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ASPECTS OF DIAGENESIS AND ICHNOLOGY IN THE OLIGOCENE-MIOCENE BLUFF FORMATION OF GRAND CAYMAN ISLAND, BRITISH WEST INDIES

by

SUZANNE MARIE PLEYDELL

A THESIS

SUBMITTED TO THE FACULTY OF GRADUATE STUDIES AND RESEARCH
IN PARTIAL FULFILMENT OF THE REQUIREMENTS FOR THE DEGREE
OF MASTER OF SCIENCE

DEPARTMENT OF GEOLOGY

EDMONTON, ALBERTA

FALL, 1987

ISBN 0-315-41015-9

THE UNIVERSITY OF ALBERTA

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DEGREE FOR WHICH THESIS WAS PRESENTED MASTER OF SCIENCE

YEAR THIS DEGREE GRANTED FALL, 1987

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Supervisor

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....

Fred J Longstaffe

Date June 12, 1987.

Dedication

Dedicated to my parents.

Abstract

The Oligocene-Miocene Bluff Formation of Grand Cayman Island has a complicated diagenetic history resulting from the passage of the rocks from the marine phreatic, through the mixing and freshwater phreatic to the vadose diagenetic zones, in response to tectonic and/or sea level fluctuations. The result is a highly porous dolostone produced by extensive aragonite and dolomite dissolution.

Differential aragonite dissolution of skeletal material (largely coral molds), revealed abundant borings that were preserved as casts. 75% of the borings are dominated by *Entobia. Trypanites, Gastrochaenolites, Maeandropolydora, Talpina, Caulostrepsis,* microborings and two new ichnogenera are also locally abundant. It is apparent that bioerosion during middle Tertiary times was comparable, in terms of both the organisms involved and the scale of damage, to bioerosion in modern reefs.

Pervasive dolomitization which occurred prior to or penecontemporaneous with aragonite dissolution, took place away from the sediment-water interface. Petrographic, and oxygen and carbon stable isotope geochemistry indicate that dolomitization occurred in normal marine to slightly hypersaline waters, prior to the initial emergence of the island.

Emergence of the island after middle Miocene times saw the influx of meteoric waters into the dolostones. This led to the development of poikilotopic calcite from progressively fresh porewaters of the mixing and freshwater phreatic zones. The development of this poikilotopic texture can be considered a dedolomitization process through time separated stages of (1) precipitation of limpid dolomite cements, (2) dissolution of the cores of those cements, and (3) precipitation of sparry calcite cements in and around the hollow dolomite rhombs. As the rocks moved into the vadose zone, calcite dissolution, $CaCO_3$ cementation, dedolomitization by concomitant dolomite replacement by micrite, and sparmicritization occurred. Since this initial emergence of the island, the dolostones have remained predominantly in the vadose zone.

Despite the similarities of the dolostones of the Bluff Formation with the Tertiary and Quaternary dolostones of other Caribbean areas, dolomitization was not a synchronous Caribbean-wide event.

Acknowledgements

I am indebted to my supervisor, Dr. Brian Jones, for his guidance, encouragement and support during the course of this study. I am especially appreciative of his editorial skills which have improved this thesis considerably.

I am grateful to George Braybrook for his invaluable assistance and expertise with the Scanning Electron Microscope; to Dr. Fred Longstaffe for allowing me use of the isotope laboratory facilities and for reviewing parts of this thesis; to Diane Caird for her assistance with isotopic analysis; to Dr. George Pemberton for his advice in the field, and to Ian Hunter for his helpful discussions on coral taxonomy. My thanks extend to Leni Honsaker fot her generous assistance with photography, German translations and many other aspects of the thesis, and to my fellow graduate students, especially Duncan Smith and Mark Hadley, for their encouragement and moral support.

The thesis was made possible by financial support provided by the Natural Sciences and Engineering Research Council of Canada (Grant No. A6090 to Dr. Brian Jones). Accommodation on Grand Cayman Island was kindly provided by Sam Ng of the Cayman Water Authority.

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I. INTRODUCTION

The origin of dolomite in pervasively dolomitized sequences of rock is still in debate, even after decades of extensive investigation (e.g. Zenger *et al.*, 1980; Morrow, 1982a, 1982b; Land, 1983a, 1985; Machel and Mountjoy, 1986; Hardie, 1987). The idea that dolomite may undergo a series of recrystallization events during burial diagenesis (Land, 1980, 1983a), which may result in modification of both the petrographic and isotopic signatures of the dolomite, renders interpretation of ancient dolomites difficult.

The pervasively dolomitized coralgal rocks of the Bluff Formation on Grand Cayman Island studied herein are particularly suitable for the investigation of dolomitization, since they are relatively young and have not been subject to any burial since deposition. Hence, the dolomites probably have not undergone any significant recrystallization since formation, and the petrographic and stable isotopic data may be indicative of the original dolomite that formed.

Dolomitization of the rocks of the Bluff Formation was associated with aragonite dissolution, which has produced a very porous dolostone. A complex sequence of dolomite and calcite cements have been precipitated in the pores, cavities and skeletal molds, reflecting the fluctuating porewater chemistries of the shallow diagenetic regimes through which the rocks have evolved. Both the dolomitization textures and the cement sequences have been compared to similar textures and cements in dolostones of other Caribbean areas (Sibley, 1980; Jones *et al.*, 1984). On the basis of similarities between the Seroe Domi Formation of Bonaire with the dolostones of the Hope Gate Formation of Jamaica as described by Land (1973a, 1973b) and the dolostones of San Salvador, Bahamas as described by Supko (1977), Sibley (1980) suggested that dolomitization events were common to the entire Caribbean area. Jones *et al.* (1984) included the Bluff Formation of Grand Cayman Island and the dolostones of Great Abaco Island, Bahamas (described by Kaldi and Gidman,1982) into the list of similar dolostones and argued that the same sequence of diagenetic events should be found on most islands of the Caribbean. Ward and Halley (1985) reported dolomitized limestones from the Yucatan Peninsula, that comprised a similar cement sequence. However, they refuted the idea of regional dolomitization events, and emphasized the importance of individual environmental and hydrological characteristics in any one area. The detailed study of dolomitization of the rocks of the Bluff Formation of Grand Cayman Island provides a good opportunity to test such a hypothesis.

The dissolution of skeletal components during diagenesis of the Caymanian carbonates has revealed well preserved casts of borings of endolithic organisms. Even though dolomitization has removed many of the original skeletal components in which the borings occur, the casts of the borings themselves have been preserved in fine detail. The abundance and diversity of the borings attest to the significance of bioerosion in middle Tertiary times. Detailed investigations of ancient bioerosion assemblages in reef environments are rare (e.g. Kauffman and Sohl, 1974; James *et al.*, 1977; Warme, 1977), although studies of bioerosion in recent reefs are extensive. Thus, the good preservation, abundance and diversity of the borings preserved in skeletal molds of the Bluff Formation provide an excellent opportunity to carry out such a study, and to contrast the bioerosion assemblage preserved with those of modern reefs.

A. OBJECTIVES

The main objective of this thesis is to provide an overall documentation of the borings and diagenesis (especially early diagenesis) of the Bluff Formation based on surface outcrop alone, as such an overview has not been attempted since the investigations of Matley (1926). The borings, in particular, have never been investigated previously. The specific objectives are:

1. To define the age of the Bluff Formation on Grand Cayman Island.

2. To identify the borings that occur in the skeletal molds of the Bluff Formation.
3. To describe and document the new ichnogenera and ichnospecies that are present.
4. To compare and contrast the fossil boring assemblage of the Bluff Formation with that of modern assemblages found in reef environments.
5. To decipher the complex sequence of diagenetic events that have affected the rocks of the Bluff Formation, with the aid of petrographic (light microscopy, cathodoluminescence and SEM) and geochemical (stable isotope) analysis.
6. To derive a dolomitization model that is compatible with the petrographic and geochemical evidence.

B. METHODS

Forty seven samples were chosen for thin section study. Forty one of these were ground to 30 μm and impregnated with blue epoxy to enhance porosity. The remaining 6 were ground to 60 μm, impregnated with blue epoxy and finely polished for cathodoluminescence study. All thin sections were stained with alizarin red solution and potassium ferricyanide in dilute HCL (Evamy, 1963; Dickson, 1965), to distinguish dolomite and calcite and to determine the presence of any ferroan counterparts. No ferroan carbonates were discovered. Thin sections were examined on a Jena polarizing microscope. Relative abundance of components was determined by visual estimation using the methods of Terry and Chilingar (1955). Thin sections prepared for cathodoluminescence study were examined on a Technosyn Cold Cathode Luminescence Model 8200 Mark II.

