3
4

PLATE 9

SUBMARINE DIAGENESIS: CEMENTATION
(Q5 unit)

9-1. Isopachous fringe of interparticle submarine cement in ooid grainstone to packstone. Scale bar 0.2 mm. (Thin section, microphotograph, MO-2-G).

9-2. Intraparticle submarine cement in pelecypod (or ostracode?). Parallel arrangement of cement fibres suggests substrate control. Cement occurs in ooid grainstone. Note coarseness of intraparticle spar relative to interparticle spar. Scale bar 0.2 mm. (Thin section, microphotograph, BP-7-D).

9-3. Optically continuous submarine cement overgrowing aragonitic coral (*Diploria* sp.). Arrow points to coral-cement contact. Scale bar 0.1 mm. (Thin section, microphotograph, H-2).

9-4. Submarine cement of Plate 9-3. Compare with Plate 9-5. Arrangement of the tabular crystals suggests substrate control. Scale bar 20 μm. (SEM photograph, H-2).

9-5. Composite needle submarine cement of Plate 9-3. Compare with Plate 9-4. Scale bar 50 μm. (SEM photograph, H-2).

9-6. Detail of Plate 9-5. Note individual crystals (1,2,3,4) which comprise composite needle. Also note rhombohedral cement overgrowing submarine cement. Scale bar 10 μm. (SEM photograph, H-2).

9-7. Submarine cement infilling *Halimeda* utricle. Note rhombohedral spar further occluding porosity. Scale bar 10 μm. (SEM photograph, MO-6-L).

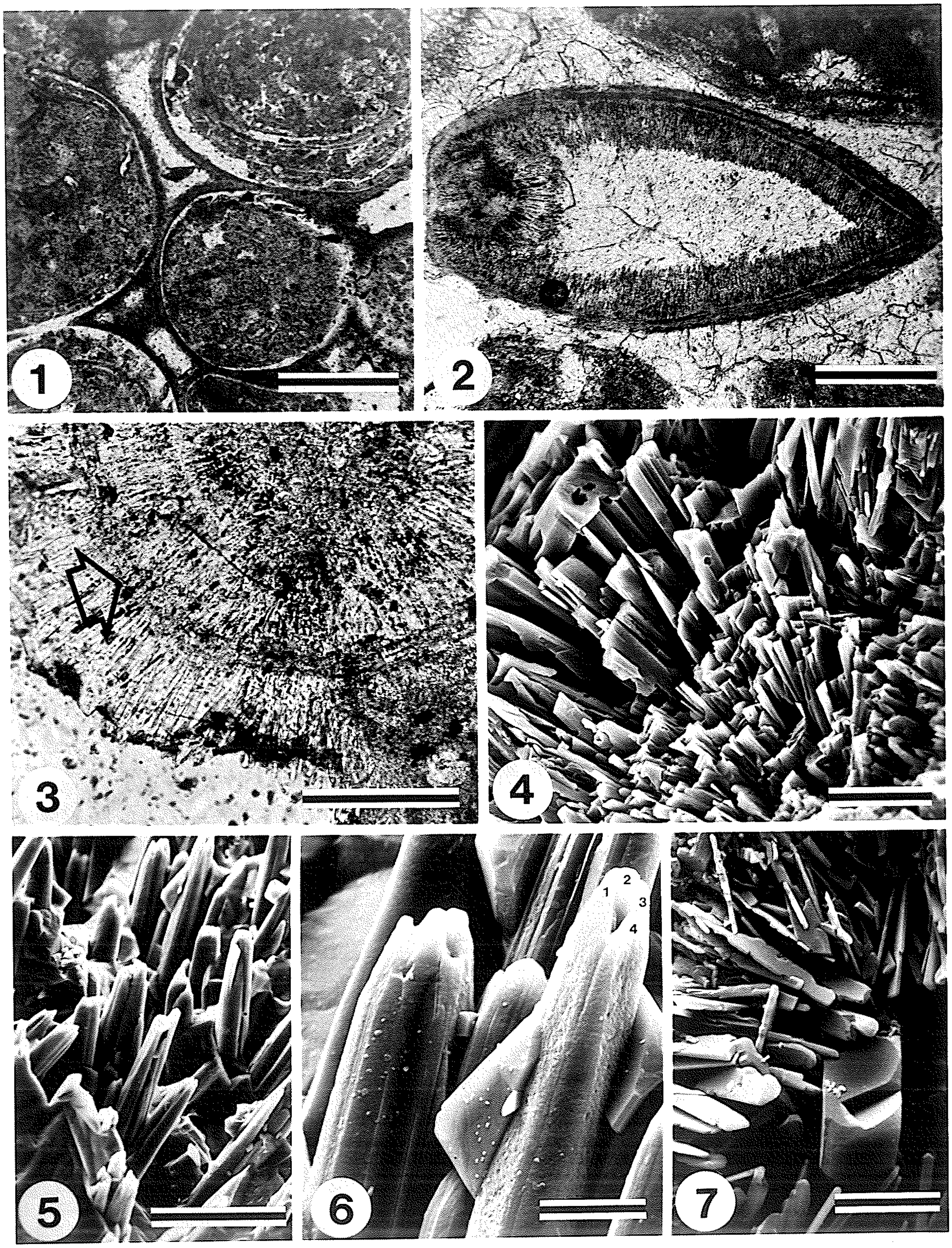
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PLATE 10

SUBMARINE DIAGENESIS: OOID MICRITIZATION
(Q5 unit)

10-1. Distribution of algal microborings in ooid cortex (arrows) which is entirely replaced by calcite spar (except for microboring/molds). Scale bar 0.2 mm. (Thin section, microphotograph, BP-7-K).

10-2. Surface texture resulting from dissolution of ooid cortex. Arrow points to microboring partially filled with submarine cement. Scale bar 50 μm. (SEM photograph, MO-6-D).

10-3. The most common algal microboring mold (A) is typically irregularly shaped. Rare filamentous molds (B) of uncertain origin. Scale bar 10 μm. (SEM photograph, MO-6-A,B).

10-4. Detail of rare filamentous microboring mold consisting of randomly oriented aragonite rods. Scale bar 2 μm. (SEM photograph, MO-2-A).

10-5. Common spherical to subspherical algal microboring molds. Scale bar 0.1 mm. (SEM photograph, MO-6-L).

10-6. Detail of molds from Plate 10-5. The microboring molds are partially replaced by calcite as demonstrated by rhombohedral crystals (arrow). Scale bar 20 μm. (SEM photograph, MO-6-L).

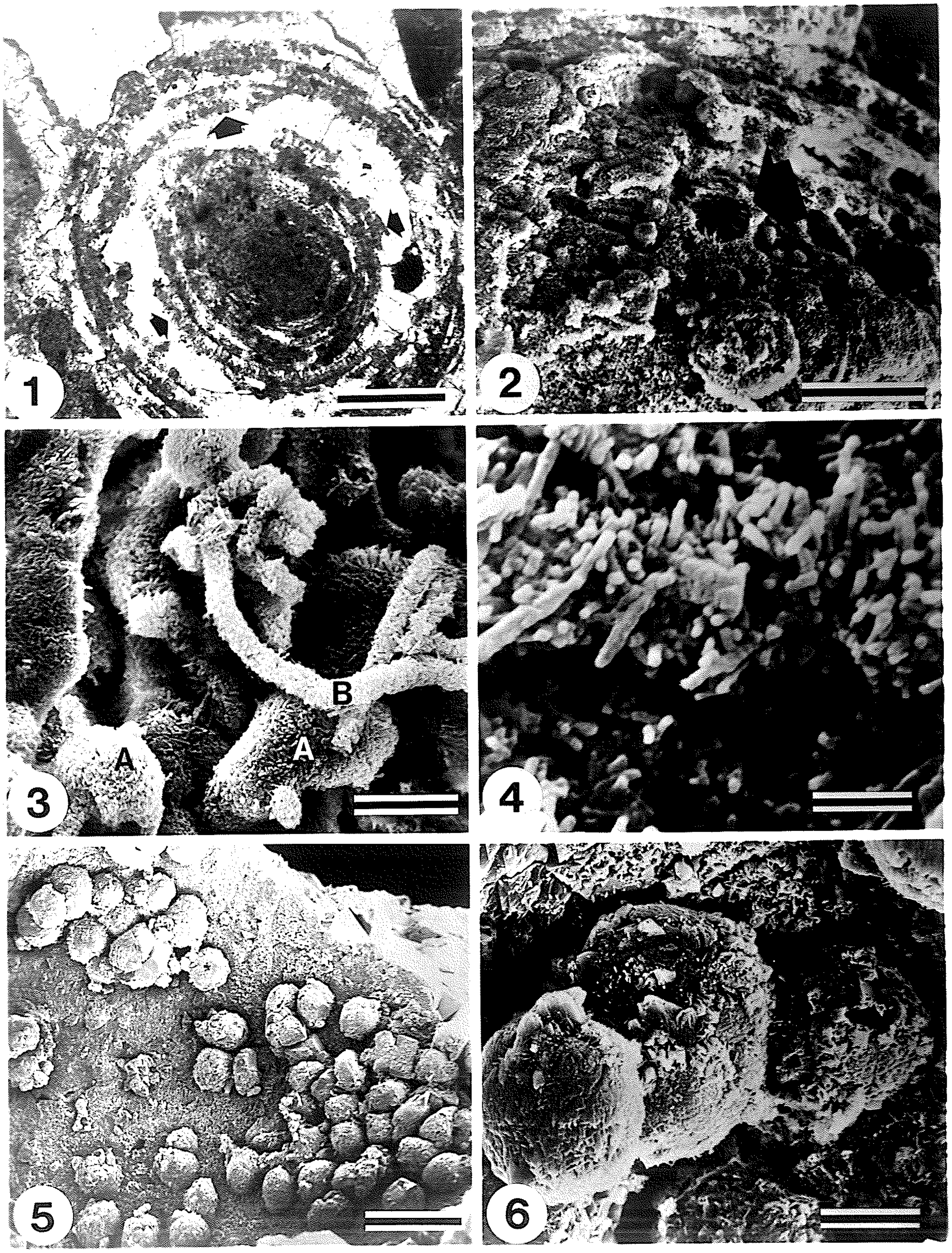
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PLATE 11

CALICHE FABRICS

(Core and outcrop samples, Q5 unit surface, except where noted; scales as shown).

11-1. Finely laminated non-porous crust. Laminae thicken in depressions and thin over highs. Q4 unit surface. (MO-6-H,I).

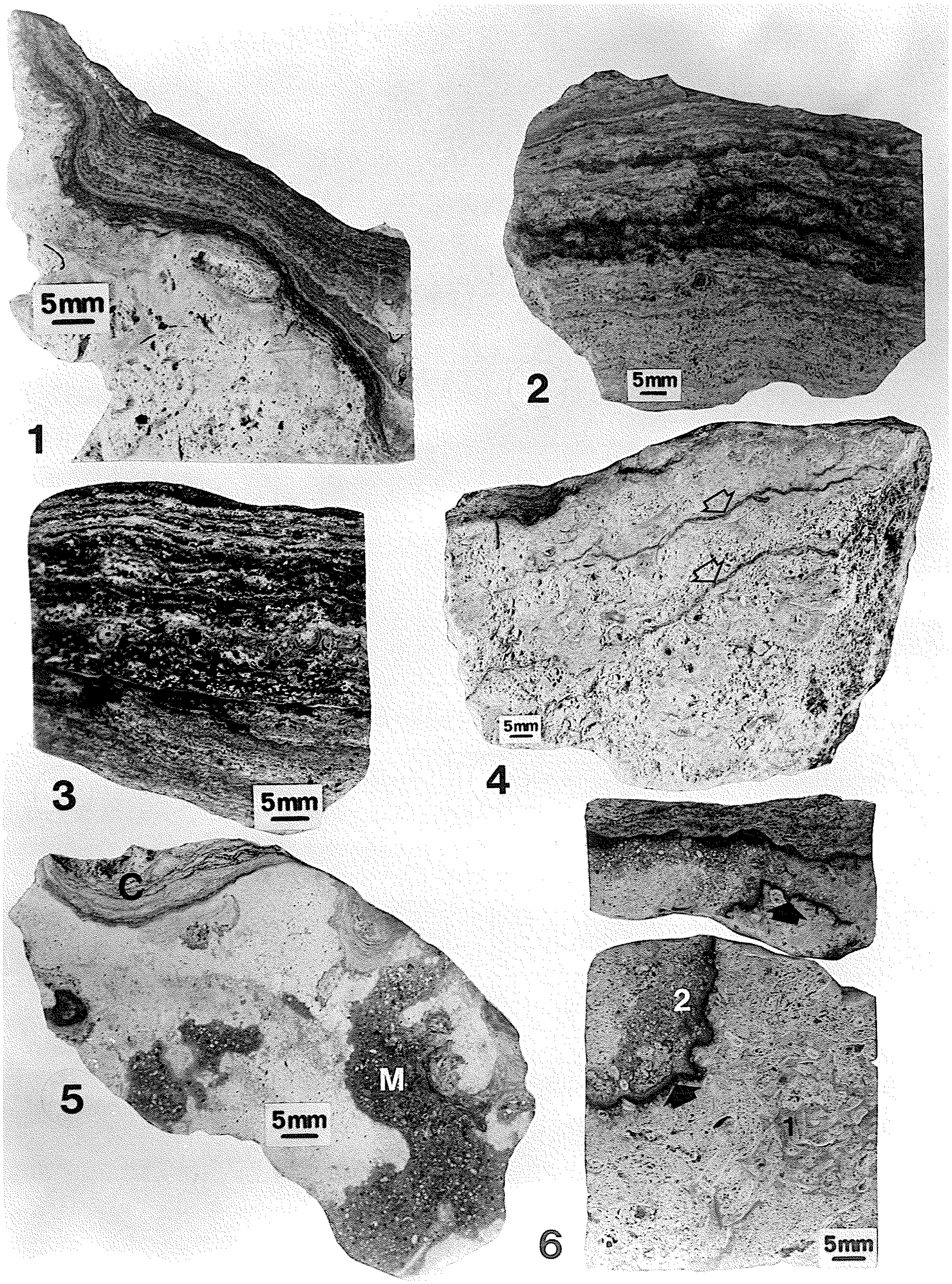
11-2. Irregular, diffusely laminated crust. (H-8).

11-3. Finely laminated crust. Numerous tubular pores were produced by roots or worms. Note truncation of lower laminae by horizontal upper laminae. (BP-6-A).

11-4. Subsurface micrite stringers (arrows) developed in peloid-bioclast packstone to grainstone. Thin surface crust is non-porous and non-laminated. (H-3).

11-5. Distinct laminated crust (C) and irregularly shaped micrite cement zone (M). (H-9).

11-6. Minor subaerial surface (arrow) developed on peloid-bioclast packstone to grainstone (1) acts as a permeability barrier to later deposited peloid-ooid packstone to grainstone (2). Micrite coatings on grains in (2) are more pronounced than in (1). (BP-2-B,C).


5mm
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5mm

PLATE 12

COMPONENTS OF CALICHE: CEMENTS, PELOIDS, AND MICROBORINGS (Q5 unit surface)

12-1. Fibrous calcite cement botryoids occlude irregular vuggy porosity in laminated crust. Scale bar 0.2 mm. (Thin section, microphotograph, crossed nicols, H-15).

12-2. Random needle fibres in peloidal grainstone to packstone. Scale bar 0.2 mm. (Thin section, microphotograph, crossed nicols, BP-2-G).

12-3. Alternating light and dark micritic laminae of laminated crust. Note numerous pin-point vugs (white). Scale bar 1 mm. (Thin section, microphotograph, H-21a).

12-4. Peloids and clotted texture of micrite cement zone. Scale bar 0.2 mm. (Thin section, microphotograph, crossed nicols, MO-6-A).

12-5. Fungal microborings in laminated crust (arrows). Unidentified hopper crystals in centre of photo. Scale bar 50 μm. (SEM photograph, H-9).

12-6. Fungal microborings on grain surface in micrite cement zone. Note prominent micrite cement coat (M) on grain. Scale bar 0.1 mm. (SEM photograph, MO-2-A).

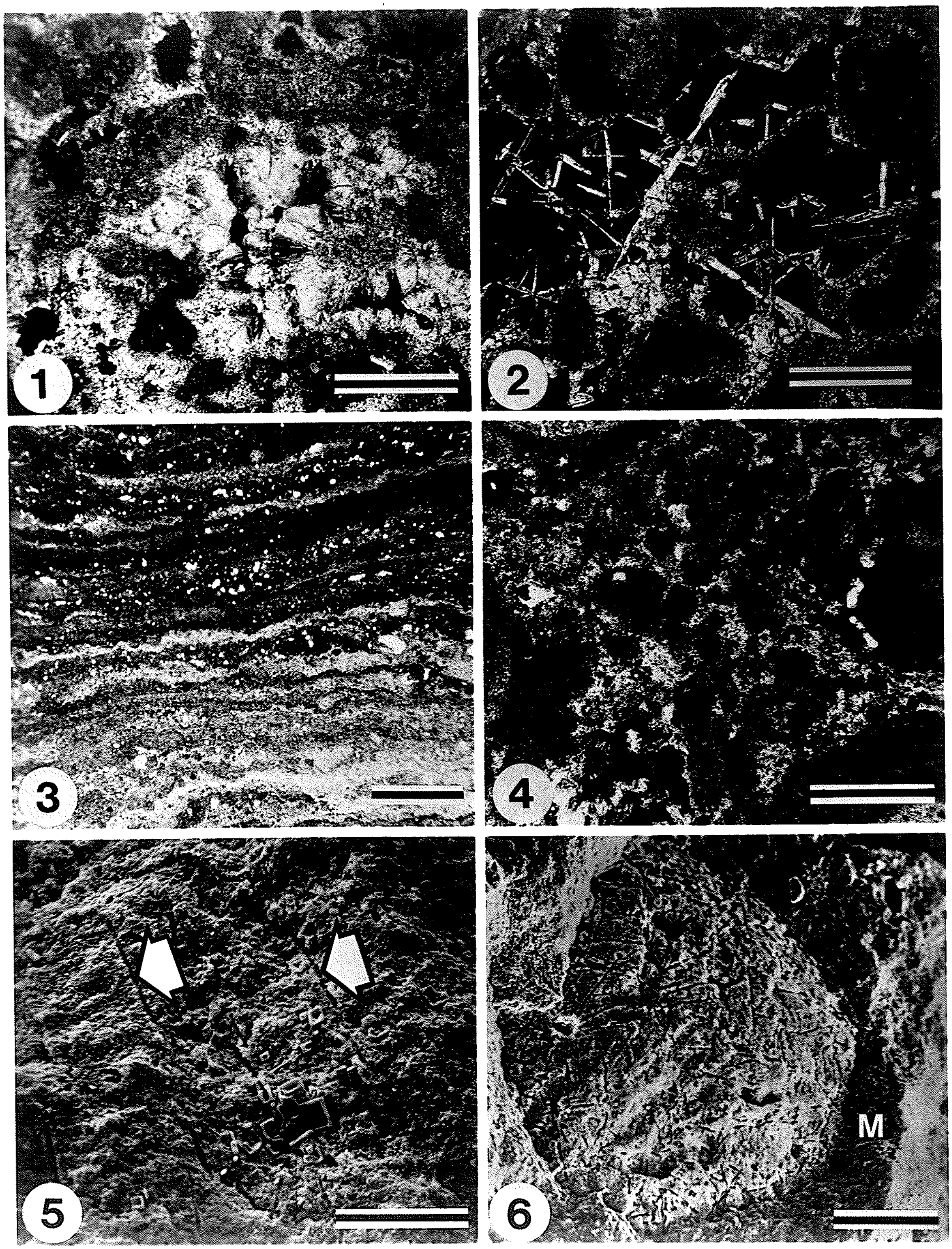
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PLATE 13

COMPONENTS OF CALICHE: CALCIFIED FILAMENTS (Q5 unit surface, except where noted).

13-1. Small filaments branching radially outward from large filament suggest root hair origin. Filaments occur within a soil infilling secondary porosity. Q4 unit surface. Scale bar 0.2 mm. (Thin section, microphotograph, BP-2-M).

13-2. Chasmoliths branching from large filament in interparticle pore of ooid grainstone to packstone. Scale bar 0.2 mm. (Thin section, microphotograph, BP-1-B).

13-3. Calcified filaments and opaque and translucent material (right half of photo) is a soil which infills secondary porosity in highly altered rock (left half of photo). Q4 unit surface. Scale bar 0.2 mm. (Thin section, microphotograph, BP-2-M).

13-4. Chasmoliths and endoliths (arrow) in micrite cement zone. Note variable diameter of calcified filaments. Scale bar 0.2 mm. (SEM photograph, MO-6-A,B).

13-5. Detail of chasmoliths in Plate 13-4. Scale bar 50 µm. (SEM photograph, MO-6-A,B).

13-6. Equant micrite crystals which comprise calcified filament. Scale bar 5 µm. (SEM photograph, MO-6-A,B).

13-7. Porous 'wisp' network caused by micrite cement precipitation on calcified filaments. Scale bar 0.2 mm. (Thin section, microphotograph, BP-1-A).

13-8. 'Wisps' produced by micrite cement precipitation on calcified filaments. Scale bar 20 µm. (SEM photograph, MO-2-A).

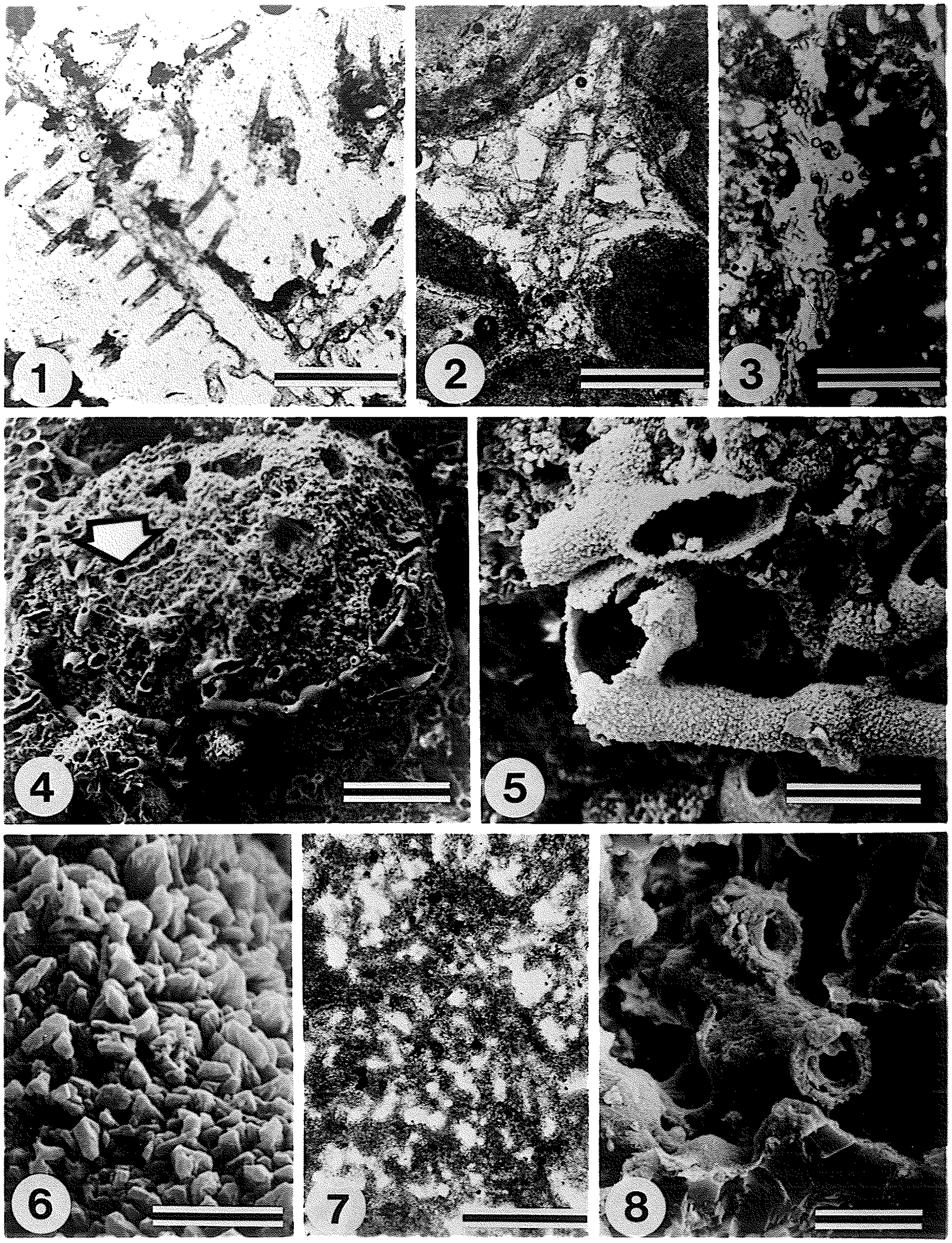
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PLATE 14

MICRITE CEMENT ZONE FABRICS
(Q5 unit surface, except where noted).

14-1. Varying colours of micrite cement which coat a highly altered ooid. Scale bar 0.2 mm. (Thin section, microphotograph, MO-2-A).

14-2. Pores in micrite coating suggest micrite precipitation over chasmoliths. Compare with Plate 13-4. Scale bar 0.2 mm. (Thin section, microphotograph, MO-6-B).

14-3. Typical micrite cement zone fabric with coated grains and coated calcified filaments ('wisps'). Scale bar 0.5 mm. (Thin section, microphotograph, MO-2-A).

14-4. Micrite cement zone with abundant coated grains sharply abutting relatively unaltered sediment. A small micritic stringer (arrow) is the permeability barrier which caused this juxtaposition. Scale bar 2 mm. (Thin section, negative print, MO-2-A).

14-5. Grain coatings preserve outlines of grains which have been leached. The pores were subsequently infilled by sparry cement. In this sample, the micritic cement grain coats have a high proportion of tangentially oriented micrometre - size needles. Q4 unit surface. Scale bar 0.2 mm. (Thin section, microphotograph, BP-7-H).

14-6. Micrite coated grain. Scale bar 0.2 mm. (SEM photograph, MO-2-A).

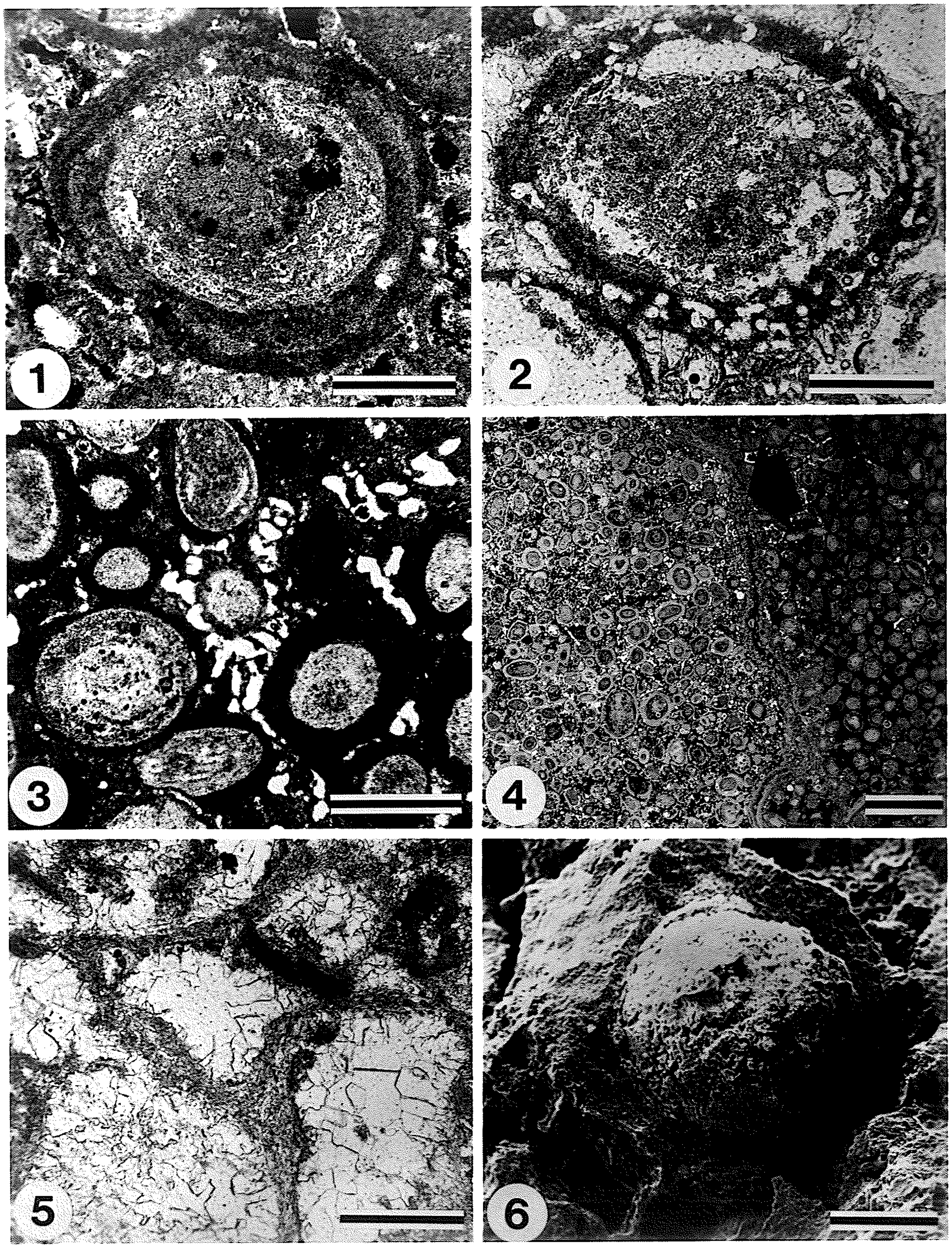
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PLATE 15

SUBSURFACE DIAGENESIS: OOID ALTERATION
(Q5 unit, except where noted).

15-1. Partially leached ooid with inner cortex laminae well preserved (note psuedo-uniaxial isogyre). Secondary porosity (P) developed in outer laminae partially occluded by calcite cement (C). Scale bar 0.2 mm. (Thin section, microphotograph, crossed nicols, BP-7-D).

15-2. Ooid cortex and nuclei replaced by coarsely crystalline calcite. Original concentricity of cortex laminae preserved by algal microboring molds. Low density of microborings suggests possible replacement by calcite. Q4 unit. Scale bar 0.2 mm. (Thin section, microphotograph, BP-7-I).

15-3. Incipient cortex alteration. Selected cortex laminae are preferentially leached (arrows). Compare with Plate 15-4. Scale bar 50 µm. (Thin section, microphotograph, MO-2-A).

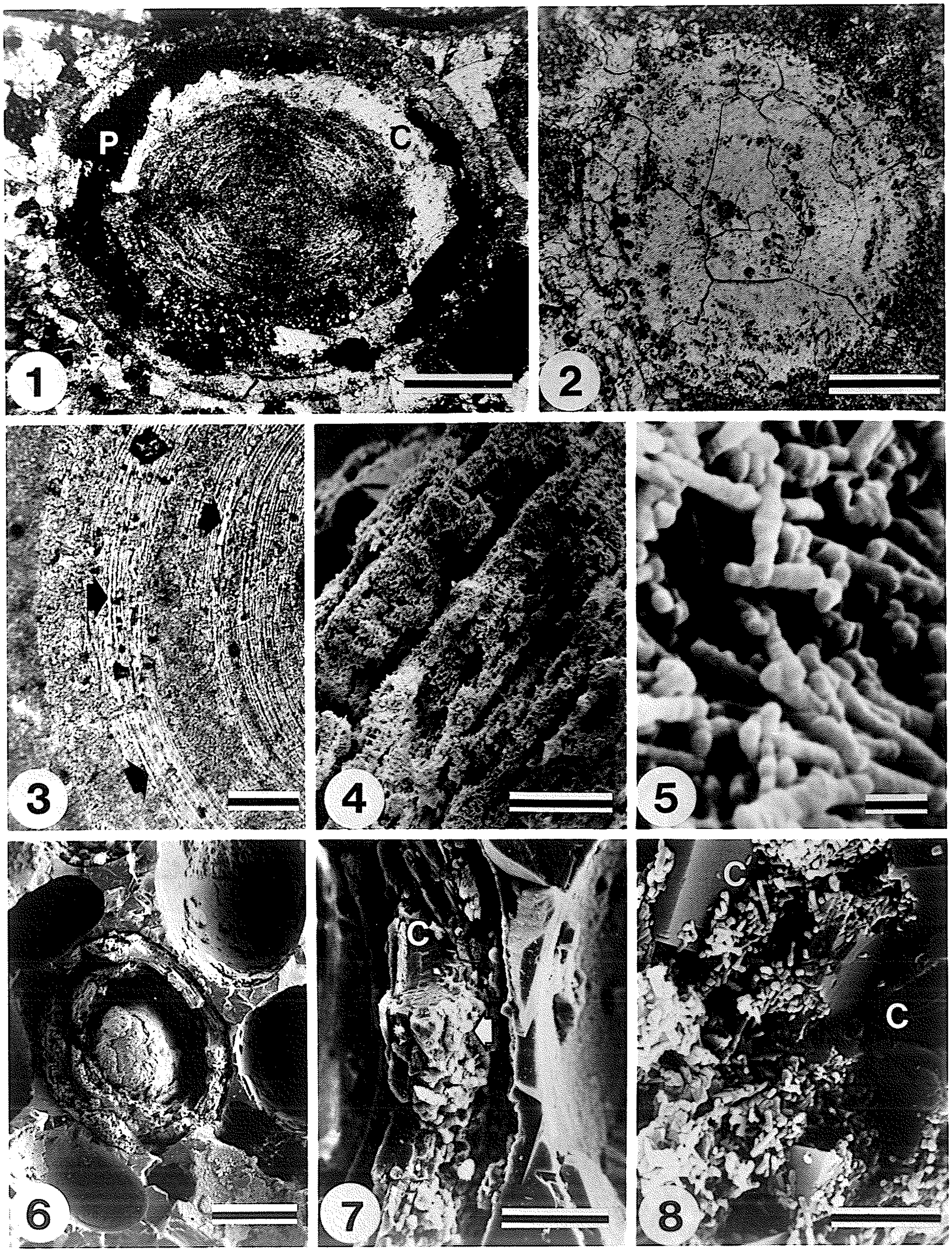
15-4. SEM detail of cortex alteration shown in Plate 15-3. Scale bar 20 µm. (SEM photograph, H-5).