Fifty seven samples were prepared for Scanning Electron Microscopy (SEM). Small fractured samples were mounted onto SEM stubs using conductive paint. These were then sputter coated with gold and analysed on a Cambridge Stereo Scan 250 SEM. An attached EDX allowed element analysis in conjunction with morphological study.

Samples chosen for isotopic analysis were first examined by X-ray powder diffraction using either a Phillips or Rigaku Geigerflex X-ray diffractometer, to determine mineral composition. Samples were manually crushed to a fine powder using a pestle and mortar. The powder was then smeared onto a glass slide with alcohol. The XRD's were operated at 50kV 20mA, using CoKα radiation, and the samples were run between 25° and 52° at 1°2θ/min. Quartz was used as an internal standard. Relative proportions of calcite and dolomite were estimated using the calcite- dolomite peak $d_{(104)}$ ratios calibrated on the curve determined by Royse *et al.* (1971) after Tennant and Berger (1957).

Dolomite nonstoichiometry was determined by measuring the position of the $d_{(104)}$ spacing (Goldsmith and Graf, 1958). The $d_{(104)}$ spacing of the dolomite lattice is shifted due to the incorporation of calcium ions in excess of the stoichiometric ($Ca_{0.5}Mg_{0.5}CO_3$) dolomite composition. The stoichiometric composition of dolomite (50% $CaCO_3$) is reflected in a $d_{(104)}$ value of 2.886Å. Lumsden (1979) used the equation:

$$^{N}CaCO_3 = md + b,$$

where $^{N}CaCO_3$ is the mole % $CaCO_3$ in dolomite, m is 333.33, *d* is the observed $d_{(104)}$ value, and b is -911.99. Other elements may substitute for the cations in the lattice and thus influence the peak positions, but only iron may be abundant enough to pose a real problem (Runnells, 1974). No samples of Bluff Formation dolostone reacted with potassium ferricyanide stain and therefore iron is not abundant enough (i.e. less than 1% FeO) to shift the peaks (Lindholm and Finkelman, 1972; Lumsden and Chimahusky, 1980).

Fifty one samples were chosen for isotopic analysis and were powdered (< 44 μm) either with a dental drill or manually with a pestle and mortar. The procedure of Walters *et al.* (1972), modified after McCrea (1950) was then followed. The samples were reacted with 100% H_3PO_4 at 25.2°C. In samples containing both dolomite and calcite, the calcite CO_2 was extracted after 1 hour, any subsequent gas produced was pumped away after 3 hours, and the dolomite CO_2 was extracted after 10 days. The gases were then analyzed on a 602D Micromass Mass Spectrometer. To calculate the results of calcite and dolomite, values of 1.01025 and 1.01110 respectively were used for the carbonate - phosphoric acid carbon dioxide fractionation factor at 25°C (modified after Sharma and Clayton, 1965). Accuracy is better than 0.2‰. Results are presented in the δ (‰) notation relative to the *Belemnitella americana* from the Peedee Formation.

Rough percentages of amount of boring were determined by visual estimation of both hand and field samples. More accurate determination was made on a limited number of slabs (10), which were impregnated with liquid fibreglassing resin before slabbing. These slabs were taken from the middle of corals and were assumed to be representative of the whole coral. Acetate overlays were used to trace and quantify the amount of boring. Most of the coral molds represent only fragments of the original coral colony. This must be considered when comparing the data obtained with determinations made from slabs from whole colonies in the recent environment. The borings were analysed using a binocular microscope, and in some cases thin section and SEM study was implimented.

All samples collected and studied herein are documented in brackets, in the relevant sections, by locality abbreviation (refer to Figure 1B) and University of Alberta catalogue number. Specimens of borings illustrated in this thesis have University of Alberta Type Collection numbers, while those not illustrated but described, have the same numbering sequence as the petrographic samples.

C. LOCATION

The three Cayman Islands are located in the western part of the Caribbean Sea, between Jamaica and Cuba (Fig.1A). Grand Cayman (81°15'W, 19°20'N), the largest of the islands, is approximately 280 km north northwest of Jamaica and 240 km south of Cuba. Cayman Brac and Little Cayman are approximately 95 km northeast of Grand

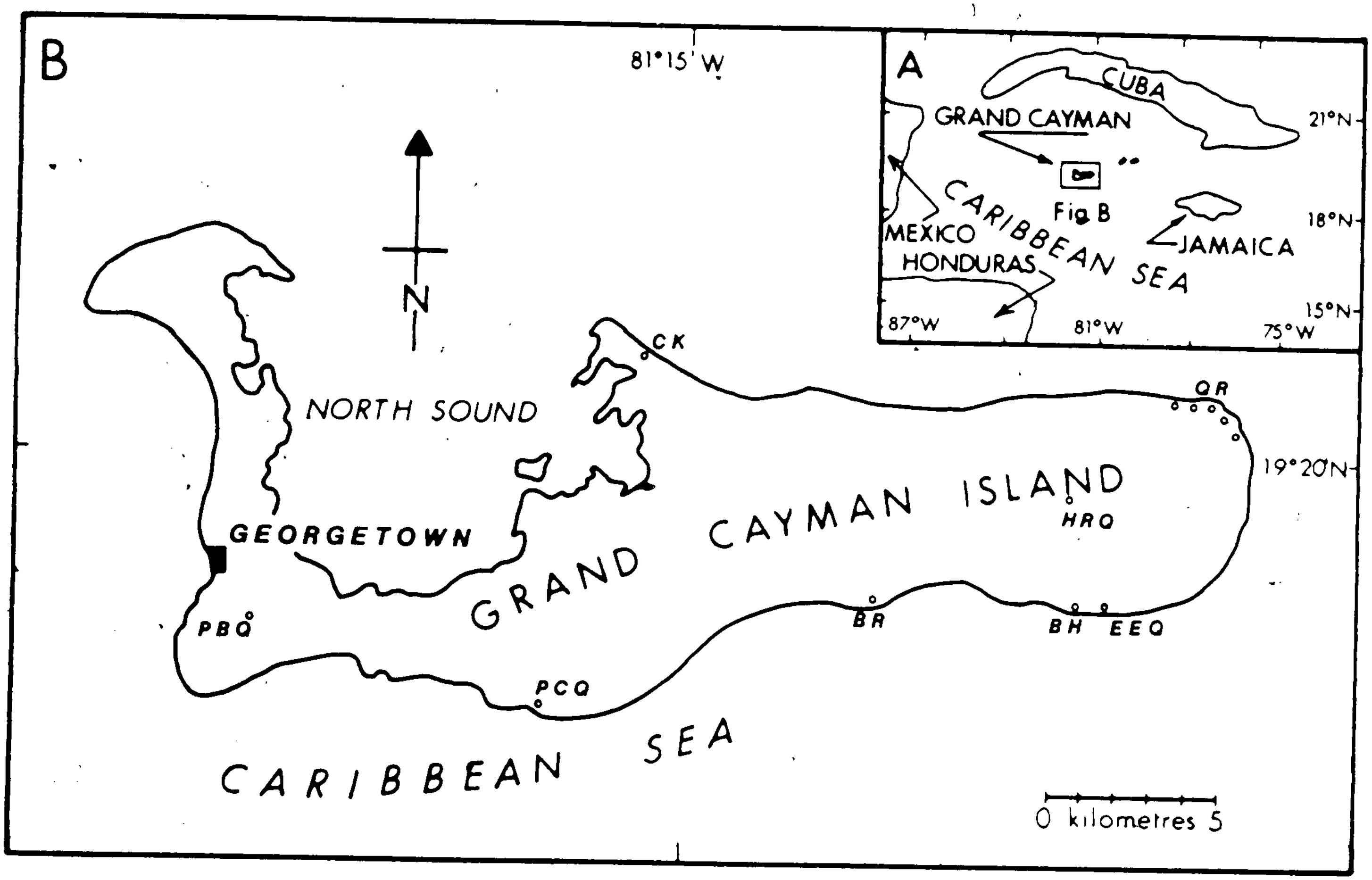


Figure 1. Location map of Grand Cayman Island. BH, Blowholes; BR, Breakers; CK, Cayman Kai; EEQ, East End Quarry; HRQ, High Rock Quarry; PBQ, Paul Bodden Quarry; PCQ, Pedro Castle Quarry; QR, Queens Road.

Cayman (Fig.1A). The Cayman Islands are peaks on the Cayman Ridge, which is a submarine extension of the Sierra Maestra Range of Cuba (Rigby and Roberts, 1976).