15-5. Cortex aragonite rods. Note approximate tangential orientation and large amount of inter-rod porosity. The irregular shape of the rods may result from incipient dissolution. Scale bar 1 µm. (SEM photograph, BP-7-C).

15-6. Development of oomolidic porosity with minor calcite cementation. See Plate 15-7 for detail. Scale bar 0.2 mm. (SEM photograph, BP-7-C).

15-7. Detail of outer ooid cortex from Plate 15-6. Calcite cement (C) has partially occluded secondary porosity. Microboring molds (arrow) are apparently unaltered. Scale bar 50 µm. (SEM photograph, BP-7-C).

15-8. Rhombohedral calcite cement crystals (C) growing in pore spaces of the ooid cortex. Scale bar 5 µm. (SEM photograph, BP-7-F).


P
C
1
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5
6
C
7
C
C
8

PLATE 16

SUBSURFACE DIAGENESIS: CEMENTATION AND NEOMORPHISM (Q5 unit, except where noted).

16-1. Bladed spar (arrow) on peloid in ooid grainstone facies. Equant spar which fills most of interparticle pore space is typical. Scale bar 0.2 mm. (Thin section, microphotograph, BP-7-G).

16-2. Pore filling fabric and meniscus texture in ooid-bioclast grainstone. Micrite cement (M) postdates calcite spar cement and is included in meniscus texture. Scale bar 0.5 mm. (Thin section, microphotograph, crossed nicols, BP-7-K).

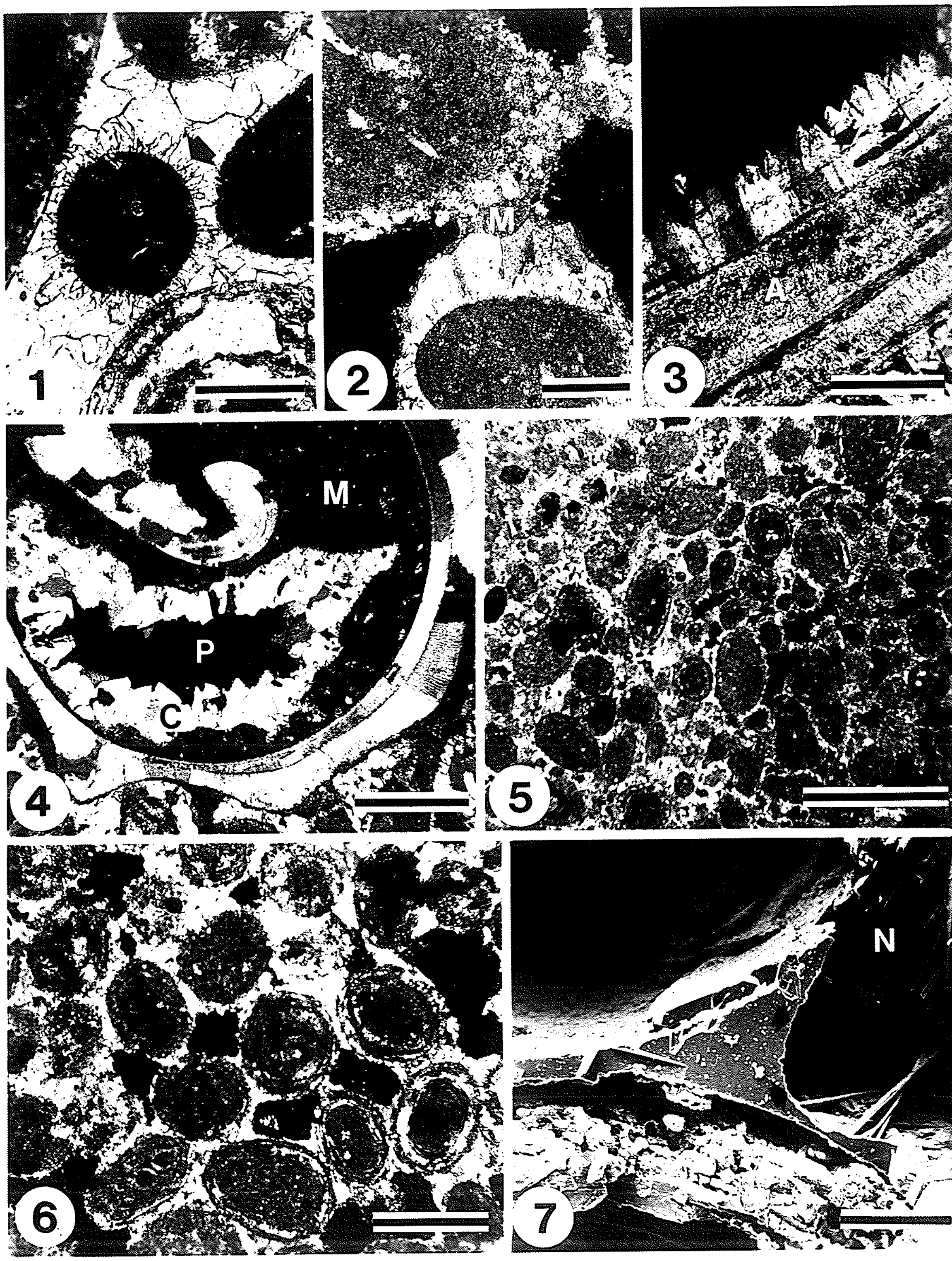
16-3. Bladed calcite is both neomorphic replacement of mollusc shell prismatic layer and scalenohedrally terminated calcite cement. Arrows point to relict organic textures in shell fragment. Unaltered mollusc aragonite (A). Scale bar 0.2 mm. (Thin section, microphotograph, crossed nicols, BP-7-K).

16-4. Leaching of internal micrite sediment (M) producing secondary pore (P) in neomorphosed gastropod. Porosity partially occluded by calcite cement (C). Q3 unit. Scale bar 0.5 mm. (Thin section, microphotograph, crossed nicols, MO-6-U).

16-5. Typical vadose zone cementation heterogeneity in grainstone. Scale bar 1 mm. (Thin section, microphotograph, crossed nicols, MO-6-E).

16-6. Typical vadose zone mensicus cement texture in ooid grainstone. Scale bar 0.5 mm. (Thin section, microphotograph, crossed nicols, BP-7-C).

16-7. Pore-rounded meniscus cement (arrows). Note random needle fibres (N) in pore. Scale bar 50 μm. (SEM photograph, BP-7-C).


1
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M
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A
4
M
P
C
5
6
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N

PLATE 17

SUBSURFACE DIAGENESIS: CEMENTATION AND NEOMORPHISM (Q5 unit, except where noted).

17-1. Partial preservation of original texture in neomorphically altered mollusc shell fragment. Minor development of secondary porosity (P). Scale bar 0.5 mm. (Thin section, microphotograph, MO-2-G).

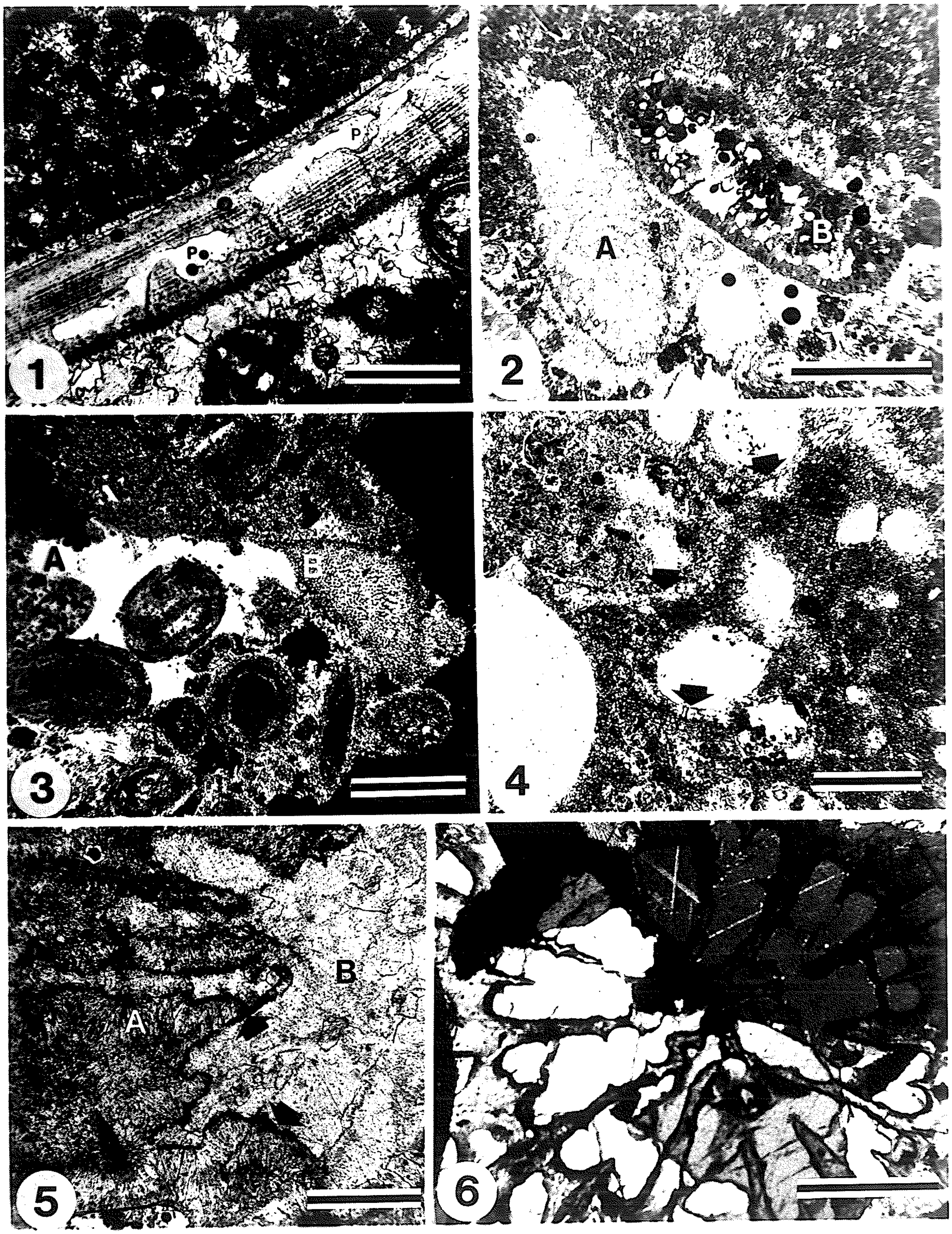
17-2. Completely neomorphosed Halimeda plate (A) adjacent to largely unaltered Halimeda plate (B). Scale bar 1 mm. (Thin section, microphotograph, MO-6-N).

17-3. Poikilotopic spar (white) is optically continuous with echinoderm fragments (A,B). Scale bar 0.5 mm. (Thin section, microphotograph, crossed nicols, BP-1-B).

17-4. Micrite neomorphically altered to microspar. Arrows point to neomorphic 'halos' around grains which have been leached. Scale bar 0.2 mm. (Thin section, microphotograph, BP-7-L).

17-5. Zone of chalky aragonite (arrows) separates apparently unaltered coral aragonite (A) from neomorphic spar (B). Coral is Diploria sp. Scale bar 0.5 mm. (Thin section, microphotograph, H-2).

17-6. Montastrea annularis illustrating cross cutting mosaic of coarse calcite spar. Q4 unit. Scale bar 1 mm. (Thin section, microphotograph, BP-3-M).



1
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APPENDIX A - THE KEY LARGO ENIGMA

INTRODUCTION

The origin of the Florida Keys has been a topic of great interest since the mid 1960's. The problem centres around the paleoenvironmental interpretation of the Key Largo Limestone. Based mainly on a literature survey, the problem is discussed here. The similarity of the Q5 Key Largo Limestone to the Q4 and Q3 reef limestones suggests that comparable arguments may be applicable to these older units.

DISTRIBUTION AND THICKNESS

The present-day Florida Keys form an arcuate chain of small islands with a maximum width of approximately 5 km (3 mi), which stretches from Soldier Key (a few kilometres south of Miami) to Key West, a distance of 240 km (150 mi). The islands are the expression of the sea level high stand of 125,000 years BP represented by the Q5 unit. According to Hoffmeister and Multer (1968, p. 1489), the greater area of the Keys lies only about 1 m (3 ft) above high tide level, and mangrove swamps occupy at least half of the total area. There is little surface relief. The northern Keys, however, may be more elevated along their central elongated axis. Maximum elevation is 5.5 m (18 ft) on Windley Key (Stanley, 1966, p. 1929).

The Key Largo Limestone comprises the bedrock of the upper and middle Keys whereas the bedrock of the lower Keys is Miami Limestone. The surface contact between the two formations occurs on Big Pine Key (Hoffmeister and Multer, 1964, p. 60; this work).

Boreholes have demonstrated that Key Largo Limestone-type lithologies underlie the lower Keys and extend as far as the Dry Tortugas, 113 km (70 mi) west of Key West. Here, Key Largo Limestone was found to lie approximately 9 m (30 ft) below sea level (Hoffmeister and Multer, 1968, p. 1490).

The thickness of the Key Largo Limestone is variable. Hoffmeister and Multer (1968, p. 1490) reported 44 m (145 ft) near the northern tip of Key Largo, but 16 km (10 mi) further south along the Key the thickness is 23 m (75 ft). A well at Tavernier (near Key Largo's southern tip) passes through 30.5 m (100 ft) of Key Largo Limestone. 52 m (170 ft) was reported from Grassy Key, while at Big Pine Key and Key West, more than 55 and 52 m (180 and 170 ft, respectively) were reported.

Perkins (1977, plate 3) shows the thickness of the Key Largo Limestone to be 30 m (97 ft) or more from boreholes on Big Pine Key, Windley Key, and Little Molasses Island. Stratigraphically, the Key Largo Limestone spanned the Q3 through Q5 units at these localities.

COMPOSITION AND ZONATION

The modern analogues with which the Key Largo Limestone is usually compared occur in the present-day Florida reef tract. The tract contains approximately 96 km (60 mi) of platform edge reefs located along the shallow shelf edge, and a 'lagoon' which contains over 6000 patch reefs (Marszalek et al., 1977, p. 224). The typical well developed platform edge is

> "characterized by a reef-flat formed of in situ dead encrusted Acropora palmata skeletons and rubble. Dominant reef-flat benthic macrobiota include small heads of Acropora, Porites, and Siderastrea, encrusting Millepora ... A fringe of massive oriented colonies of Acropora palmata forms the seaward face of the reef to a depth of about 4 m ... Deeper portions of the reef exhibit a diverse coral assemblage dominated by large heads of Montastrea annularis ..." (Marszalek et al., 1977, p. 224).

Reefs with Acropora palmata can also occur slightly landward (as far as 1.6 km [1 mi]) from the outer platform margin, but such is the case only when a large break exists in the outer reef chain (Shinn, 1963, p. 93). Patch reefs, on the other hand, lack the coral zonation found on the platform edge reefs, and significantly, lack Acropora palmata. However, the reef biota may otherwise be similar to that of the platform edge reefs (Marszalek et al., 1977, p. 227). The chief corals are large, rounded heads of Montastrea annularis, Diploria spp., and Siderastrea. Porites astreoides is an abundant encrusting form and the branching corals Porites and Acropora cervicornis may also occur (Stanley, 1966, p. 1937).