Grand Cayman is 35 km long, east to west, and is between 6 and 14 km wide (Fig. 1B). It encompasses an area of 197 km^2 (Spencer, 1985, p. 60). The maximum elevation is 18 m (compared to 40 m of Cayman Brac), but most of the island is less than 3 m above sea level. The eastern part of the island is generally covered in mangrove swamps, which prevent access to much of the interior of the island. A coastal road provides good access to the perimeter of the eastern end.

D. GEOLOGIC SETTING

The first major geological observations of the Cayman Islands were made by Matley (1926), who described the Bluff and Ironshore formations on all three islands.

On Grand Cayman Island, the Bluff Formation makes up much of the eastern half of the island, and most of the high ground. The rest of the island is formed of the Ironshore Formation (Fig. 2) which is comprised of five "facies": reef, backreef, lagoonal, shoal and beach ridge (Brunt *et al.*, 1973). The Ironshore Formation is a limestone, whereas the Bluff Formation is a dolostone unit (Jones *et al.*, 1984). The Ironshore Formation is late Pleistocene in age (124 000 B.P.) and rests unconformably on the Oligocene-Miocene Bluff Formation (Fig. 3). The contact between the two formations is variable (Brunt *et al.*, 1973), but always well defined.

The Bluff Formation is a dense, very hard, white to beige, crystalline dolostone with local patches of dolomitic limestone. It has vaguely defined bedding at some localities, but otherwise appears massive, and has three well defined joint systems. Corals, molluscs, foraminifera and red algae are common skeletal constituents. Surface exposures of the Formation are characterized by a well developed karst topography, consisting of

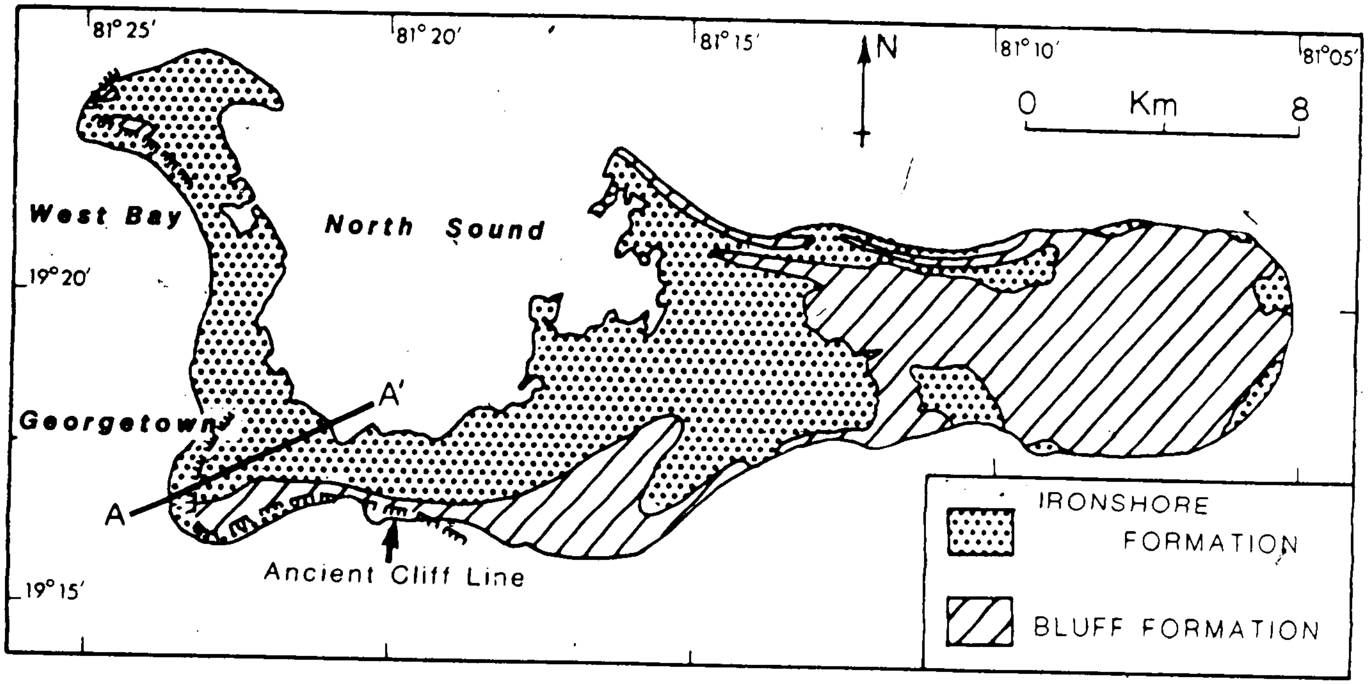


Figure 2. Geological map of Grand Cayman Island (from Brunt *et al.*, 1973)

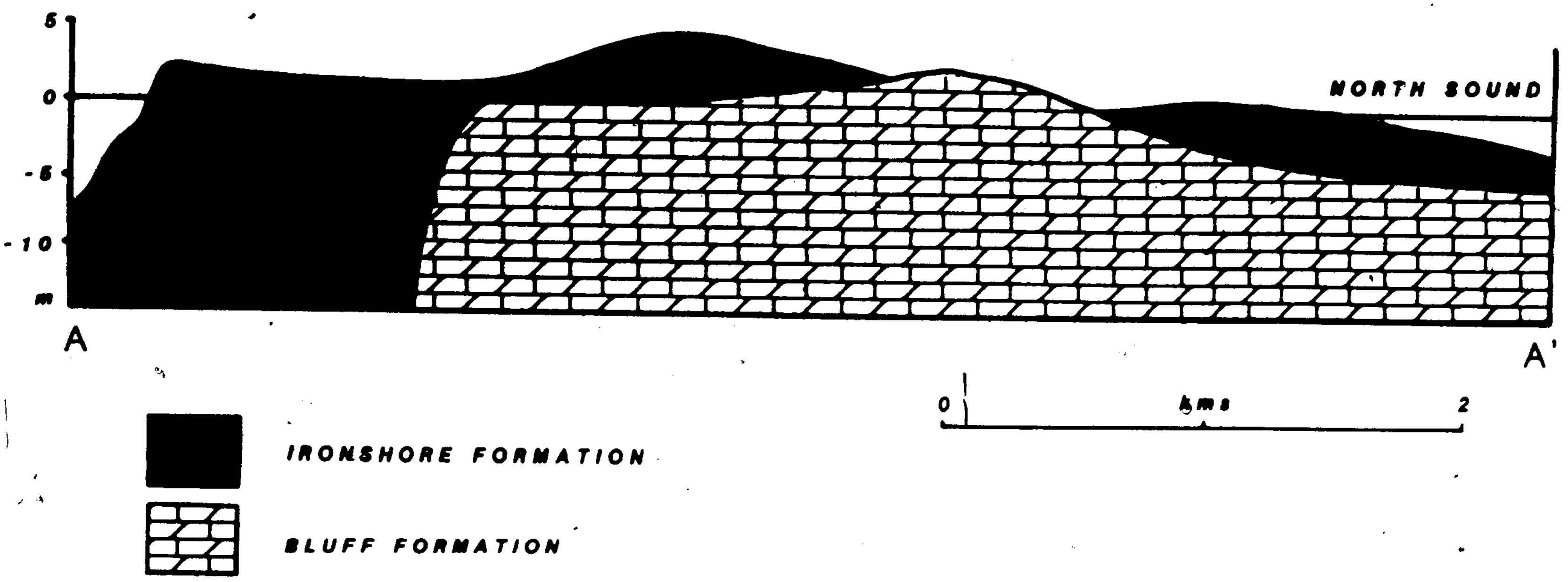


Figure 3. Geological cross-section of Grand Cayman Island (from Brunt *et al.*, 1973)

sharp pinnacles, ridges, sinkholes and fissures (Matley, 1926; Doran, 1954). Black phytokarst caused by the boring activity of microendoliths (Folk *et al.*, 1973) is common.

E. TECTONIC SETTING

The Cayman Islands form pinnacles on top of the asymmetrical Cayman Ridge. This ridge extends from the Sierra Maestra of Cuba almost to British Honduras, where it disappears beneath sediment cover in the Yucatan Basin (Perfit and Heezen, 1978). The northern side slopes gently into the Yucatan Abyssal Plain while the southern slope drops precipitously more than 5000 m into the Cayman Trench, (Rosencrantz and Sclater, 1986).

The Cayman Trench, formally the Bartlett Trough (Taber, 1922), extends from 86°30' W, just north of the Bay Islands, to 77°W, north of Jamaica. It is between 120-180 km wide, and is bounded to the south by the precipitous northern slope of the Nicaraguan Plateau.

The Cayman Trench marks the present day location of the strike slip plate boundary of the northern Caribbean Plate (Molnar and Sykes, 1969). The boundary is largely defined by the Oriente Transform Fault in the northern and eastern side of the Trench, and by the Swan Island Transform Fault along the southern and western edge (Fig. 4). These have left lateral strike slip motions.. Holcombe *et al.* (1973) described a north-south trending valley at 81°40'W, bounded on each side by north-south trending rugged topography. This was interpreted to represent an actively spreading oceanic rise, which connected the two transform faults.

. Rosencrantz and Sclater (1986) suggested that the total opening of the Trench has been 1100 km, and that this has occurred in two stages: an early phase of fast spreading, and a later episode of slower spreading (15 ± 5 mm yr^{-1}). A distinct break between the two occurred 30 Ma ago. Land (1979) argued that the Trench most probably opened during the Eocene.

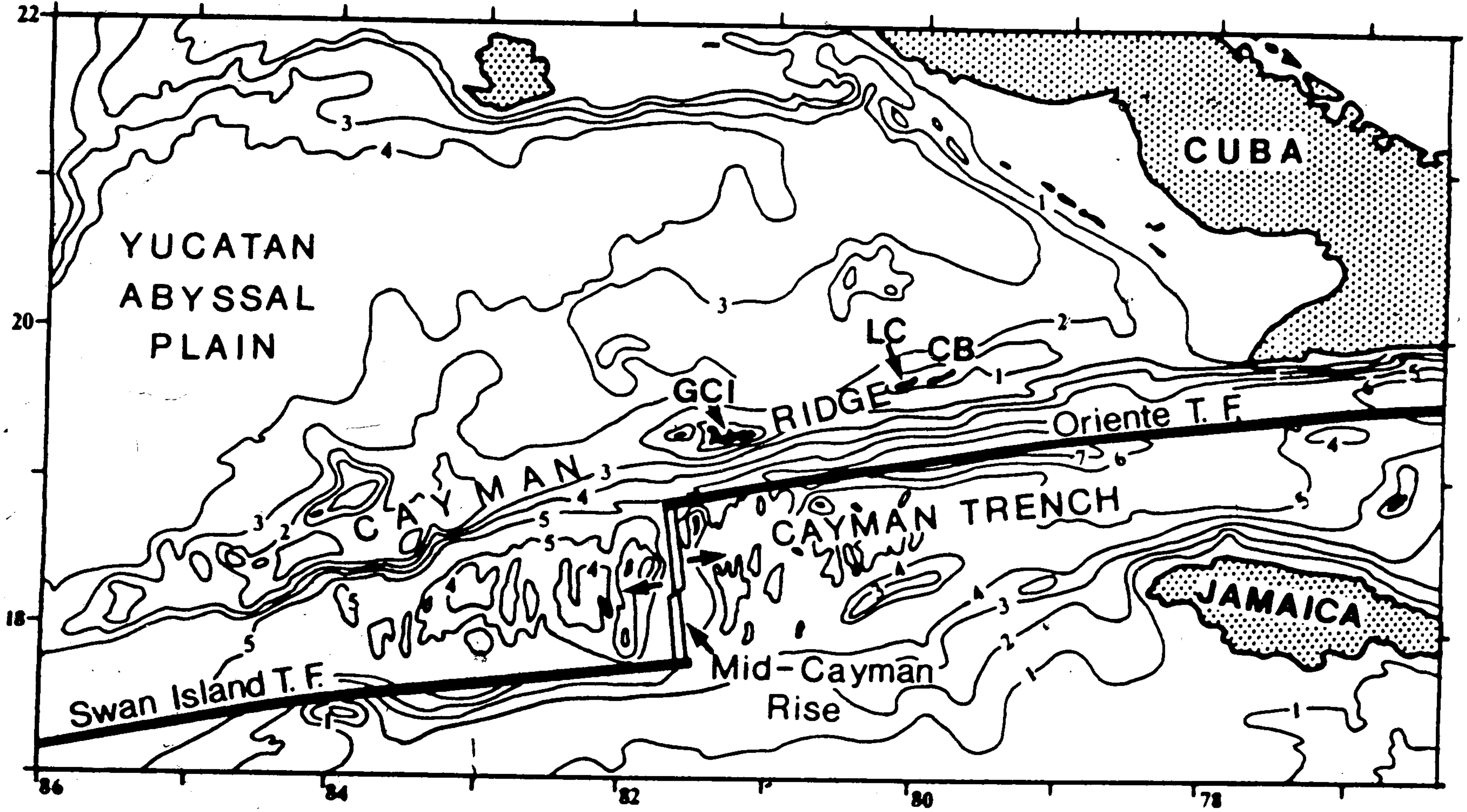


Figure 4. Map showing the tectonic setting of Grand Cayman Island. Water depths in thousands of metres. Figure courtesy of B. Jones.

The activity of the trench can be tied in with Caribbean plate motions. Perfit and Heezen (1978) presented a detailed discussion of the tectonic evolution of the Caribbean during the Mesozoic and Cenozoic. They proposed that the Cayman Trench represents the missing link between Central America and the Greater Antilles. In pre Cretaceous times, the Cayman Ridge and Nicaraguan Plateau were a single unit extending from just east of the Honduras coast. In the middle to late Mesozoic left lateral strike slip faulting began along the ancient Motagua-Polochic Fault Zone. In the early Cretaceous, South America began to rotate clockwise with respect to North America, and southerly subduction of the Atlantic began under the Caribbean Plate. A volcanic arc, represented by Cretaceous rocks from the Cayman Ridge, Nicaraguan Plateau, Jamaica, south Cuba and Haiti, developed above the Benioff zone. Subduction stopped in the early Tertiary and the Caribbean plate began to move relatively eastwards. The southward movement of South America relative to North America caused extension, and led to the opening of the Proto Cayman Trench. Subparallel "leaky" faults developed, until continued plate divergence caused a spreading centre to form. The trench widened and lateral faults became localised along the trench walls. The Cayman Ridge and south east Cuba were rifted from the volcanic arc, and since the Eocene have been increasingly left laterally offset.

The Cayman Ridge was a shallow carbonate bank, but began to subside in the Oligocene (Emery and Milliman, 1980) to Miocene (Perfit and Heezen, 1978). Subsidence rates have been estimated at 6 cm (Perfit and Heezen, 1978) to 10 cm 1000yr^{-1} (Emery and Milliman, 1980). Localised uplift elevated the Cayman Islands, the Swan Islands, Jamaica and South Cuba above sea level after the middle Miocene, and continued to elevate Central America, while subsidence continued in the other areas (Perfit and Heezen, 1978). It has been suggested that the three Cayman islands occur on distinct fault blocks that have been subject to differential vertical tectonic movement until the last interglacial (Woodroffe *et al.*, 1983). Since then, they have been relatively stable (Emery, 1981; Woodroffe *et al.*, 1983).

F. SEA LEVEL

A sea level fall, instead of, or coupled with tectonic uplift, could have accounted for the Island's emergence after the middle Miocene. Hallam (1984) documented a minimum global sea level in the middle to late Miocene, which certainly would have exposed the rocks of the Bluff Formation.

Evidence from the evolution of land snails of Grand Cayman Island suggests that at least part of the island has remained subaerial "...since well back into the Tertiary Period." (Emery, 1981). A well defined marine terrace at 2 m above sea level and higher ones on the island attest to periods of higher sea level in the past. The 2 m terrace can be correlated with marine terraces throughout the Caribbean region (Emery, 1981), and represents the Sangamon Interglacial. During this time, the Ironshore Formation was deposited unconformably on the Bluff Formation. The higher terraces cannot be correlated but probably represent glacio-eustatic fluctuations of the late Tertiary and Pleistocene. Evidence from Bermuda (Harmon *et al.*, 1983) indicated a sea level 1 m higher than present at about 200 000 BP (middle Pleistocene), and Vail and Hardonbol (1979) documented a high of about 100 m above present sea level during the Pliocene. The Bluff Formation of Grand Cayman Island has a maximum elevation of 18 m, and therefore, this Pliocene high would have submerged the island. However, the details and magnitude of Vail and Hardonbol's curve have been questioned (Miall, 1986). Indeed, Hallam (1984) documented a much smaller rise in sea level at this time. Thus, it is probable that Grand Cayman Island has remained largely subaerial since its initial emergence after the middle Miocene.