In a study based on the excellent exposures of Key Largo Limestone at the Windley Key quarry and the Key Largo Waterway, Stanley (1966) described the elements of the Key Largo Limestone and found that the rock consisted of an organic framework dominantly composed of hermatypic corals and an interstitial calcarenite. The dominant head corals are

Montastrea annularis, Diploria strigosa and Diploria labrynthiformis, and encrusting corals are Porites astreoides and Diploria clivosa. The framework was estimated at 30% of the rock volume with Montastrea annularis making up half of this (Stanley, 1966, p. 1929-30). Hoffmeister and Multer (1968, p. 1491), based on core study, reported that the coral framework is most prolific in the upper 30 m (100 ft) of the formation and that with depth calcarenite becomes more imporant. Furthermore, their cores indicated that the relative abundances of coral species is maintained in the subsurface, suggesting that there was little ecologic variation spatially or temporally.

No vertical or horizontal zonation of coral species was observed by Stanley (1966, p. 1932), and he concluded that the Key Largo Limestone represents a homogeneous community of frame-building organisms. A well sorted calcarenite sub-facies of the interstitial material was interpreted as inter-reef channel deposits. Typically a poorly sorted sub-facies composed of 40% micrite and 60% coarse bioclasts surrounds the coral heads. The general order of abundance is mollusc > coral > coralline algae > foraminifera, with 15% unidentifiable debris.

The relative abundance of bioclasts in the Key Largo Limestone compares reasonably well with Ginsburg's (1956) data for sediments of the present-day outer bank reef sub-environment. However the greater abundance of micrite in the Pleistocene rocks led Stanley (1966, p. 1934-35) to suggest that the Pleistocene sediments were representative of less turbulent conditions. This, plus the lack of zonation and apparent homogeneity of the Key Largo reef, indicates that no significant energy gradient existed.

The absence of Acropora palmata is significant because it is an extremely selective species requiring high energy conditions, and therefore, inhabits the surf zone down to a depth of approximately 4 m (13 ft) (Marszalek et al., 1977, p. 224).

ORIGIN OF THE KEY LARGO LIMESTONE

The controversy concerning the origin of the Key Largo Limestone centres around comparisons made between it and modern day analogues and the inconsistencies therein.

With the above as a basis, Stanley (1966, p. 1940) suggested that the Key Largo reef formed in relatively deep water at depths of 6 to 12 m (20-40 ft). The coral assemblage was ill-adapted to turbulent conditions (large heads, small bases, and high centre of gravity). The "Montastrea Zone" of the Key Largo reef was then analogous to the Montastrea- dominated assemblages on the windward lower levels of platform edge reefs that contain Acropora palmata. Faunal similarities revolved around the dominance of Montastrea annularis, the secondary importance of the three Diploria spp. and Porites astreoides, and the presence of branching species of Porites and Acropora cervicornis (Stanely, 1966, p. 1938).

The lack of zonation, absence of Acropora palmata, presence of large heads of Montastrea annularis, Diploria spp., and branching corals, and nature of the associated sediments in the Key Largo Limestone also make comparison of this formation to present-day patch reefs a viable alternative. Hoffmeister and Multer (1968, p. 1497) suggested that the Key Largo assemblage represents back-reef coalescent patch reef deposition in shallow water. As with Stanley's (1966) deep water Montastrea zone hypothesis, Hoffmeister and Multer's (1968) shallow water coalescent patch reef hypothesis reflects an environment of low wave energy and explains the absence of Acropora palmata in light of information from present-day reefs.

WATER DEPTH AND PRESENCE OF ACROPORA PALMATA

The Atlantic Coastal Ridge is a northeast-southwest trending cross-bedded oolite and reaches a maximum elevation of 7.6 m (25 ft) above sea level (Hoffmeister and Multer, 1968, p. 1495). The oolite is contemporaneous with the Key Largo reef (discussed in Chapter 2) and as such this elevation is presumed to denote the approximate elevation of sea level during maximum Q5 transgression (Hoffmeister and Multer, 1968, p. 1495). The difference in elevation between the ridge and the maximum elevation of 5.5 m of the Key Largo reef from Windley Key is approximately 2 m. This depth, or perhaps slightly more, is suggested for the reef at the peak of Q5 transgression. During at least the close of Q5 deposition, the reef was in water depth appropriate for the growth of Acropora palmata.

Further suggestion of a shallow water origin is provided by Perkins (1977, p. 179):

> "Red soil pockets and laminated crust marking the upper unconformable surface of the Q4 unit are exposed in places in the floor of the quarry located on Windley Key. The surface has an

> elevation of approximately 3 ft (1 m) above sea level. If we assume that the maximum elevation of the oolite bars marks the approximate position of sea level during maximum Q5 transgression, water depths over the Q4 erosional surface on Windley Key could never have exceeded 22 ft [6.7 m] assuming there was no deposition during the Q5 transgressive phase. Such a figure is unrealistic, for much, if not most, of the coral growth recorded at this locality likely occurred during the transgressive phase with corals keeping pace with a rising sea. Maximum water depth over the frontal edge of this transgressive coral-rich unit probably never exceeded 10 ft (3 m)."

The above implications for a shallow water reef argue strongly in favour of a patch reef origin, as the absence of *Acropora palmata* would be difficult to explain in a platform edge reef of such shallow depth (Hoffmeister and Multer, 1968, p. 1495).

PRESENCE OR ABSENCE OF PLATFORM EDGE REEF

If we accept the patch reef hypothesis, there are a number of points to further consider. By analogy, if these patch reefs lie back from the shallow platform edge, then it is at the platform edge that *Acropora palmata* should occur in reefs with the characteristic zonation.

Hoffmeister and Multer (1968, p. 1496) drilled several holes in the Florida platform and platform edge, and although more or less typical Key Largo sediments constituted the Q5 unit in most of the boreholes, core number 4 drilled slightly landward of the platform edge produced two specimens of *Acropora palmata*, one at the 18 m (58 ft) level and another at 20 m (65 ft). From this evidence it was concluded that there had been an outer bank reef at roughly this position.

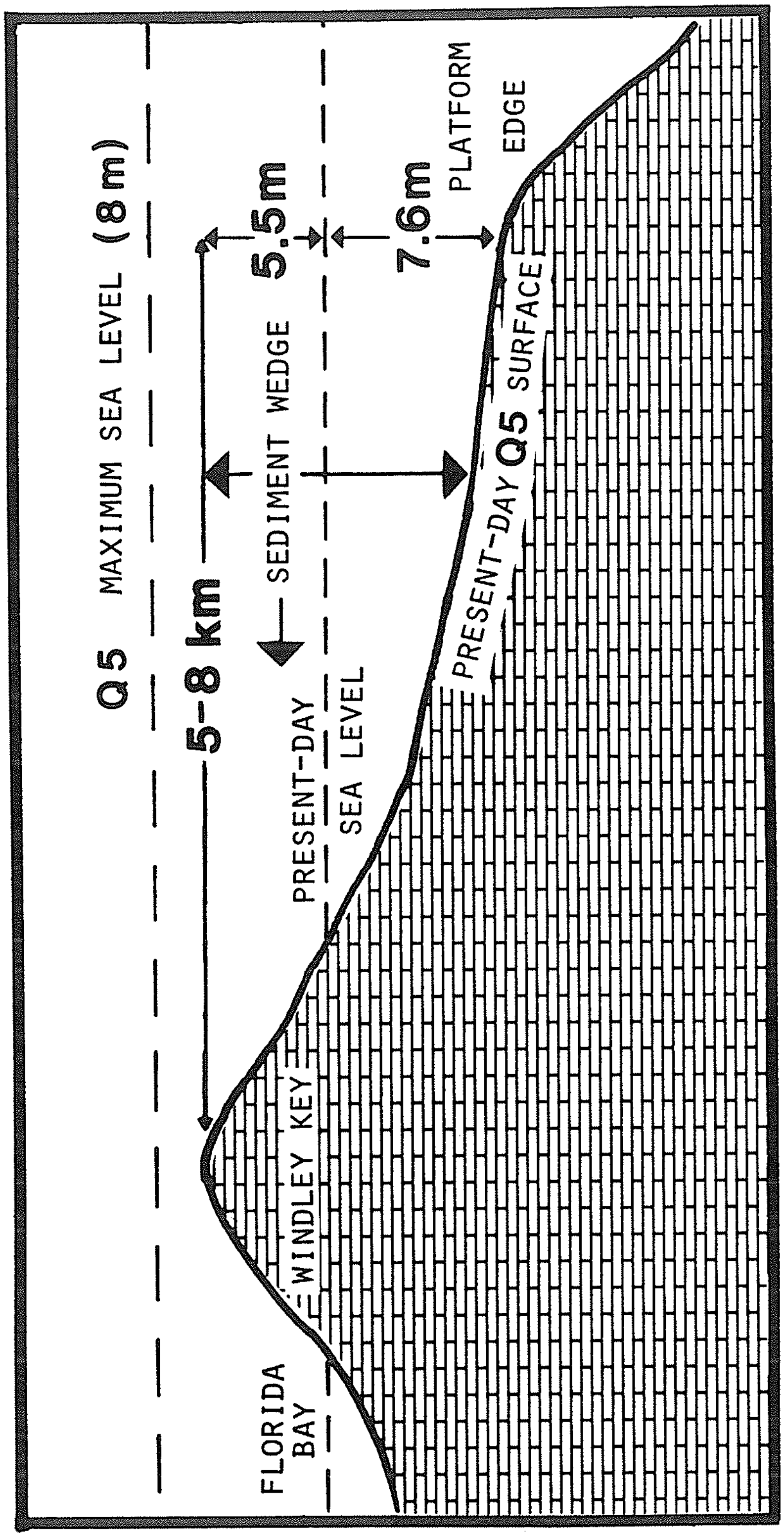
Although the existence of Acropora palmata in the previously hypothesized reefs at the shelf or platform edge position is established, the logical question to ask is whether there was indeed an Acropora palmata dominated surf-zone community analogous to present-day platform edge reefs. Resolution of the above requires more extensive study of the subsurface of the reef tract.

EROSION OF Q5 SEDIMENTS

A difficulty in accepting the coalescent patch reef origin (Hoffmeister and Multer, 1968) for the Key Largo Limestone is in establishing how the Q5 surface was eroded. Recall that the maximum Q5 Key Largo elevation is 5.5 m (18 ft) at Windley Key, within the trend of the hypothesized coalescent patch reef. Present-day platform edge reefs stand on a rocky (Q5) surface approximately 7.6 m (25 ft) below sea level (Stanley, 1966, p. 1939). If the Q5 platform edge reef and Q5 patch reefs were at approximately the same elevation with respect to sea level during growth, then a wedge of sediments several kilometres wide (from Keys to platform edge) and approximately 13 m (44 ft) or more in thickness would have had to be eroded to produce the present Q5 surface (Figure A-1). How and when this bevelling took place is open to question.

> "If such bevelling had been accomplished by wave action during the post glacial sea level rise, [erosion during transgression] one might expect the line of bevelling to run approximately parallel with the continental shelf. However, according to

Figure A-1. Schematic diagram to illustrate magnitude of erosion necessary to accept the coalescent patch reef hypothesis. A wedge of sediment greater than 13 m thick and the width of the reef tract must have been removed if the proposed platform edge reefs and coalescent patch reefs were built up to similar elevations. This wedge, however, had considerable internal relief (not shown), and consequently the quantity of sediment removed would have been significantly less than the diagram implies.


Q5 MAXIMUM SEA LEVEL (8 m)
5-8 km
5.5m
SEDIMENT WEDGE
PRESENT-DAY SEA LEVEL
7.6m
PLATFORM EDGE
PRESENT-DAY Q5 SURFACE
FLORIDA BAY
WINDLEY KEY

> recent estimates the last 25 ft [7.6 m] of sea level rise along the Atlantic coast has occurred only within the last 4000 - 6000 years ..., and it is impossible that the vast amount of bevelling required could have been accomplished during this brief time interval"
>
> (Stanley, 1966, p. 1939)

Such bevelling would also have to explain the arcuate relict coalescent patch reef forming the Keys landward but parallel to the platform edge. Stanley (1966, p. 1939) further stated that the broad central ridge of the Keys is the dominant part of an ancient reef arc and the orientation is precisely that to expect of such a reef.

Erosion of the platform is thought by Hoffmeister and Multer (1968, p. 1947) to be possible if it took place on unconsolidated Q5 sediments during the regressive phase of Q5 time. Because the seaward platform was in longer contact time with wave energy, the seaward dipping platform was thus created. Furthermore they suggested that Q5 depositional topography was similar to today's reef tract topography - that patch reefs and bank reefs may rise to the surface of the water, but water depths away from the reefs may be 8 to 12 m (25 to 40 ft) deep, thus considerably reducing the volume of sediments to be eroded in the wedge.

The orientation of the Keys was thought by Hoffmeister and Multer (1968) to be a product of growth rather than erosion and they cited that patch reefs orient themselves parallel to the platform edge, often forming on summits of elongate ridges of calcarenitic material. They also suggested the possibility of other similar patch reefs existing in Q5 time.

> "In all probability, ... one and possibly two other similar elongated series of parallel reef patches occupied the space between them [the reefs forming the Keys] and the bank reef..."
>
> (Hoffmeister and Multer, 1968, p. 1500)

Interestingly, the above inadvertently supports an erosional origin for the present shape and slope of the Q5 shelf.

Stanley (1966, p. 1939) justified his Montastrea zone's broad, low, homogeneous reef morphology by suggesting that it may have been related to growth on a wide, flat platform several kilometres behind the platform edge. Also, the absence of Acropora palmata possibly prevented the development of a surf-zone community and instead of growing upward, the reef spread laterally (Stanley, 1966, p. 1940). The platform edge was hypothesized to have been too deep to permit effective colonization by hermatypic corals. There is difficulty, however, in explaining Key Largo Limestone flooring virtually all of the carbonate shelf, as work by Hoffmeister and Multer (1968, p. 1496-1497) demonstrated. A possible explanation from the deep water advocates' point of view is that definite biohermal construction began at some time into the transgressive phase of Q5 time. Previously, corals may not have formed significant build-ups but instead migrated back from the shelf margin.

Perkins (1977, p. 181-183) suggested a type of patch reef origin for the Key Largo Limestone.

> "The key to this mechanism lies in an examination of a modern shelf-margin sand shoal, White Bank, which lies behind the line of living outer reefs in the Florida reef tract. This skeletal sand-shoal and patch-reef complex parallels the arcuate shelf break and the line of living outer bank reefs; it is asymmetrical in cross section, being shallowest on its leeward edge, usually 10 ft (3 m) or less. Extensive coral patches growing on this sand shoal and behind its sheltered leeward margin become enveloped in sand as this shoal migrates. Such a sand-shoal-patch-reef complex migrating landward during a

rising sea would produce a seaward sloping, wedgelike deposit of corals encased in skeletal sand. The leading, shallowest edge of such a migrating sand shoal would be exposed as a linear high during subsequent lowering of sea level, would undergo cementation, and would be preserved as an arcuate coral patch reef and skeletal sand complex. Such a sand-shoal-patch-reef interpretation is favored for the deposition of the Q4 and Q5 coralline facies of the Florida Keys. The migrating shoal may have been initiated at a paleotopographic break in slope on the outer shelf margin and does not require a seaward reef barrier for its inception."

PRESENT-DAY ANALOGUES - SIGNIFICANCE OF *ACROPORA PALMATA*

The complexities become greater. The type of reef described by Stanley (1966) has no modern analogue in the Florida area, while the living patch reefs are nowhere nearly so extensive or continuous as to be a proper analogue of Hoffmeister and Multer's (1968) coalescent patch reef (Dodd et al., 1973, p. 3995). The migrating sand shoal-patch reef complex proposed by Perkins (1977) also has a scale drawback. If the Keys do represent a platform edge reef (shallow water) and as *Acropora palmata* is known from Pleistocene rocks of the Caribbean, then perhaps it is the higher selectivity of this particular organism which accounts for its absence here. For example, the species is absent from the various reefs of the Bermuda platform, while the other prominent Key Largo species are ubiquitous there.

Perhaps some combination of salinity, temperature, and turbidity may have precluded the species from growing here during the Pleistocene. Looking at modern reefs, Ginsburg and Shinn (1964) suggested that the western margins of the Florida and Bahama platforms are unfavourable to the reef community because warmer and more saline waters are moved westward across the platforms by easterly winds. Along the same lines of reasoning, the reef community favored development leeward of islands where they are shielded from the tidal runoff of platform waters. This effect was also noted by Marszalek et al. (1977, p. 228) in their study of reef distribution in the Florida reef tract. They noted an inverse correlation between the existence of patch and platform edge reefs and the existence of tidal passes between the ocean and Florida Bay. The south Florida reefs are not uniformly distributed. Most reefs are located in the upper Keys (seaward of Elliot Key and Key Largo) and to a lesser extent between Big Pine Key and Dry Tortugas. The middle Keys with their numerous tidal channels separating them have only rare patch reefs. Platform edge reefs are occasionally developed.