G. AGE OF THE BLUFF FORMATION

Vaughan (1926) suggested an Oligocene age for the Bluff Formation on Cayman Brac on the basis of Foraminifera, most notably Lepidocyclinids. Vaughan (in Matley,

1926) based on coral and foraminiferal evidence, suggested a Miocene age for the formation on Little Cayman and Grand Cayman. Merren (in Richards, 1955) reported both Oligocene and Miocene foraminifera from the Bluff Formation of Grand Cayman. Gradstein and Poag (in Emery and Milliman, 1980) documented the presence of amphisteginids in the Bluff Formation on Grand Cayman, and attributed a maximum Oligocene age for the formation.

The presence of amphigestinid foraminifera is confirmed by the present study. In addition, a few specimens of Lepidocyclinids, abundant *Stylophora* sp., *Diploria* sp., *Montastrea* sp., *Agaricia* sp., *Agathiphyllia* sp., *Antillocyathus* sp., *Dichocoenia* sp., poritid and faviid corals; *Diodora* sp. and *Siliquaria* (?) sp., not hitherto documented from the Bluff Formation of Grand Cayman, have also been found. These suggest that the Bluff Formation has an upper limit of middle Miocene age. This conclusion is supported by the ideas of Perfit and Heezen (1978) who suggested that the Cayman Islands were uplifted after the middle Miocene, and by Hallam (1984) who documented a massive eustatic sea level fall in the middle to early-late Miocene.

II. THE BORINGS OF THE BLUFF FORMATION

Abundant ancient biogenic sedimentary structures, or trace fossils, are well preserved in the Bluff Formation. They are the product of contemporaneous boring activities of marine organisms, many of which had no preservable hard parts. This boring assemblage is thus an important testimony to the diversity of the marine community. The borings occur as natural casts, which preserve much of the fine detail of the original borings in negative form. Examples of such excellent preservation are rare. Unless otherwise stated, the terminology used (appendix I) is consistent with that of Ekdale *et al.* (1984, p. 301-316).

The boring assemblage can be divided into macro- and microborings, based on the size of the individual borings. Even though microorganisms are usually less than 100 µm in diameter (Golubic *et al.*, 1975), Schroeder (1972) found calcified filaments up to 300 µm. The microborings can therefore be considered as those under 300 µm in diameter. Macroborings are those over 300 µm in diameter. The significance of this size distinction stems from the different organisms involved in the two categories. Microborings are predominantly produced by algae, fungi and bacteria, but some of the common macroborers such as sponges, bryozoans and annelid worms, may sometimes produce borings which fall into the upper size range of the microborings. Thus, it is a convenient working boundary rather than a definitive one and is useful for descriptive purposes.

A. PRESERVATION

Preservation of the Macroborings

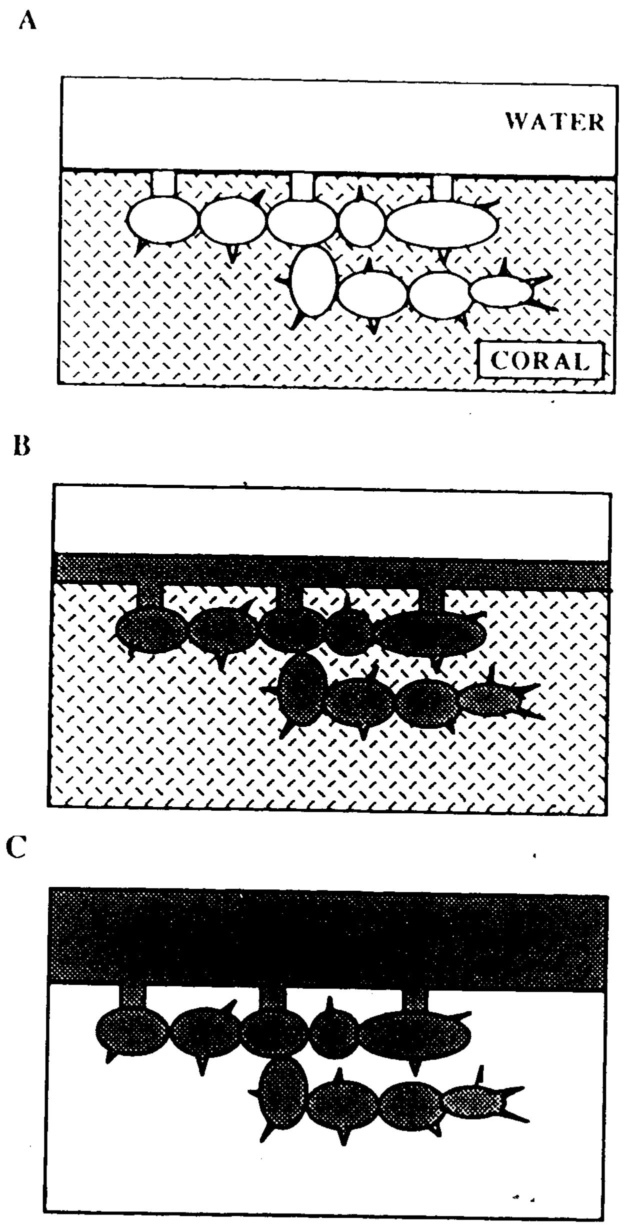
The casts of the macroborings are most common in corals or in coral molds, which are the most common faunal element in the Bluff Formation. The borings also occur on the internal casts of gastropods and bivalves, and less commonly, in the coralline algal crusts which coat many of the coral molds. Kleemann (1982) documented a similar type of differential preservation in the Miocene Leithakalk of Austria, wherein, trace fossils made by large boring bivalves are preserved in coral molds.

The casts of the borings have been preserved by dolomite, which is petrographically similar to that in the host rock. The presence of skeletal allochems in the dolomite in the casts indicates that the borings were passively filled with sediment. The sediment that originally filled the borings was significantly different from the aragonitic skeleton of the host coral. The fact that even the smallest chambers of the borings are filled, attests to the fine grained nature of the original sediment.

The filling of the borings by sediment may have killed the endoliths. This may have been a result of either high sedimentation rates or burial of the host skeletons. In the case of the corals this could have occurred after the endoliths had weakened the bases of the corals and caused them to collapse. At most localities (all except BH) coral fragments rather than whole corals are common. Thus, the endoliths may have contributed to their own demise.

The sediment in the casts was then lithified and later dolomitized along with the surrounding matrix. Preferential leaching of aragonitic skeletons occurred prior to or penecontemporaneous with pervasive dolomitization. This formed skeletal molds in which only the boring casts were left (Fig. 5). The dissolution of the molluscs also left internal and external molds of the valves and exposed the borings. The originally HMC (high magnesium calcite) coralline algal crusts have been largely preserved and so, the borings are only visible in thin section, where they disrupt the algal laminations. Minor dolomite cementation, in some cases followed by calcite cementation, then occurred on the casts of the borings. This cementation has obscured some of the fine detail of the borings.

Some of the borings are now hollow (Plate 8C, F) suggesting that either some dissolution of the lithified fill occurred penecontemporaneously with its replacement by dolomite, or dissolution of the dolomite in the central region of the borings has occurred after dolomitization.



Boring in aragonitic skeletal substrate, with numerous apertures connected to the substrate surface.

Marine sediment fills borings, and is lithified, producing boring casts.

Diagenetic fluids preferentially leach aragonitic host skeleton, but not the infilled borings. The borings are subsequently dolomitized.

Figure 5. Preservational sequence of the macroborings of the Bluff Formation.

The sequence of events can thus be summarized (Fig. 5):

1. Boring of the skeletons,
2. Filling of the borings with sediment,
3. Lithification of the sediment fill,
4. Dolomitization and penecontemporaneous differential dissolution of aragonite skeletons to expose natural casts of the borings in the molds,
5. Dolomite and calcite cementation around boring casts and in skeletal molds.

Preservation of the Microborings

There are three types of microborings and filaments in the Bluff Formation:

1. recent, uncalcified filaments,
2. calcified filaments,
3. dolomitized microborings.

The first two were not part of the original endolithic community, as they are not dolomitized and are therefore younger than the dolomitized microborings. The dolomitized microborings are inferred to be Oligocene to Miocene in age like the macroborings, because they are intricately associated with the submarine macroborings.

The borings range in diameter from 2 to 300 µm; however, their lengths are generally indeterminable. Their common occurrence in microcavities means that much of their structure is hidden, thus making taxonomic identification difficult. They are inferred to be either algal or fungal in origin; Golubic *et al.* (1975) discussed the problems of differentiating these. It is possible that some of the microborings were formed by sponges, worms or bryozoans.

The casts of the microborings are formed of small subhedral to euhedral dolomite rhombs 1 to 4 µm long (Plates 11E, F; 7F), which are comparable to the dolomite cement commonly found lining pores in the Bluff Formation. On many microborings, larger (up to 30 µm long) euhedral dolomite crystals may completely coat the fine dolomite or may occur as isolated crystals on the microborings (Plate 11E, F). Some of the microborings are hollow (Plate 7G).

The microborings commonly occur in microcavities, that average 200-600 µm in diameter. The possibility exists that some of these may not be borings *sensu stricto* , but may be chasmoliths, passively nestling in a pre-existing cavity. It is more probable, however, that the filaments existed as both euendoliths boring through the substrate, until they reached cavities whereupon they became chasmoliths. The euendolithic mode was again established at the other side of the cavity. This is a common mode of life for marine endolithic green algae (Kylin, 1936; Schroeder, 1972). Since it is not possible to distinguish between these modes of life in the Bluff Formation, the term 'microborings' is considered appropriate.

There are three possible alternatives for the preservational sequence of these microborings (Fig. 6).

(1) They have been preserved as natural casts, in a similar manner to that of the macroborings. The microorganisms bored into skeletons producing cavities. Fine grained sediment then settled into the borings, was lithified and later dolomitized. The aragonitic host substrate was preferentially leached prior to, or simultaneous with dolomitization. Dolomite cement then grew on the microboring casts. This mechanism may have been valid for the larger microborings, but it is difficult to envisage even fine grained sediment settling into the smallest borings.

The hollow microborings may have formed during dolomitization, whereupon only part of the sediment fill was dolomitized. Alternatively, the dolomite in the cores of the borings may have been leached later in diagenesis.

(2) The microorganism formed a cavity producing a microenvironment conducive to the precipitation of high magnesium calcite (Schroeder, 1972) or aragonite (Harris *et al.*, 1979). Cement was then able to precipitate in the boring. The cause of such cementation is problematic. The boring organism could have been directly or indirectly responsible for it,

or it could have been triggered by bacteria, or by passive inorganic precipitation (Bathurst, 1966; Schroeder, 1972; Kobluk and Risk, 1977a, 1977b). The cemented boring was then dolomitized, with pene-contemporaneous leaching of the aragonitic host substrate. Subsequently, dolomite cement was precipitated on the dolomitized boring cast.

The hollow centres may have formed during the initial cementation in the boring, where cement was precipitated only as a lining or envelope in the boring. Liljedahl (1986) documented such incomplete cementation in microborings in the Silurian of Gotland. Subsequent dolomitization thus preserved only this internal envelope.

(3) A similar mechanism to (2) can be invoked to explain the preservation of the borings of chasmolithic organisms. During life, or following partial decay of the organism, HMC or aragonite cement was precipitated around the filament. This first stage fringing cement thus preserved the outline of the filament. When the organism decayed completely the internal core was cemented by a second stage cement. Schroeder (1972) called these cemented filaments, "calcified filaments". In the Bluff Formation, these calcified filaments were then dolomitized, and dolomite cement was subsequently precipitated. Hollow cores of the borings could have resulted from incomplete calcification of the filament. The chasmolithic filaments may have undergone the first stage cementation, thus preserving the shape of the filament. The second stage of filling was inhibited or restricted to a lining only. Thus, subsequent dolomitization only preserved the first stage cement.

The possibility exists that some of these microborings may have originated as filaments that were primarily cemented by dolomite and these would therefore be termed "dolomitized filaments". Unfortunately. there is no way to distinguish these in the Bluff Formation. Mechanism (1) could account for the larger microborings, while (2) and (3) seem more likely for the smaller ones. It is not possible to distinguish these possible mechanisms.

The preservational sequence can thus be summarized:

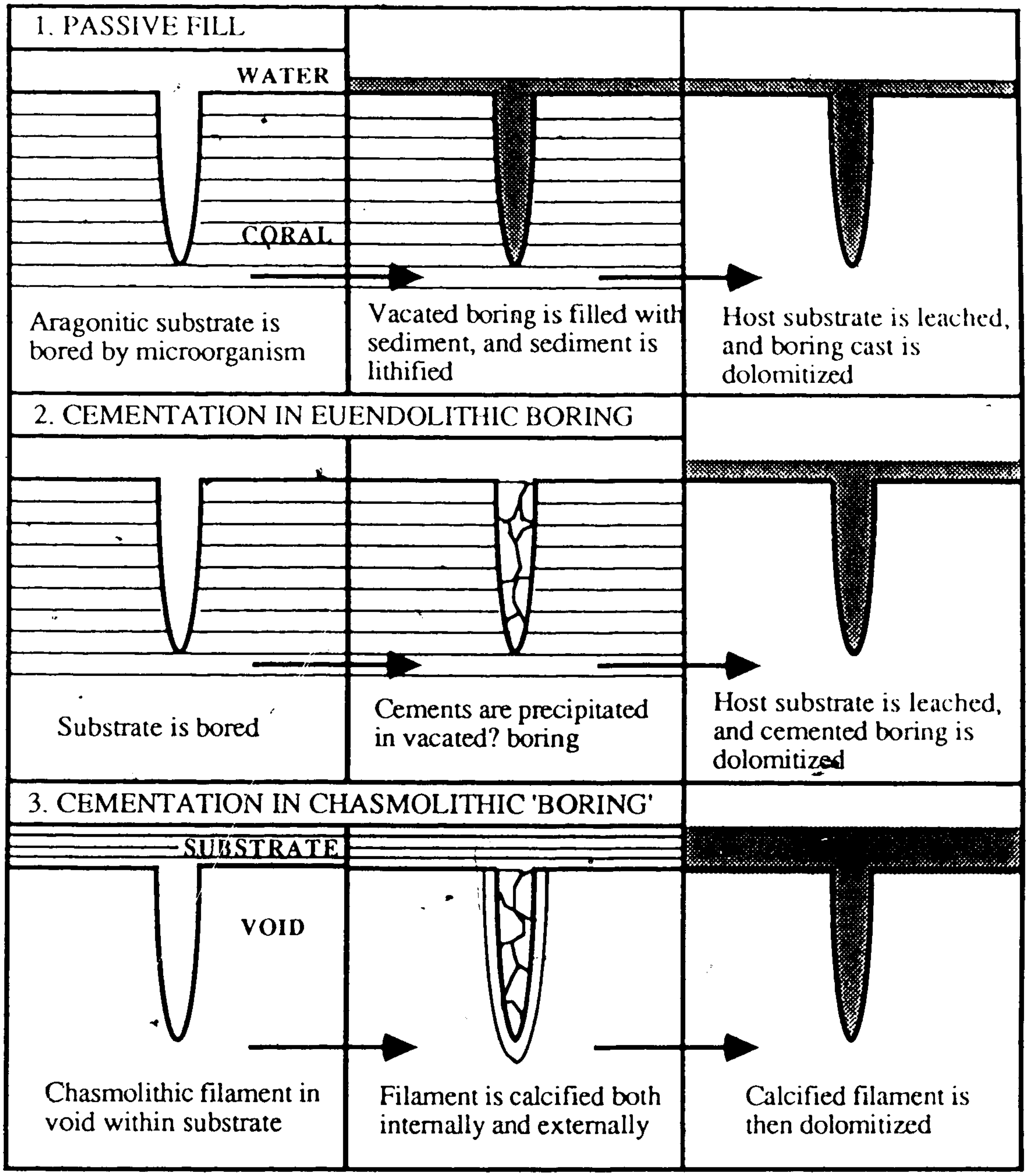


Figure 6. Three possible mechanisms for the preservation of the microborings.

1. Grain or skeleton bored by endoliths,
2. Occlusion (total or partial) of borings by aragonite or HMC by cementation or passive fill,
3. Lithification,
4. Dolomitization, and penecontemporaneous dissolution of aragonite, to reveal borings in cavities,
5. Precipitation of dolomite cement around some borings.

B. SYSTEMATIC ICHNOLOGY

Taxonomy in ichnology is frought with difficulties due to the special nature of trace fossils (Frey and Seilacher, 1985; Frey and Pemberton, 1985), and their uncertain relationships to biological taxa. Specifically this is because:

1. One organism can produce more than one ichnospecies of trace fossil.
2. Many unrelated species of organisms can produce the same trace fossil.
3. The ontogeny of the trace fossil does not always relate to that of the tracemaker.
4. An individual trace fossil can represent more than one behavioural trait.
5. Full taxonomic ranks cannot be applied to trace fossils. Taxa are presently restricted to ichnogenera and ichnospecies.