Warm waters and high salinity are not the only factors which affect modern reef distribution. Marszalek et al. (1977, p. 228) cited evidence that demonstrates that reef growth is discouraged by mixing with cold waters. The large shallow expanse of Florida Bay allows cold fronts and high wind velocities to quickly lower its temperature to 15°C or less and thus these colder waters mix with reef tract waters and the latter is cooled. The average water temperature in the Florida reef tract is near the minimum that will support reef growth, and a slight cooling may drive *Acropora palmata* further south allowing the hardier species to remain behind (Stanley, 1966, p. 1937). According to Lighty et al. (1978, p. 60), pronounced temperature fluctuations and unusually cold bottom waters in the shallow waters off southeast Florida are preventing active reef growth north of Miami.

Salinity reduction via mixing with lower than normal salinity currents is also a factor which may preclude reef development or the existence of highly salinity specific organisms. The Loop Current, which connects the Yucutan and Florida Currents in the eastern Gulf of Mexico, may, at times of maximum development, extend as far north as the Mississippi delta and may reach the west Florida shelf. In these areas, low salinity coastal waters may be incorporated into the current as it flows southward to eventually reach the Keys (Marszalek et al., 1977, p. 228). This may account for the absence of bank reefs along the platform edge near Dry Tortugas.

The death of an early Holocene barrier (*Acropora palmata*) reef located along the shelf break off Miami coincided with the beginning of flooding of the Florida platform during the Holocene transgression approximately 7000 years BP (Lighty et al., 1978, p. 59). They suggested that cold waters (bottom temperatures along the eastern Florida shelf may reach 10°C during March, April, and May) and turbidity (originating from the south where soil and lagoonal deposits of fine lime mud were widespread) possibly led to the reef's demise.

Based on modern reefs it can be seen that any number of factors may have led to colonization of the Key Largo reef with *Acropora palmata* being absent, if assuming a platform edge reef origin. Work by Shinn et al. (1977) on Holocene reefs of the Florida reef tract has disclosed a number of platform edge reefs at Marker G and Dry Tortugas which host no

Acropora palmata, and they concluded (regarding the Key Largo reef) that "it is no longer necessary to call on a patch reef origin. Reefs can form and keep pace with sea level at the platform margin, for reasons not understood, without the help of *Acropora palmata*" (Shinn et al., 1977 p. 6).

Dodd et al. (1973) have described the Newfound Reef which lies between 0.8 km (0.5 mi) and 1.2 km (0.7 mi) off the southern shores of Newfound Harbor Keys and Big Pine Key. The reef is approximately 60 m (197 ft) wide and is located on the seaward edge of a bedrock terrace (Dodd et al., 1973, p. 3996). Water depths range from 4.5 m (15 ft) on the landward side to 7 m (23 ft) on the oceanward side. The biota, geometry and geography suggested to Dodd et al. (1973, p. 4000) that the reef may be the Holocene analogue to the Key Largo reef. Biota is similar with the possible exception of the presence of *Siderastrea* and *Solenastrea hyades* being common in the Newfound Reef. Sediment texture in and around the reef also compares favourably. "As sea level continues to rise, Newfound Reef will probably expand vertically and laterally, and gradually incorporate the present patch reef belt; such a composite reef tract would compare even more favourably with the Key Largo Limestone in scale and geometry" (Dodd et al., 1973, p. 4000).

The elevated sea levels during the Q5 transgression inundated large areas of coastal Florida, especially the southern tip of the peninsula (Figure 2-9, text). Effects associated with such a large platform (several times the size of Florida Bay) are difficult to speculate on but Harrison (pers. comm., 1981) suggested that the Great Bahama Bank may be used as a possible model where salinities above normal sea salinity develop and persist (see Bathurst, 1976, p. 276-292).

OOID SHOALS OF THE MIAMI LIMESTONE

The facts that the Q5 oolite and reef are contemporaneous (Osmond et al., 1965; Broecker and Thurber, 1965) and that the oolite is crossbedded (Hoffmeister et al., 1967, p. 178) suggest very strongly that a shallow water platform edge reef origin for the Key Largo reef is reasonable. A deep water *Montastrea* zone reef that Stanley (1966) proposed does not explain the obvious cross-bedding in the oolite.

However, it is interesting to note that the proximity of oolitic and reefal deposits in the Pleistocene of south Florida is anomalous based on relationships observed on the present-day Bahama Bank where oolitic sands and reefs are mutually exclusive (Ginsburg and Shinn, 1964). Although Stanley (1966, p. 1939) admitted that the best developed oolite of the Miami Limestone is northeast and southwest of the main Key Largo exposures, the presence of oolite platformward of the reef is used to support his deepwater *Montastrea* zone hypothesis. This situation would allow sufficient wave energy to by-pass the reef and form ooids. The question to ask is, why is the best-developed oolite flanking the reef?

The patch reef hypotheses of Hoffmeister and Multer (1968) and Perkins (1977) also have difficulty in explaining the high energy oolitic deposits that are farther bankward. Perkins' (1977) sand shoal-patch reef hypothesis neatly explains the Key Largo Limestone underlying the present reef tract but does not attempt to elucidate on the reef-oolite relationship. The proposed migrating sand shoal-patch reef complex must have terminated for some reason at the present Florida Keys. Could a paleotopographic slope change at this point have halted the migrating complex and caused a platform edge reef to subsequently keep pace with a rising sea level by essentially building upwards *in situ* with the accompanying development of the ooid shoal?

The lack of emergent Key Largo reef seaward of the lower Florida Keys may reflect termination of reef growth in this area as a result of the formation of incipient shoals. Further shoal growth effectively precluded any more reefal development in this area. These incipient shoals may have been prompted by large passes in the early Key Largo reef or a coral community, which for some reason, did not keep up with its contemporaries to the east, thus allowing wave energy to continue largely unimpeded over the reef and initiate ooid development.

SUMMARY AND CONCLUSIONS

In summary, previous workers have proposed a number of possible origins for the Key Largo reef. These include a deep water *Montastrea* zone hypothesis (Stanley, 1966); a shallow water coalescent patch reef hypothesis (Hoffmeister and Multer, 1968); and a migrating sand shoal-patch reef hypothesis (Perkins, 1977). Recently it has been demonstrated that reefs of the Florida reef tract are not necessarily dependant on

Acropora palmata to keep pace with rising sea level (Shinn et al., 1977), and a platform edge reef origin is enhanced. Studies on the distribution of modern reefs in relation to physical and chemical factors in their environments have shown that corals may be quite selective as to where colonization takes place (Ginsburg and Shinn, 1964; Marszalek et al., 1977). The ooid shoals adjacent to the reef complex further suggest a platform edge origin for the Key Largo reef. When we stop to consider the host of possible effects related to the inundation of a large featureless land mass producing a large, shallow platform, perhaps the absence of a highly selective species such as Acropora palmata is not so surprising (Harrison, pers. comm., 1981).

APPENDIX B - BIG PINE KEY CORE SUMMARY

The following text describes briefly the prominent macroscopic and microscopic compositional and textural attributes of the cores from Big Pine Key, shown in the visual core summary diagrams that follow. Depth values on the diagrams are metres below land surface.

1. Rock Colour: White (10 YR 8/1, 10 YR 8/2 - Munsell Colour) (Excluding subaerial surface features - see text for details).

2. Grain Size: Well sorted grainstone typically from medium to coarse sand. As sorting decreases, grain size may span silt to granule size.

3. Borehole Notes:

BP-1
- a. Vug and channel porosity well developed throughout unit.
- b. *Strombus*.
- c. Unit is well cemented with intense vug and channel porosity developed. Foraminifera and articulated coralline algae *Amphiroa* common.
- d. Subaerial surface features extend to this depth or possible minor subaerial exposure surface.

BP-2
- a. Vug and channel porosity is moderately developed in upper metre and becomes more intense with depth.
- b. Local areas of micrite cement zone extend to this depth.
- c. Unit is very well cemented with intense development of vug and channel porosity. Foraminifera common.

BP-3
- a. Upper metre of unit has moderate development of vug and channel porosity. Below this, porosity is well developed.
- b. *Spondylus*.
- c. Interval between 3 m and the Q4 subaerial surface is extremely crumbly due to intense development of vug and channel porosity.
- d. Vug and channel porosity intensely developed. Cementation same as in Q5 unit.

BP-4
- a. Vug and channel porosity is well developed throughout unit.
- b. *Strombus*

BP-5
- a. Throughout Q5 unit, vug and channel porosity is moderately developed.

b. Below this point, core scrambled. Position of Q4 surface determined by assuming equal recovery throughout interval from 3.1 to 6.1 m.
c. Unit is well cemented with development of vug and channel porosity. Foraminifera and silt size bioclastic debris common.

BP-6
a. Development of vug and channel porosity is minor until 1.5 m depth. From 1.5 m to bottom of unit development is moderate to intense.
b. Local domains of micrite cement zone extend to this depth.
c. Unit is well cemented with intense development of vug and channel porosity. Foraminifera and silt size bioclastic debris common.

BP-7
a. Upper 1 m of core has only minor development of vug and channel porosity. Development is moderate elsewhere. Cross-bedding exerts strong influence on orientation of these pores.
b. High density of worm (?) tubes. Possible subaerial micrite cement.
c. Unit is well cemented with intense development of vug and channel porosity.

MO-2
a. Well-developed vug and channel porosity exists throughout entire Q5 unit.
b. Local areas of micrite cement zone extend to this depth.
c. Bedding defined by thin (few millimetres), roughly horizontal, alternating medium and coarse grained laminae. These exert an influence on orientation of vug and channel pores.
d. Unit is well cemented but crumbly due to intense development of vug and channel porosity. Silt size bioclastic debris common.
e. Large (several centimetres) dark grey, dense, foraminifera-rich, grainstone clast.
f. Unit is well cemented with intense development of vug and channel porosity. Foraminifera common in upper unit. Encrusting coralline algae occur locally.

MO-6
a. Upper 1.5 m of core has only minor development of vug and channel porosity.
b. Packstone interval characterized by crumbly core due to intense development of vug and channel porosity.
c. Below this depth, development of vug and channel porosity is not pronounced.
d. Well cemented interval (several centimetres) with vug and channel pores stained medium grey.

e. Unit is well cemented but with profuse development of vug and channel porosity. Fragmented foraminifera are abundant throughout unit.
f. Abundant silt size foraminiferal debris at bottom of unit.
g. Rock is very well cemented, vug and channel pores well developed throughout unit. Foraminifera locally common. Silt size bioclastic debris common.
h. Minor subaerial surface.

CORE SUMMARY: LEGEND

GRAINS

- OOID
- PELLET
- PELOID
- INTRACLAST
- F SKELETAL (<1 CM) : DOMINANTLY *HALIMEDA* AND MOLLUSC FRAGMENTS. OTHER FOSSILS SUBORDINATE EXCEPT WHERE NOTED.

FOSSILS >1 CM

- GASTROPOD
- PELECYPOD
- BRYOZOA (*SCHIZOPORELLA FLORIDANA*)
- BRANCHING CORAL (*PORITES* SP.)
- HEAD CORAL (*MONTASTREA ANNULARIS* EXCEPT WHERE NOTED)
- (E) ECHINODERM
- ? UNIDENTIFIED SKELETAL