Like trace fossils in general, borings must be kept taxonomically distinct from biological taxa. Borings, however, tend to relate to their tracemakers to a greater degree than is the case with soft substrate trace fossils. This close relationship has served to confuse their taxonomy. This confusion is most apparent in the taxonomy of algae, fungi and bryozoa, and their corresponding borings. The problem arises from the recognition of borers in their borings in recent environments. In some cases, these organisms may not have even bored their dwellings, but have settled in pre-existing cavities excavated by other organisms. In other cases, the organism may have subsequently modified a pre-existing boring, thereby producing a xenomorphic boring. Some authors have attempted to name recent borings after the apparent borer (Pohowsky, 1974), thus confusing matters considerably when similar borings have been, or are consequently found in ancient settings.

In recent years, there has been a concerted effort to re-evaluate the taxonomy of borings, in order to provide a consistent and useful working classification (e.g. Voigt, 1975; Bromley and D'Alessandro, 1984; Kelly and Bromley, 1984).

Ichnogenus *Entobia* Bronn, 1837

(Plates 1A-G; 2A-G; 3A-G)

Selected Synonymy

1808 Bodies Parkinson.
1814 Cavities Conybeare.
1822 Parasitical bodies Mantell.
1837 *Entobia* Bronn.
1838 *Entobia* - Bronn.
1843 *Entobia* - Portlock.
1850 *Clionites* Morris (in Mantell).
1868 *Cliona* - Fischer.
1962 *Entobia* - Häntzschel.
1970 *Entobia* - Bromley.
1975 *Entobia* - Häntzschel.
1984 *Entobia* - Bromley and D'Alessandro.

Type ichnospecies: *Entobia cretacea* Portlock, 1843

Ichnospecies: The taxonomy of *Entobia* was recently reviewed and emended by Bromley and D'Alessandro (1984). Out of 23 existing ichnospecies of supposed sponge borings, they validated only 6: *cretacea* , *glomerata*, *megastoma*, *paradoxa*, *parisiensis* and *retiformis*. To these they added 6 new ichnospecies: *cateniformis*, *geometrica*, *laquea*, *mammillata*, *ovula* and *volzi*.

Stratifgraphic Range: Triassic to Recent.

Remarks: *Entobia* Bronn has been referred to the work of Cenozoic and Mesozoic boring sponges, generally of the family *Clionidae* (Morris, 1851; Fischer, 1868; Stephenson, 1952; Bromley, 1970). *Entobia* borings can also be produced by some species of spirastrellids, as well as clionids. For example, *Anthosigmella varians* can produce borings that are indistinguishable from those of *Cliona aprica* (Pang, 1973). Paleozoic borings that resemble those of recent boring sponges are not included in *Entobia*, as they need re-examination (Bromley and D'Alessandro, 1984). They therefore, retain their old names of*Clionolithes, Clionoides, Filuroda* and*Topsentopsis* (see de Laubenfels, 1955).

The diagnosis of *Entobia* (Bromley and D'Alessandro, 1984) includes those borings with single chambers and those with networks or boxworks of chambers. These are all considered to be the products of boring sponges. *E. glomerata* Morris is the only ichnospecies of *Entobia* that has single chambered borings. In the corals of the Bluff Formation there are numerous well preserved specimens of single chambered borings (of boring sponges), generally much larger and more ~~distinct~~ than the poorly described and "dubious" (Bromley, written comm., 1986) *E. glomerata*. Significantly, these single chambered borings occur in the same coral molds as entobians that form networks of chambers. They therefore represent a distinct behavioural trait of the trace maker, and as such, should be differentiated at the ichnogeneric level. This would avoid overcrowding *Entobia* and would allow for species distinction of this new single chambered form.

An allusion to the fate of clavate borings highlights the problem of using one ichnogenus to cover a wide range of boring morphologies. Problems of overcrowding and excessive lumping occurred when Bromley (1972) included a wide range of single entrance borings into the ichnogenus *Trypanites* Mägdefrau. This was not generally accepted, and led Kelly and Bromley (1984) to remove the clavate borings from *Trypanites*, and to reassign the names *Gastrochaenolites* and *Teredolites* to cover these clavate borings in lithic and lignic substrates respectively.

The establishment of a new ichnogenus for single chambered borings requires a modification of the *Entobia* diagnosis of Bromley and D'Alessandro (1984). *E. glomerata* can then be removed from *Entobia* and transferred to a new ichnogenus.

Emended diagnosis: Borings in carbonate substrates comprising networks or boxworks of galleries connected to the surface by several or numerous apertures. "Morphology changes markedly with ontogeny. The galleries show progressive increase in diameter during growth; in some forms, inflation at more or less regular distances produces a system of closely interconnected chambers; in other forms, chamber development is restricted to only a brief ontogenetic stage; in still other forms, no cameration is developed. The surface of the boring bears a cuspate microsculpture that may be lost in gerontic specimens. Fine apophyses arise from all or most surfaces of the system." (Bromley and D'Alessandro, 1984).

Entobia new ichnospecies A

(Plate 1, Figs. A-C)

Material: UA 7793, UA 7794, UA 7795.

Occurrence: Not abundant in the Bluff Formation. Occurs in stylophorid molds at CK and HRQ.

Diagnosis: Vaguely camerate network of branching galleries. Main axes of branches parallel substrate surface. Galleries maintain constant diameter, except at branch junctions, where they thicken slightly to accommodate new branches. Branching is less than 90° from main axis; network has a dendritic pattern. Lumps or buds are common on galleries. In areas where cameration is developed, chambers are rectangular to equidimensional.

Description: No growth stages are visible, but densest areas of the boring may represent oldest part. Network can consist of up to six tiers, and where densely packed may coalesce. The densely packed structure together with the complexly branching galleries, make it difficult to determine whether these structures represent one or many borings. Maximum boring depth of the best specimen is 11 mm. Galleries are 1 - 1.2 mm in diameter. At branch junctions diameter is up to 1.7 mm. A coating of cement (0.1mm) may account for the lack of apophyses expected on a sponge boring.

Remarks: The buds may represent new areas of growth, i.e. preliminary branches. This has not been seen in *Entobia* before and might suggest that this new ichnospecies is not a sponge boring. Phoronids can produce dense pseudocolonies, with a branching form resulting from asexual budding. The galleries of these are however, much finer than the new ichnospecies (about 150 μm in diameter) and are non-camerate (Marcus, 1949; Emig, 1979). Henderson and McNamara (1985) described *Gnathichnus* occurring as semi-positive reliefs on steinkern surfaces from the Maastrichtian of Western Australia. These consisted of ridges (30 to 250 μm in diameter) which branched and exhibited a faintly beaded texture. They interpreted these as imprints of phoronids, which had etched themselves into the shell. New ichnospecies A occurs well within the substrate and is thus a borer. It also has a characteristic dense, robust network of galleries, unlike the slender ridges with just two or three branches. The new ichnospecies is most similar to *E. megastoma* which is a non-camerate entobian. However, it differs by virtue of its dendritic branching habit, vaguely camerate nature of parts of the galleries and its common development of buds. It differs from *Talpina* sp. by the development of cameration and its large and robust character.

Entobia cateniformis Bromley and D'Alessandro, 1984

(Plates 1F, G; 4G)

1984 *Entobia cateniformis* - Bromley and D'Alessandro, p. 238, pl. 16, figs. 1, 3, 5; pl. 17, fig. 3; pl. 27, fig. 3.

Material: UA 7796, UA 7797, UA 7810, also many other well preserved specimens.

Occurrence: Occurs commonly in both coral molds and bivalve molds, at all localities.

Diagnosis: Entobian developed as an open camerate system, comprising long rows of cylindrical chambers forming an irregular maze through branching and anastomosis. Many chambers elongated, commonly T-, L- or cross-shaped where rows intersect. Elsewhere, no fusion of chambers. Intercameral canals reduced to constrictions. Apertures small, irregularly distributed; apertural canals well developed. Ontogenetic phases A, B and C well developed, D reduced.

Remarks: The highly distinctive T, L and cross shapes of the chambers of this ichnospecies are well developed. At least one specimen is xenomorphic after a *Trypanites* type boring (Plate 1G). Chambers are commonly 4 x 2 mm in growth phase C. Phase D has not been observed.