STRUCTURES

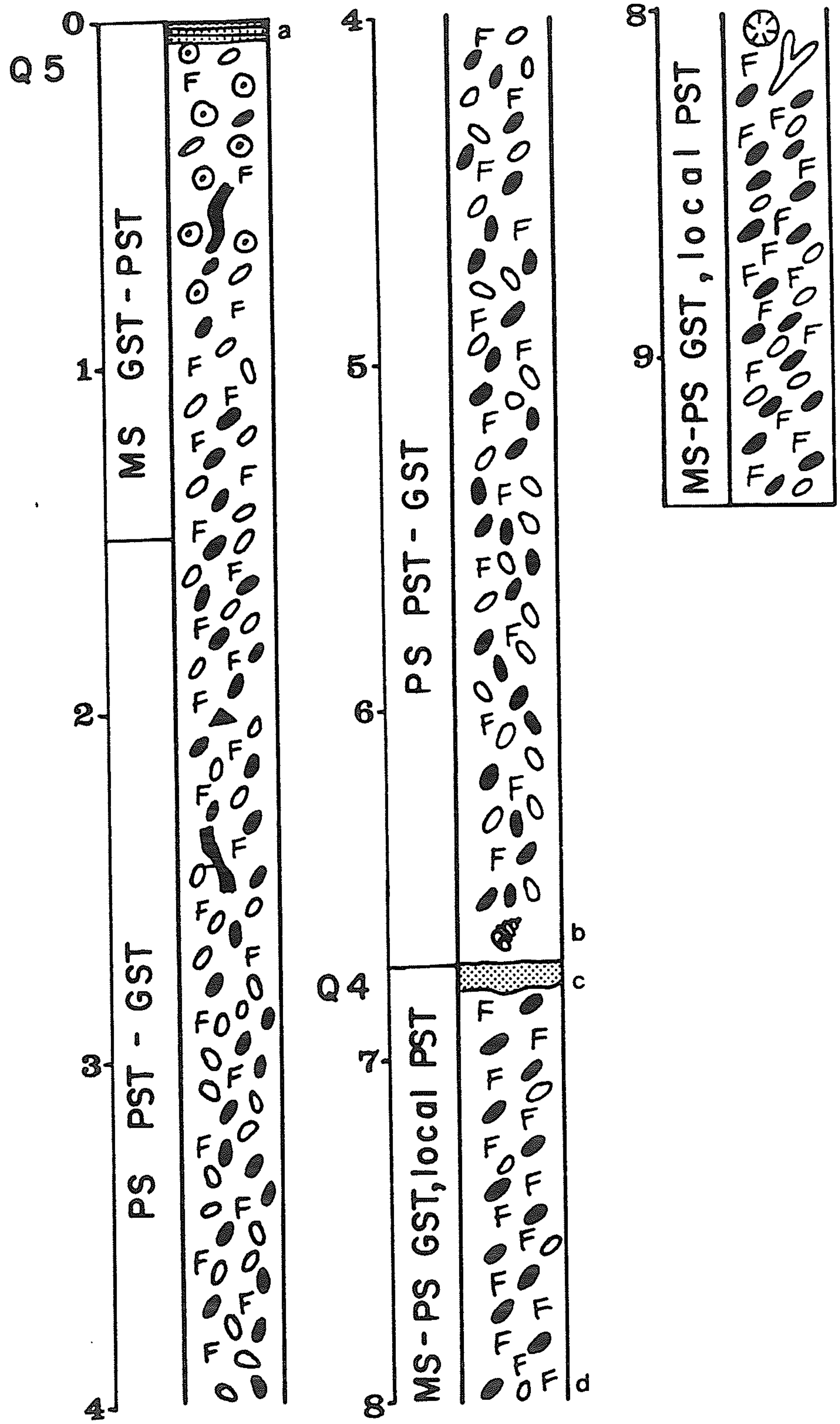
- BURROW
- CROSS BEDDING
- HORIZONTAL BEDDING
- DISTURBED BEDDING

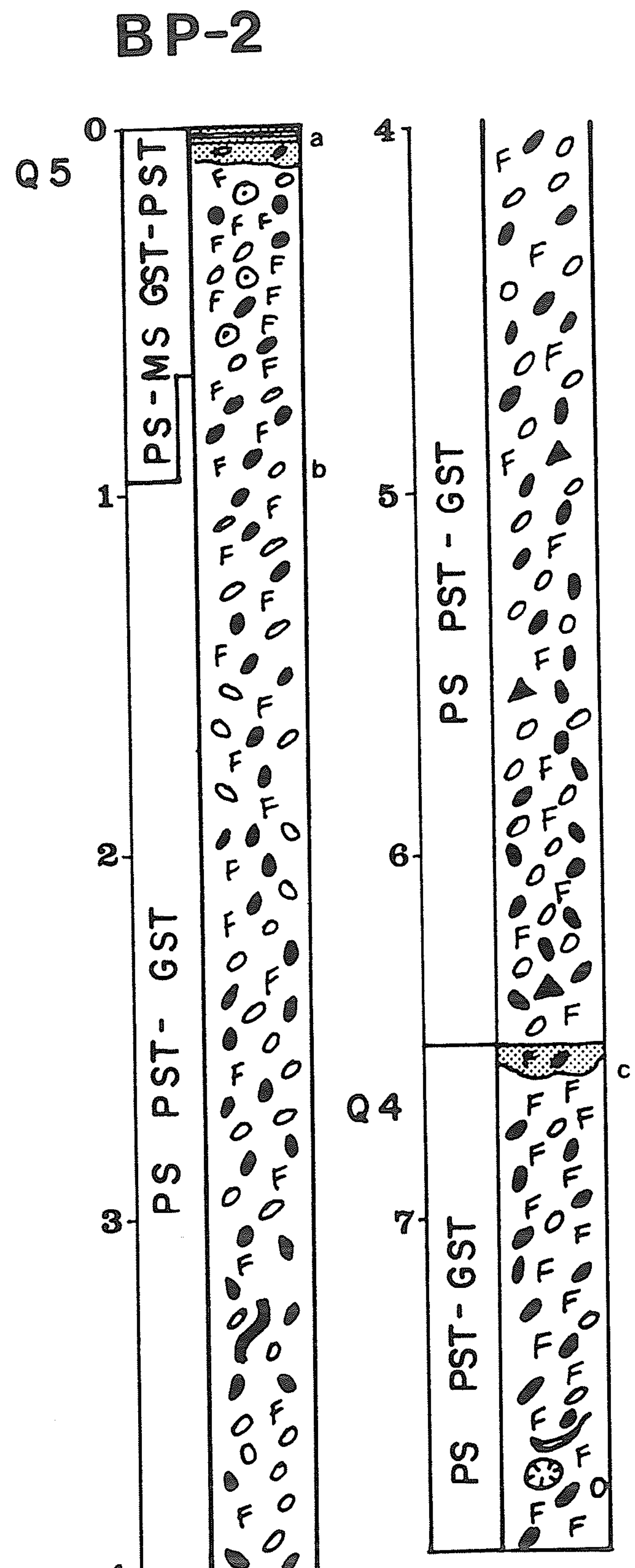
SUBAERIAL SURFACE

- LAMINATED CRUST
- MICRITE CEMENT ZONE
- MASSIVE, FEATURELESS CRUST

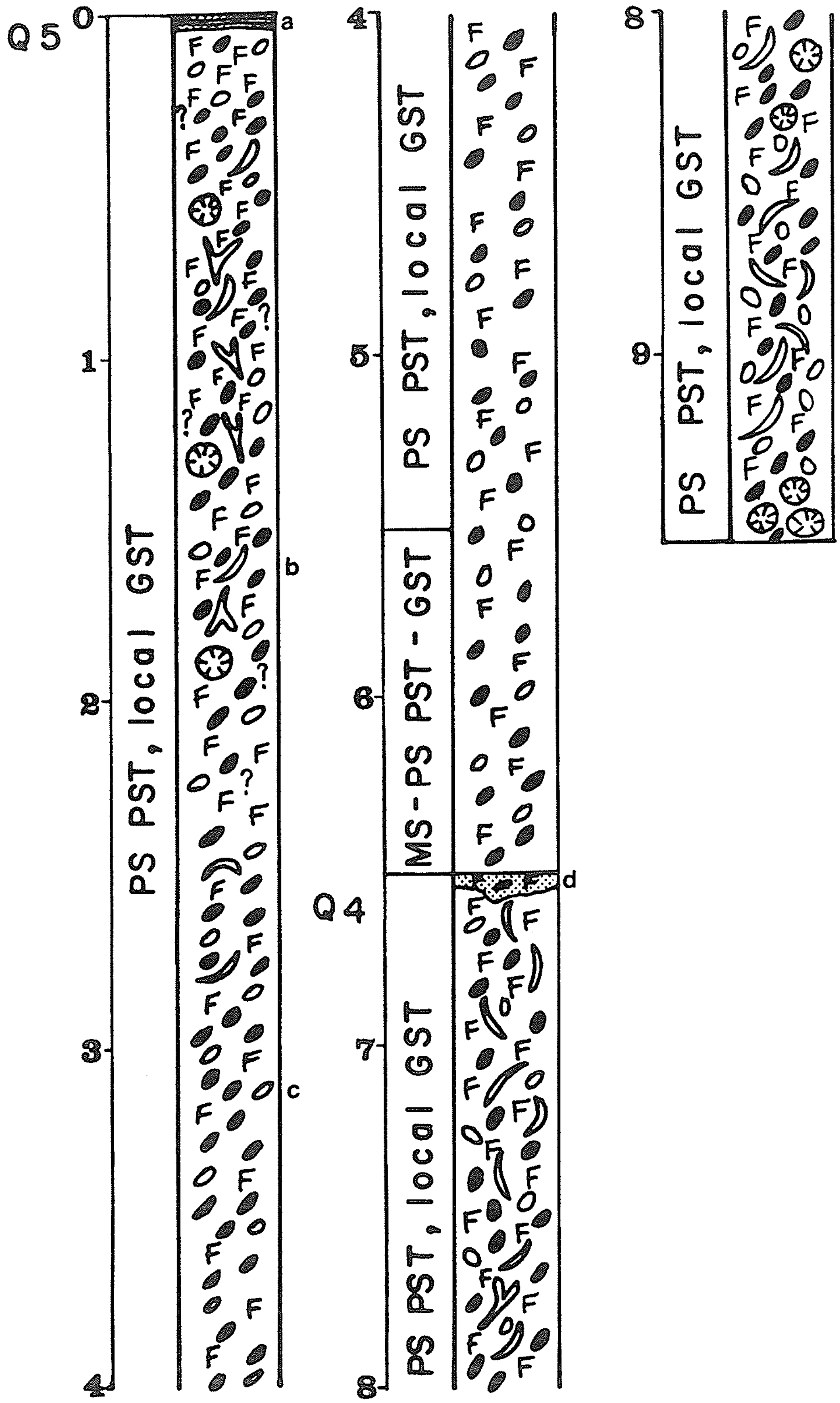
ABBREVIATIONS : GST - GRAINSTONE, PST - PACKSTONE, WST - WACKESTONE, WS - WELL SORTED, MS - MODERATELY SORTED, PS - POORLY SORTED

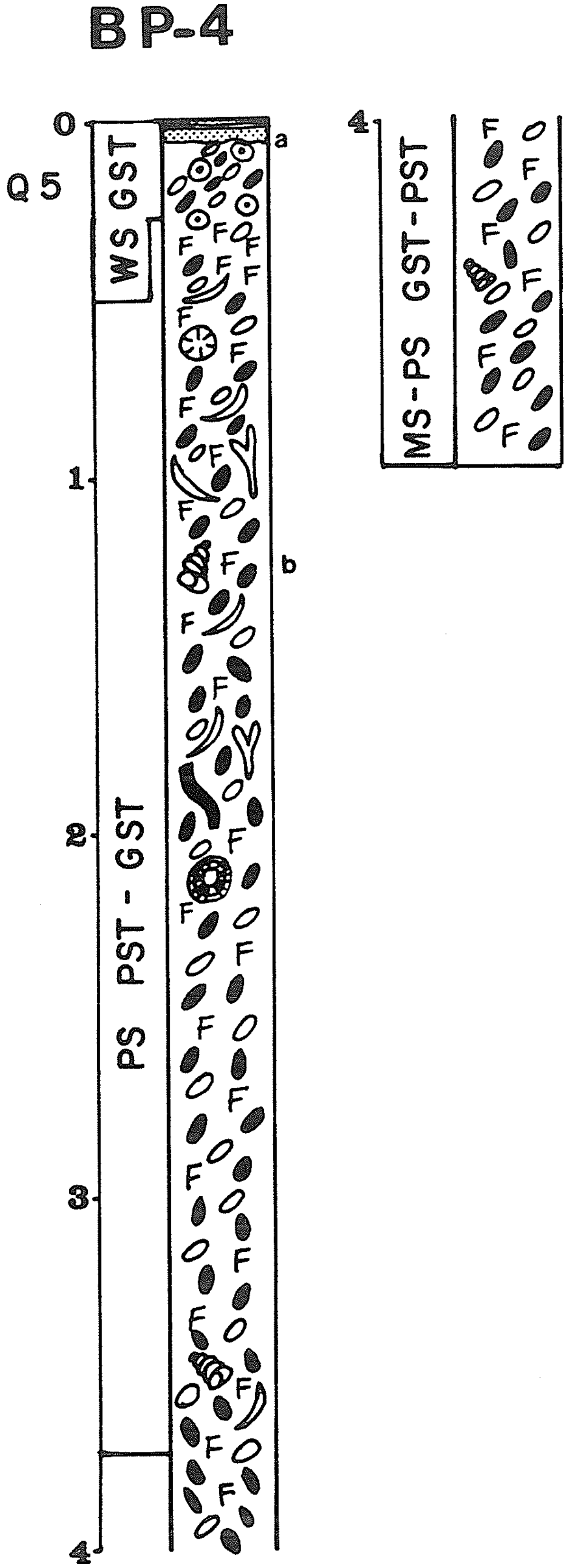
BP-1

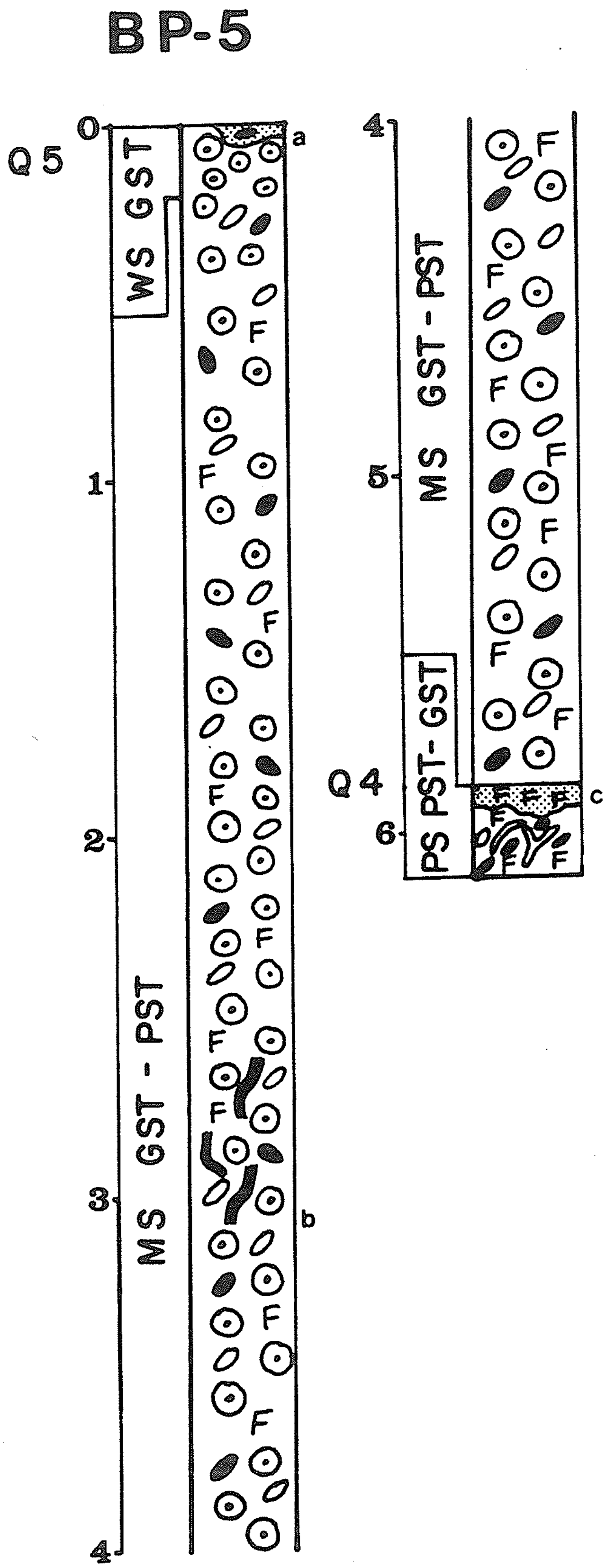

0
Q 5
a
MS GST - PST
1
2
3
4
PS PST - GST
4
5
6
7
8
PS PST - GST
b
Q4
c
MS - PS GST, local PST
d
8
9
MS-PS GST, local PST
F

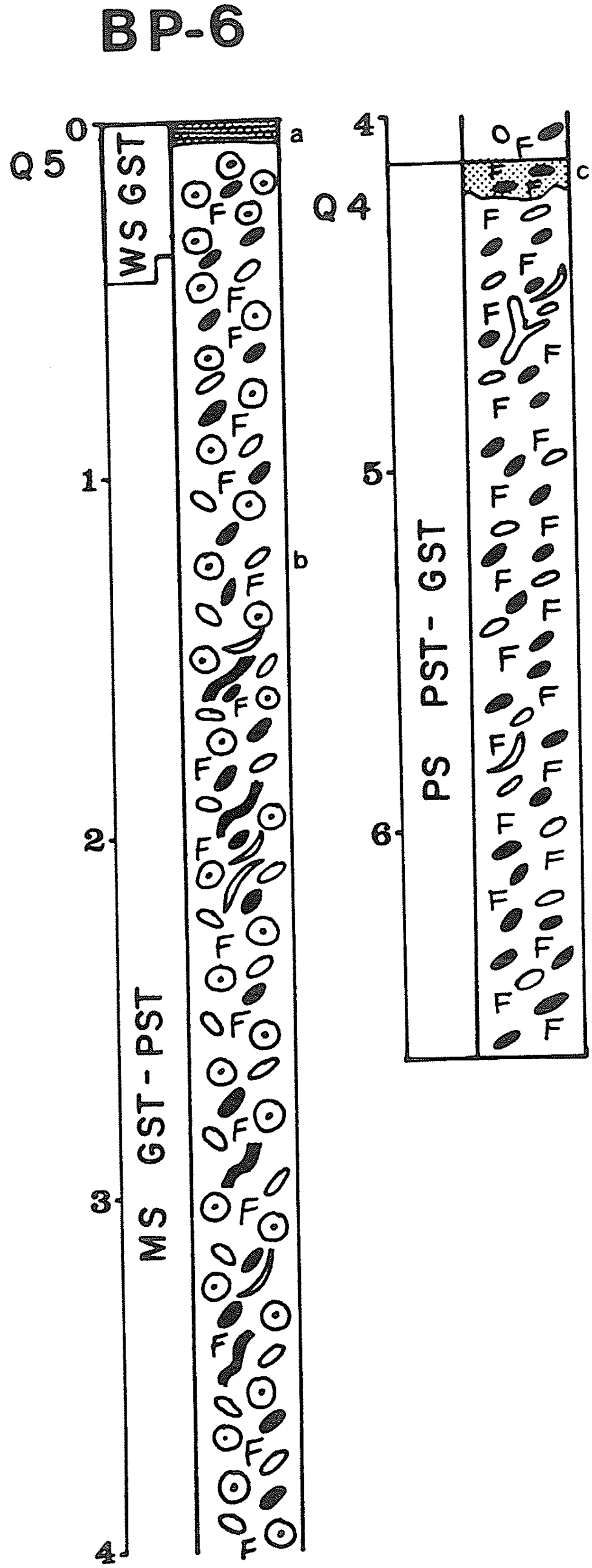

BP-2
0
Q5
PS-MS GST-PST
a
b
1
2
PS PST- GST
3
4
4
5
PS PST - GST
6
c
Q4
7
PS PST - GST

BP-3

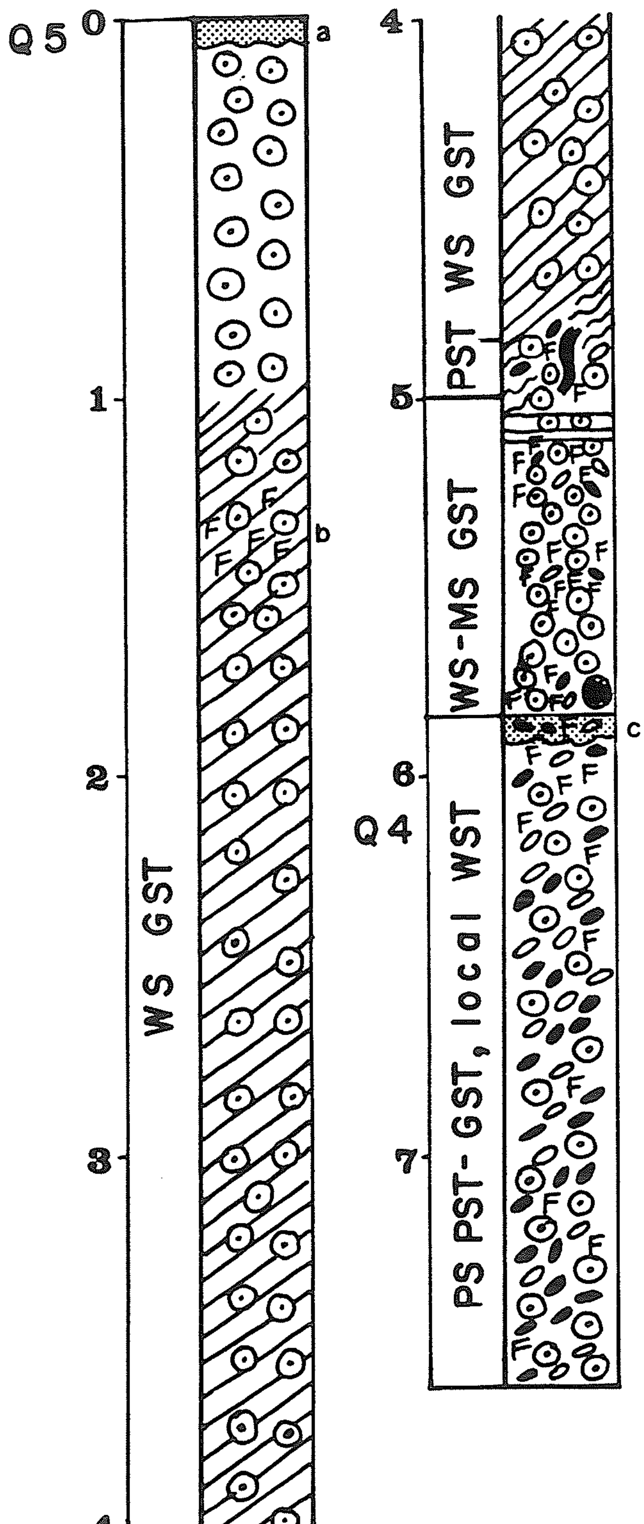

Q5
0
1
2
3
4
PS PST, local GST
a
b
c
4
5
6
7
8
PS PST, local GST
MS-PS PST-GST
Q4
d
PS PST, local GST
8
9
PS PST, local GST
F
O
?

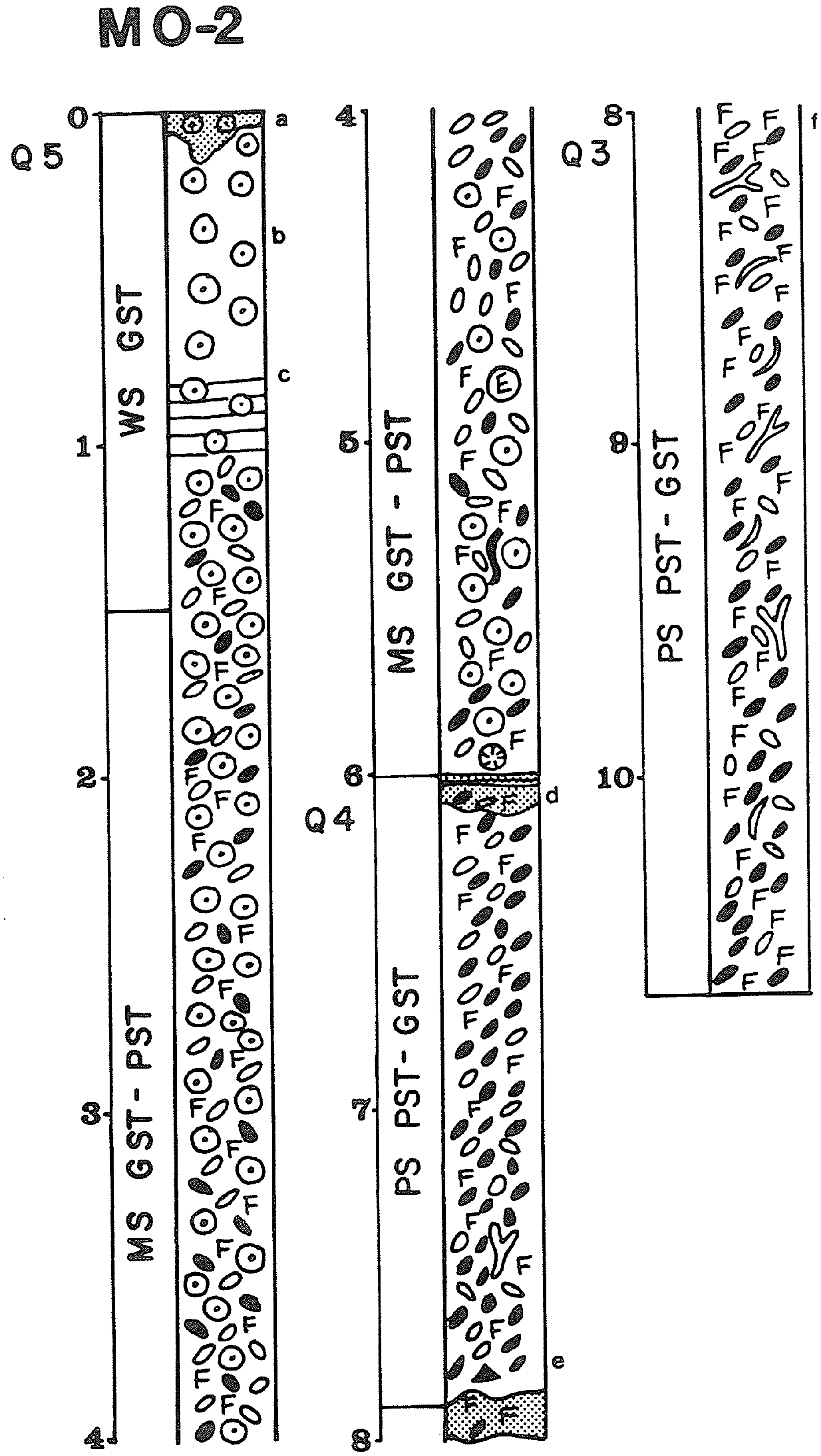

BP-4
0
1
2
3
4
Q 5
WS GST
PS PST - GST
a
b
4
MS-PS GST-PST
F

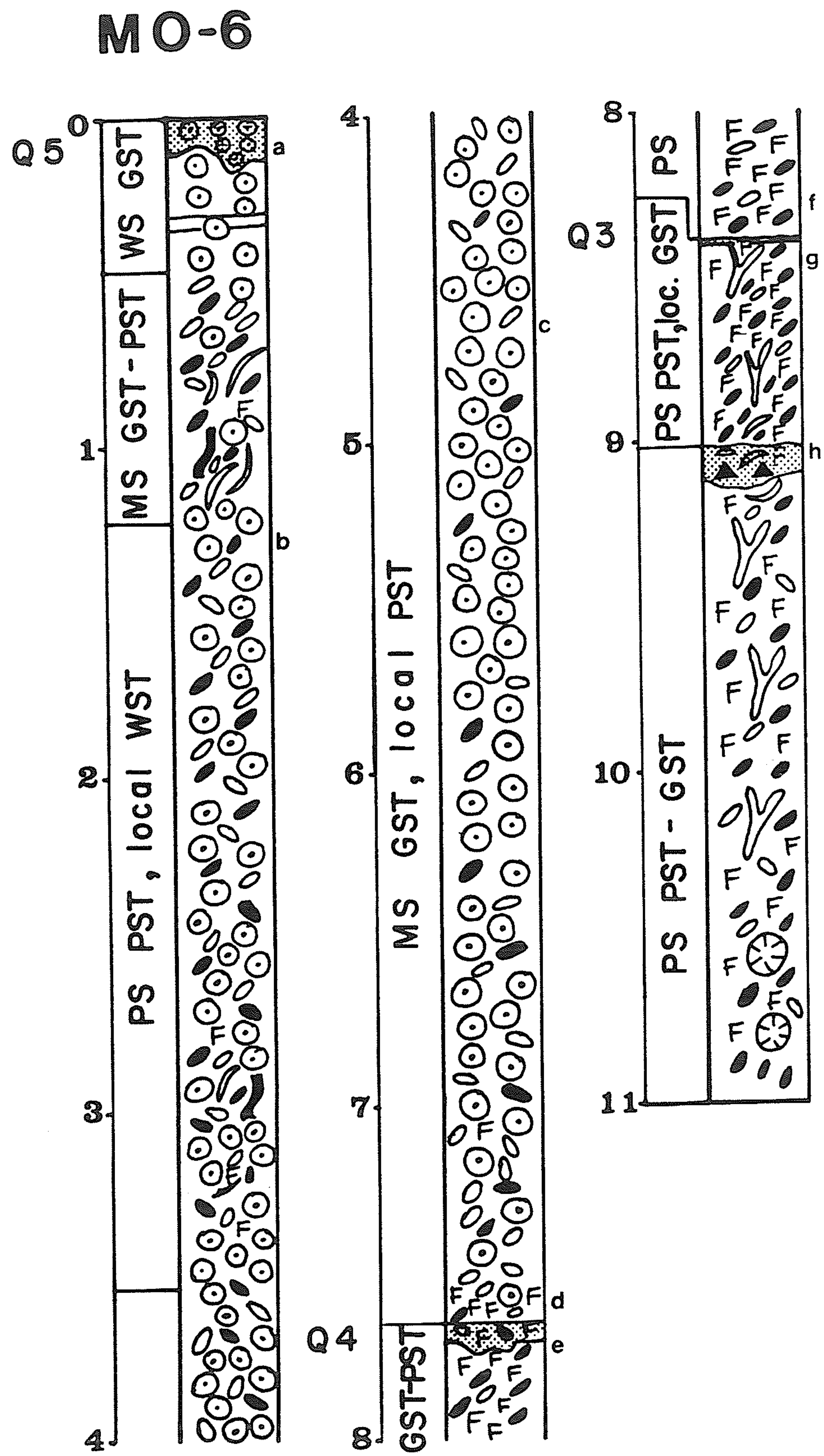

BP-5
0
Q 5
WS GST
a
1
2
3
b
4
MS GST - PST
F
4
5
6
Q 4
MS GST - PST
PS PST- GST
c
F


BP-6
Q5
Q4
WS GST
MS GST - PST
PS PST - GST
0
1
2
3
4
5
6
a
b
c
F

BP-7


Q5
0
a
1
b
2
WS GST
3
4
4
PST WS GST
5
WS-MS GST
c
6
Q4
PS PST- GST, local WST
7


MO-2
Q5
WS GST
MS GST - PST
MS GST - PST
Q4
PS PST - GST
Q3
PS PST - GST
a
b
c
d
e
f
E
F
0
1
2
3
4
5
6
7
8
9
10



MO-6
Q5
Q4
Q3
0
1
2
3
4
5
6
7
8
9
10
11
WS GST
MS GST-PST
PS PST, local WST
MS GST, local PST
GST-PST
PS
PS PST,loc. GST
PS PST - GST
a
b
c
d
e
f
g
h

APPENDIX C - POINT COUNTING

The locations of core thin sections examined in this study are presented in Table A-1. Twenty-one thin sections (Table A-2) were point counted using a 0.5 mm grid and counting 300 to 400 grains per thin section (methodology outlined in Hutchinson, 1974) in order to provide a quantitative basis for subdivision of the Q5 unit into three facies. An additional 24 thin sections were point counted in order to quantitatively compare primary and secondary 'matrix' porosity within and among the Q units (Table 8-1, text). Thin sections counted were judged to be representative of the interval from which they were taken.

Superficial and multiple ooids were counted as ooids. Pellets and peloids were differentiated on the basis of criteria outlined in the text (Chapter 3). Skeletal material included all bioclasts (unaltered and recrystallized). Spar, micrite, and microspar were lumped together based on general intimacy of micrite and microspar and the complete gradation of microspar to spar. Porosity types were differentiated as outlined in the text (Chapter 8).

TABLE A-1. THIN SECTION LOCATIONS

Borehole	Thin Section	Depth from Land Surface (metres)
BP-1	A	0
	B	0.6
	7	1.3
	C	2
	D	2.4
	E	3.3
	F	4.5
	G	4.8
	H	6.2
	J	7.3
	K	7.8
	L	8.2
	M	9.1
BP-2	A	0
	B	0.1
	C	0.1
	D	0.5
	E	0.9
	F	1.3
	G	2.2
	H	3.1
	I	4.5
	J	4.9
	K	5.5
	L	6.3
	M	6.3
	N	7.5
	73	7.6
	O	7.8
BP-3	A	0
	B	0.1
	C	0.4
	D	1.1
	E	2.2
	F	5.4
	G	6.6
	H	6.7
	I	6.4
	J	7.1
	K	8.2
	L	8.3
	M	9.3
BP-4	A	0
	B	0.2

	C	0.4
	D	1.7
	E	2.3
	F	2.7
	G	3.1
	H	3.5
	I	3.8
	J	4.2
	K	4.6
BP-5	A	0
	B	0.2
	C	0.9
	D	1.6
	E	2.2
	F	3.0
BP-6	A	0
	B	0
	C	0.3
	D	0.5
	E	1.3
	F	2.2
	G	2.5
	H	2.8
	I	3.6
	J	3.9
	K	4.1
	L	4.4
	M	5.9
BP-7	A	0
	B	0.1
	C	0.8
	D	1.2
	22	1.4
	E	1.5
	F	1.9
	G	2.8
	L	4.9
	K	5.1
	H	5.8
	I	6.1
	J	7.3
MO-2	A	0
	B	0.1
	C	1.3
	D	2.3
	E	2.8

	F	3.7
	G	5.3
	I	6.1
	J	6.8
	K	7.2
	L	7.7
	M	7.9
	N	8.0
	O	10.2
	P	10.6
MO-6	A	0
	B	0
	C	0.3
	D	0.6
	E	1.1
	F	1.9
	G	2.7
	H	3.0
	I	3.2
	J	4.1
	K	4.6
	L	7.1
	M	7.4
	N	7.4
	O	7.6
	T	7.6
	P	7.9
	79	8.3
	Q	8.4
	U	9.0
	V	9.1
	R	9.2
	S	10.9

TABLE A-2. POINT COUNT DATA
PERCENTAGE OF GRAINS, MATRIX, AND POROSITY IN THIN SECTION - Q5 UNIT

Thin Section	Points Counted	Ooid	Pellet	Peloid	Spar, Micrite, Microspar	Skeletal	Primary Porosity	Secondary Porosity	Total Porosity
BP-3 -D	343	0		56.9*		18.1	2.3	22.7	25.0
-E	324	0		62.0*		17.9	0	20.1	20.1
-F	301	0		54.5*		13.0	0	32.6	32.6
AVERAGE		0		57.8*		16.3	0.8	25.1	25.9
BP-5 -B	351	52.1	3.1	3.4	25.9	2.0	9.4	4.0	13.4
-C	349	31.5	7.2	9.7	41.5	2.6	3.4	4.0	7.4
-E	319	29.5	6.3	10.0	36.4	8.8	2.5	6.6	9.1
-F	314	45.5	11.1	4.5	29.0	5.7	2.2	1.9	4.1
AVERAGE		39.7	6.9	6.9	33.2	4.8	4.4	4.1	8.5
BP-6 -I	301	20.3	15.3	8.3	37.9	6.0	2.7	9.6	12.3
BP-7 -C	370	45.4	3.2	3.8	27.8	1.1	9.2	9.5	18.7
-D	304	41.4	1.0	1.6	31.0	6.9	8.6	9.5	18.1
-E	331	46.8	1.8	2.4	29.3	1.5	10.3	7.9	18.1
-F	365	52.9	0.5	3.6	30.1	0	10.4	2.5	12.9
-G	341	50.7	2.6	0.9	36.7	0.3	8.2	0.6	8.8
AVERAGE		47.4	1.8	2.5	31.0	2.0	9.3	6.0	15.3
MO-2 -D	279	36.4	4.9	3.8	40.2	3.5	6.2	5.2	11.4
-F	336	24.1	12.5	12.5	39.0	8.9	1.2	1.8	3.0
-G	345	13.9	6.1	11.6	37.7	21.2	3.2	6.4	9.6
AVERAGE		24.8	7.8	9.3	39.0	11.2	3.5	4.5	8.0
MO-6 -D	318	26.4	8.8	20.1	32.1	2.8	1.9	7.9	9.7
-E	247	26.3	16.6	9.3	34.4	0.8	4.0	8.5	12.6
-H	315	18.4	9.2	7.9	49.8	5.7	1.0	7.9	8.9
-K	350	46.0	2.9	5.7	25.1	8.6	3.7	8.0	11.7
-N	267	12.7	5.2	6.0	39.3	31.1	0.7	4.9	5.6
AVERAGE		26.0	8.5	9.8	36.1	9.8	2.3	7.4	9.7

* Combined value for undifferentiated pellets, peloids, micrite, microspar, and spar.

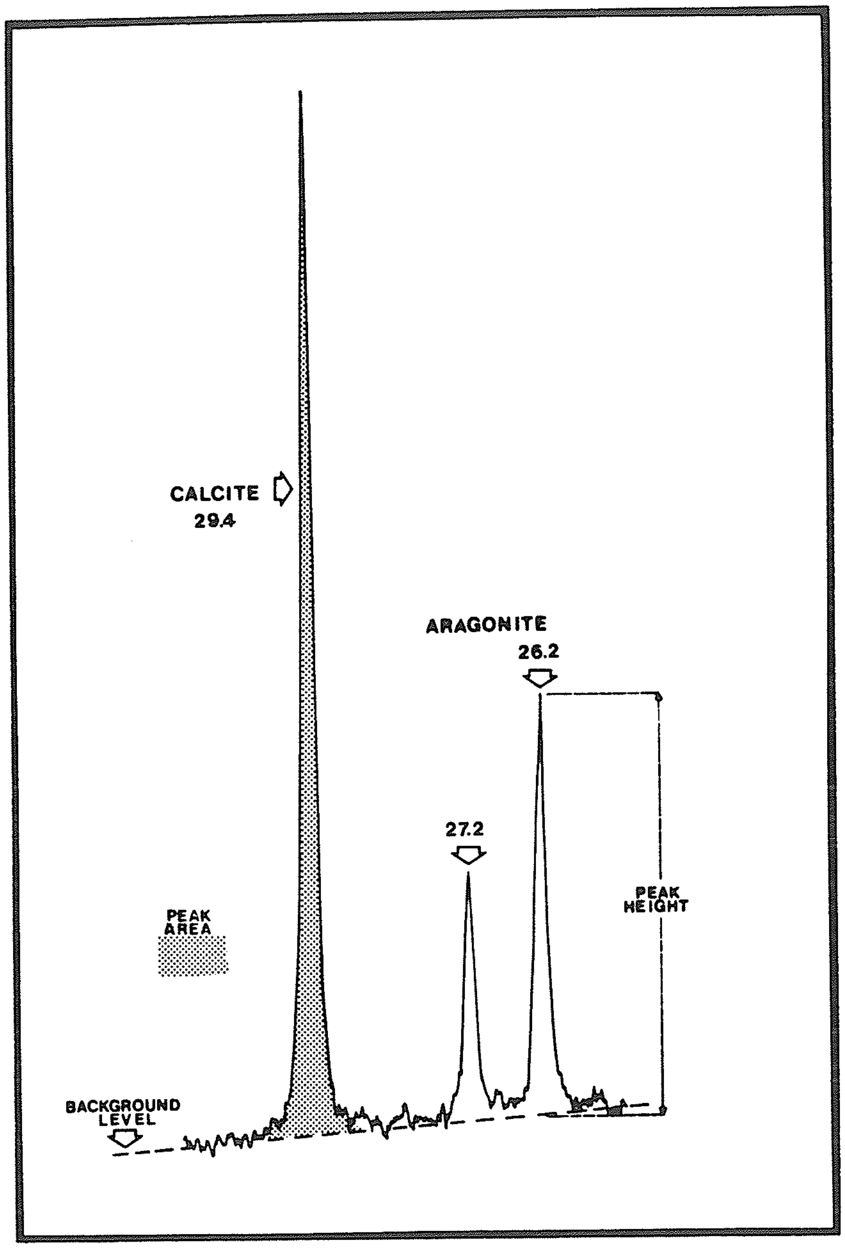
APPENDIX D - X-RAY ANALYSIS

Qualitative mineralogical evaluation using Fiegl's solution was supplemented by quantitative analysis via X-ray diffraction. The methodology is based largely on that outlined by Matthews (1965), but with some modifications. By comparing peak intensities of aragonite and calcite, an estimate of their relative weight percent in a given sample may be obtained. As the mass absorption coefficients of aragonite and calcite (no Mg) are equal, intensity is directly related to weight percent (Zussman, 1967, p. 229).

The intensity of the diffractogram peaks may be obtained in a number of ways (outlined by Milliman, 1974, p. 23). The measurement of intensity was carried out by planimetric measurement of X-ray diffraction peak area. Neuman (1965, p. 995) measured only the low 2θ half of the calcite curves then doubled the result to get the peak area and in doing so avoided possible incorporation of diffraction intensities from phases other than calcite. As no traces of Mg-calcite were found in the rocks of Big Pine Key, the peaks were measured as shown in Figure A-2. A common background line was drawn through the base of the peaks.

Powder slides were made similar to the method outlined in Matthews (1965): a small (1-2 g) sample representative of the lithology was ground in acetone with a mortar and pestle until grain size was such that it passed through a 230 mesh seive. The powder was placed on a glass slide and sufficient acetone was added to form a workable mixture which

Figure A-2. Parameters used to reduce X-ray diffraction data.


CALCITE
29.4
ARAGONITE
26.2
27.2
PEAK
HEIGHT
PEAK
AREA
BACKGROUND
LEVEL

was spread over the centre area of the slide.

X-ray equipment consisted of a Philips (PW 2113/00) x-ray diffractomoter using Cu Kα radiation, and Ni filter, operated at 50 kv and 20 ma. Each slide was scanned twice from 2θ of 25° to 31°. The slide was reversed between runs. A scanning speed of 1/2° 2θ/min. was used.

The areas under the following peaks were measured - aragonite 26.2° (111), aragonite 27.2° (021), and calcite 29.4° (104). The aragonite (111)/calcite and aragonite (021)/calcite ratios were calculated, the two ratios being independent estimates of the weight percent aragonite in the sample. According to Chave (1954), the use of these two values allows the possible detection of preferred orientations in the mixture on the slide.

The ratios were then converted to estimates of weight percent aragonite using curves published by Chave (1954) and Lowenstam (1954). The two runs provided four estimates of aragonite content. If the estimates were too dispersed, subsequent analyses were carried out. The estimates were then averaged.

Because the planimeter measurements were extremely time consuming (each calcite curve measured three times, the aragonite curves each measured 5-10 times), a best fit curve through a plot of peak height (Figure A-2) versus peak area was constructed once sufficient areas were measured. The best fit curve was based on 136 spread-out data pairs. Thus, peak areas were estimated via straightforward and quick measurement of peak height and the data treated as previously outlined.

Although there are many factors which may give rise to errors (see

Zussman, 1967, p. 300; Milliman, 1974, p. 25-27; Matthews, 1965, p. 32-33), the probable error is estimated at 5 to 10 percent for this method (Lowenstam, 1954, p. 287; Neumann, 1965, p. 993).

TABLE A-3. X-RAY DATA, BOREHOLE MO-6

SAMPLE NO.	PEAK AREA ARAG. 26.2 / CAL. 29.4	WT. % ARAG.	PEAK AREA ARAG. 27.2 / CAL. 29.4	WT. % ARAG.	AV. WT. % ARAG.
[Q5] 2	0.410 0.460	10 12	0.020 0.017	13 10	11
3	0.235 0.257	60 62	0.130 0.124	54 53	57
9	0.195	54	0.101	46	50
13 (E)	0.195 0.180	54 51	0.098 0.103	46 47	50
15	0.114	34	0.065	34	34
17	0.102	30	0.056	31	31
22	0.055 0.095	15 28	0.028 0.039	18 23	21
32	0.090 0.084	25 24	0.052 0.037	29 22	25
37 (I)	0.067 0.083	19 24	0.042 0.051	25 29	24
43	0.224 0.209	59 57	0.120 0.108	52 49	54
46 (J)	0.195	54	0.091	43	49
51 (K)	0.216 0.219	57 57	0.121 0.124	52 53	55
56	0.182 0.227	52 59	0.098 0.110	46 49	52
59	0.204 0.168	56 49	0.102 0.100	46 46	49
62	0.293	65	0.145	57	61
70 (L)	0.022	4	0.015	9	7

TABLE A-3 (Cont'd)

SAMPLE NO.	PEAK AREA ARAG. 26.2 / CAL. 29.4	WT. % ARAG.	PEAK AREA ARAG. 27.2 / CAL. 29.4	WT. % ARAG.	AV. WT. % ARAG.
72 (M,N)	0.072	20	0.043	25	23
74 (O)	0.013 0.008	2 0	0.005 0.000	0 0	1
[Q4] 75(T)	0.027 0.036	6 8	0.013 0.020	7 13	9
78	0.000 0.000	0 0	0.000 0.000	0 0	0
[Q3] 80(Q)	0.011 0.014	1 2	0.000 0.011	0 5	2
83	0.012 0.014	1 2	0.006 0.008	0 0	1
95	0.000 0.000	0 0	0.000 0.000	0 0	0

TABLE A-4. X-RAY DATA, BOREHOLE BP-3

SAMPLE NO.	PEAK AREA ARAG. 26.2 / CAL. 29.4	WT. % ARAG.	PEAK AREA ARAG. 27.2 / CAL. 29.4	WT. % ARAG.	AV. WT. % ARAG.
[Q5] 2(B)	0.349 0.402	68 71	0.191 0.202	64 66	67
9(C)	0.185	52	0.097	45	49
12	0.287	64	0.156	59	62
17(D)	0.024 0.029 0.041	5 6 10	0.017 0.020 0.016	10 13 9	9
24	0.057	15	0.024	15	15
30(E)	0.051 0.058	13 16	0.025 0.033	16 20	16
40	0.043 0.054	11 14	0.022 0.026	14 16	14
42(F)	0.049 0.053	13 14	0.028 0.030	18 19	16
51	0.054 0.031	14 7	0.029 0.017	18 10	12
52	0.207	56	0.107	48	52
[Q4] 48(G)	0.021 0.015	4 2	0.009 0.007	0 0	2
59	0.015 0.019	2 3	0.012 0.000	6 0	3
68(L)	0.021 0.023	4 5	0.018 0.017	11 10	8
76(M)	0.008 0.007	0 0	0.008 0.000	0 0	0

APPENDIX E - SAMPLE PREPARATION

The nine slabbed cores were examined on a 'reconnaissance' level in order to record facies changes visible on a macroscopic scale. Approximately 140 oriented thin section chips were cut, the choice of location of the thin sections was based on the intention to sample typical lithology and any variations. The relatively friable nature of most of the samples necessitated epoxy impregnation under vacuum. A mixture of 1:1 Petropoxy (Palouse Petro Products) and toluene was prepared. The chips were immersed in the solution in an aluminum tray and then put under 584 mm (23 inches) Hg for 5 to 10 minutes. The vacuum was released and reformed several times. The chips were removed from the solution and placed on a hot plate at approximately 125°C to initiate curing of the epoxy mixture. The epoxy impregnated chips were then baked in the vacuum oven (normal atmospheric pressure) for five to six hours to complete the curing process. The above method of preparation is similar to that in Hutchison (1974).

For SEM study, broken specimens were mounted on stubs with double-sided tape and gold-coated. The unit used was a Cambridge Stereoscan SEM.

APPENDIX F - STAINING PROCEDURE

Fiegl's Solution: This solution is used to differentiate aragonite from calcite. 1 g Ag_2SO_4 is mixed with 11.8 g $MnSO_4.7H_2O$ in 100 ml distilled water and the solution is boiled. After cooling the suspension is filtered off and one or two drops of dilute NaOH solution is added. After 1 or 2 hours the precipitate is filtered off. The solution longevity is significantly increased if kept in a dark bottle away from light. (Katz and Friedman, 1965)

To use the solution, polished samples are etched in dilute HCl. The surface to be stained is immersed in the solution for several minutes (depending on solution freshness). Aragonite is stained black while calcite remains clear. The stain was used to provide qualitative insight into mineralogy.

Alizarin Red S and Potassium Ferricyanide Stain: The solution is used to differentiate dolomite from calcite and ferroan calcite from non-ferroan calcite. Dolomite remains colourless, calcite stains red (non-ferroan), and ferroan calcite blue-purple.

The solution is made as follows: 1 g Alizarin Red S stain, 5 g potassium ferricyanide, and 2 ml of concentrated hydrochloric acid are brought to 1 L of solution with distilled water. (Katz and Friedman, 1965).

Clayton Yellow Stain: This stain is used to differentiate Mg-calcite from calcite. The colour of Mg-calcite varies from pale pink in crystals with little $MgCO_3$ (5-8%), to deep red in those with high $MgCO_3$

content. The stain is made by combining 1 g Clayton Yellow powder, 8 g NaOH, 4 g EDTA, and 1 L distilled water. These are mixed at room temperature and stored in an amber glass bottle. A procedure to render the stain permanent is also available (Choquette and Trussell, 1978).

For both the Alizarin Red S - Potassium Ferricyanide and Clayton Yellow stains, the procedure for sample preparation is similar to that outlined for Fiegl's solution.

APPENDIX G - METHODOLOGY FOR CONSTRUCTION OF MAPS OF FRESHWATER PHREATIC LENSES ON BIG PINE KEY

The data used for the delineation of the freshwater phreatic lenses of Big Pine Key are from Hanson (1980, table 6). The chloride concentration (mg/L) - depth relationship in 22 uncased wells (Figure 7-1, text) drilled by the USGS Water Resources Division was monitored during a study period between June, 1976 and April, 1977 at approximately 1 month intervals. Chloride concentrations were usually measured in 1.52 m (5 ft) intervals. The deepest measurement was made at the hole bottom and consequently this last interval may be slightly larger or smaller than 1.52 m. Data from wells 2A and 6A were not recorded for the period between June, 1976 to February, 1977.

Based on wells 2A and 6A, specific conductance profiles indicate that water salinity is relatively constant to a depth of approximately 5.5 m (18 ft) below the water table (Hanson, 1980, figs. 6,7). Below this depth, salinity increases at a faster rate. Obviously, with wells located on the outer edges of the lens, the depth at which salinity begins to increase rapidly would be shallower.

A [Cl^-] value of 500 mg/L is used as the lower boundary of the meteoric lens. Plotting of the [Cl^-] data with depth (1.52 m intervals) did not always show a smooth increase. Consequently, the depth of the 500 mg/L [Cl^-] was established by simple linear extrapolation between the two recorded values less than and greater than 500 mg/L (Table A-5). For reference, according to Deju (1971), water for drinking purposes should

TABLE A-5. DATA USED TO CONSTRUCT FRESHWATER PHREATIC LENSES OF BIG PINE KEY

USGS WATER RESOURCES WELL NO.	SEPT., 1975 DEPTH TO 500 mg/L [Cl−]* (m rel. to sea l.)	[Cl−] (mg/L) AT WATER TABLE*	MARCH, 1977 DEPTH TO 500 mg/L [Cl−]* (m rel. to sea l.)	[Cl−] (mg/L) AT WATER TABLE**
1	-0.5	200	NFW	860
2	<-5.9	48	<-7.1	350
2 A	ND	ND	-6.8	290
3	NFW	950	NFW	960
4	-2.3	120	NFW	560
5	-5.3	56	-6.2	320
6	<-7.3	180	NFW	610
6 A	ND	ND	NFW	680
7	<-7.1	110	NFW	780
8	NFW	5800	NFW	1300
9	NFW	1400	NFW	1400
10	<-6.8	96	-4.9	370
11	<-6.0	51	<-7.0	340
12	<-6.4	280	NFW	600
13	-2.8	130	NFW	520
14	-4.6	140	NFW	650
15	<-6.1	87	NFW	700
16	NFW	530	NFW	2900
17	0	500	NFW	1600
18	-3.9	55	-2.8	450
19	-1.9	150	NFW	740
20	-0.6	220	NFW	850

* calculated from Hanson, (1980, table 6)
** from Hanson (1980, table 6)
NFW: no freshwater as defined
ND: no data

not have over 225 mg/L [Cl—]. Waters having in exess of 500mg/L [Cl—] have an unpleasant taste. Although the lens boundary has been defined based on the 500 mg/L [Cl—] value, the transition from freshwater to normal marine water is a gradual one. The dispersion zone is estimated to be approximately 3 metres (10 ft) thick (Hanson, 1980, p. 16). The change of the rate of increase of [Cl—] with depth at approximately 5.5 m below the water table indicates the top of the dispersion zone.

An approximation of the height above sea level of the water table is based on data from wells 2 and 6 in Hanson (1980, table 5). For his study period, the monthly average water level was 0.35 m (1.14 ft) and 0.37 m (1.20 ft) for wells 2 and 6, respectively (values above mean sea level). The average value is 0.36 m, and it is this value that is used as the average water table height. Serious deviations from this are not expected as the nature of the island topography is fairly constant with an average height above sea level of approximately 1 m (see Hanson, 1980, table 3). Furthermore, the figures quoted for wells 2 and 6 are averages of the daily maxima and daily minima. In actualilty, the water table fluctuates 0.26 m (0.85 ft) daily, on average (calculated from Hanson, 1980, table 5). Using 1 m as the average elevation of land above sea level, the average vadose zone thickness (height of land surface minus water table height) is approximately 0.6 m.