Entobia geometrica Bromley and D'Alessandro, 1984

(Plate 1D, E)

1984 *Entobia geometrica* - Bromley and D'Alessandro, p. 241, pl. 18, fig. 1; pl. 19, figs. 1, 3; pl. 20, fig. 1; pl. 21, figs. 1, 4, 5; pl. 22, figs. 1, 2, 5.

Material: UA 7805, also numerous other specimens.

Occurrence: Common in branching coral molds and *Agaricia* fragments, at all localities.

Diagnosis: Camerate entobian. Apertures of two markedly different sizes; the larger generally unfused, variable in diameter; the smaller spread among the larger ones. Apertural canals very short. Chambers wide, generally polygonal in section in mature growth-phases, separated by thin walls perforated by numerous short intercameral canals that may be fused together. In more juvenile phases the chambers are oval or subspherical, always joined by a distinct intercameral canal. Chambers disposed in weakly developed rows, tending to form a subrectangular network. The dominant growth phase is D; B and

C are usually much reduced. The system is developed parallel to the external substrate surface, in usually only one or two tiers.

Remarks: *E. geometrica* is characterized by polygonal-shaped chambers and the very short to absent intercameral canals connecting these. These chambers are up to 1 cm in diameter in some large specimens, but thickness and depth rarely exceeds 5 mm. Only one tier is commonly developed.

Entobia laquea Bromley and D'Alessandro, 1984

(Plates 2A, D, G; 3E; 4G)

1984 *Entobia laquea* - Bromley and D'Alessandro, p. 244, pl. 17, fig. 2; pl. 19, fig. 2; pl. 23, fig. 1.

Material: UA 7810, UA 7805 also many other specimens.

Occurrence: Very abundant in coral molds at all localities.

Diagnosis: A camerate entobian composed in mature stages of networks of small chambers arranged in several tiers subparallel to the substrate surface. The chambers, variable in shape, taper abruptly near the constrictions that separate them from neighbours. They are organized in short, more or less arcuate chains that encircle small spaces in a way that resembles lace. The apertures, circular in shape, rarely fused, are small, numerous and distributed irregularly. Phase A well represented by branched exploratory threads that anastomose early to produce a slender network. Furthermore, phase A is usually present at the periphery, even in mature specimens. Phase B is greatly reduced or absent, the enclosed meshes passing almost directly into phase C. The most characteristic growth phase of the ichnospecies is C.

Remarks: *E. laquea* is commonly associated with other entobians. Chambers are small and irregularly shaped (Plate 2A). The characteristic lacework development is common.

When restricted by the size of the substrate, the meshworks become dense and stenomorphic (Plate 4G), but may be tentatively assigned to *E. laquea*.

Entobia mammillata Bromley and D'Alessandro, 1984

(Plate 2B, C, E)

1984 *Entobia mammillata* Bromley and D'Alessandro, p. 246, pl. 20, fig. 3; pl. 24, fig. 1; pl. 25, figs. 1, 2, 4; pl. 26, fig. 2; pl. 27, fig. 3.

Material: UA 7798, UA 7799, UA 7800, also several other specimens.

Occurrence: Commonly associated with *E. megastoma*. Not abundant, occurs in branching coral molds at all localities.

Diagnosis: Camerate entobian in juvenile phases, non camerate in mature phases. Network of sublinear cylindrical chambers, later fusing to galleries, organized in an irregular maze of a single tier, anastomosed and branched at wide angles to a right angle. Surface ornamented with hemispherical tubercles, each bearing a long terminal apophysis, or by a cluster of partially fused tubercles. Phase C, well developed, phase D absent. Phase E represented by wide, shallow, surface grooves, the ornamentation locally lost. Apertures extremely variable, in two sizes; a few relatively large and wide, sparsely distributed; among these, minute apertures of much smaller size.

Remarks: This ichnospecies is characterized by the development of large cylindrical chambers covered with a tuberculate ornament (Plate 2C). More flattenend, ribbon-like examples of this are common (Plate 2E). Each hemispherical tubercle bears a distinct apophysis. These latter forms are found in small molds, having an average diameter of 1-2 cm, while the diameter of the boring is from 0.5-1 cm, and the thickness being 3-5 mm.

Entobia megastoma (Fischer, 1868)

(Plate 2C)

1866 *Cliona megastoma* Fischer, p. 198 (nomen nudum?)
1868 *Cliona megastoma* Fischer, p. 165, pl. 24, fig. 2, 2a.
?1932b *Cliona cretacica* Fenton and Fenton, p. 55, pl. 7, fig. 8, 9.
?1941 *Cliona microtuberum* Stephenson, p. 54, pl. 3, fig. 1, 2,4; pl. 5, fig. 1, 2.
1984 *Entobia megastoma* - Bromley and D'Alessandro, p. 250, pl. 23, fig. 3; pl. 24, figs. 2, 3; pl. 26, fig. 1; pl. 27, fig. 2.

Material: UA 7799, UA 7805.

Occurrence: Rare in Bluff Formation. Occurs in branching coral molds at HRQ.

Diagnosis: Non-camerate entobian, organized in an irregular boxwork system becoming more or less complexly intermeshed in the mature stages. The galleries are subcylindrical, frequently bifurcated, swollen at nodal points where, usually, several galleries conjoin. Apertures large and numerous, circular or oval in shape, rarely fused, disposed irregularly. Phase A reduced; phases B-D well developed.

Remarks: *E. megastoma* is a non-camerate boring that in the Bluff Formation reaches 1.5 cm in diameter. It branches regularly to form a dense network (Plate 2C).

Entobia ovula Bromley and D'Alessandro, 1984

(Plates 2B, F, G; 3A-E; 8A, B, D)

1984 *Entobia ovula* - Bromley and D'Alessandro, p. 254, pl. 17, figs. 1, 4; pl. 18, fig. 2; pl. 21, fig. 3; pl. 23, fig. 2; pl. 26, fig. 3; pl. 27, fig. 1; pl. 28, fig. 4; pl. 29, fig. 3.

Material: UA 7798, UA 7801, UA 7802, UA 7803, UA 7804, UA 7805, also many other well preserved specimens.

Occurrence: Very common in branching coral molds at all localities. Also occurs in *Montastrea* and *Diploria*, and in gastropod molds. Phases C and D never found together in same specimen.

Diagnosis: A camerate entobian composed in the mature stage of small chambers of globose to ovoid shape, greatly crowded, arranged in a boxwork. The chambers are separated from neighbours by a very short intercameral canal, usually reduced to a constriction. In phase C, the chambers are arranged in straight strings, forked at variable angles and anastomosed, giving rise to a network in one or two, poorly distinguishable tiers. A and B phases are reduced. Apertural canals distinct, tapering distally, or slightly inflated as a barrel. The openings are relatively small, numerous, rather regularly disposed, rarely fused.

Remarks: This ichnospecies is characterized by (1) boxworks of small spherical chambers, average diameter of 1-1.5 mm, and (2) open branching networks of spherical to cylindrical chambers, average diameter of 1mm.

The boxwork is up to 1cm in depth and commonly fills the coral mold (Plates 2F; 3E). This is the mature phase D of the boring (see discussion in Bromley and D'Alessandro, 1984). The chambers are closely packed, separated by short intercameral canals or constrictions. Each chamber may be connected to six or seven others, all maintaining distinct boundaries. According to Bromley and D'Alessandro (1984) phase D is rarely present in the same sample as phases A, B and C. Phase D has never been found in the same sample as phase C in the molds of the Bluff Formation.

E. ovula phase C consists of rows of chambers, which commonly branch and anastomise (Plate 3C, D). The chambers are separated from each other by constrictions rather than intercameral canals, and are more commonly cylindrical in shape rather than spherical. Both these features are inconsistent with the description of phase C given by Bromley and D'Alessandro (1984). They observed mostly spherical to subcylindrical chambers commonly separated by short intercameral canals. In addition, a more mature phase C has been observed in the Caymanian examples. Following the development of openly branching 'strings' of chambers, a more dense meshwork can develop (Plate 3D). The branching rows of chambers become increasingly dense with chambers filling in the spaces between the rows to produce a lacy meshwork pattern. The development into a boxwork (phase D) has not been observed. This tends to suggest that phase C and D may be different forms or even different ichnospecies, with their own individual ontogenetic developments. However, Bromley and D'Alessandro indicated that phase C may rarely develop into phase D. Thus, both are kept in this ichnospecies.

E. ovula can be xenomorphic, following older *Trypanites* borings (Plate 2G), stenomorphic following the coral structure (Plate 8D), or have a xenoglyphic texture (Plate 3A). The cuspate microsculpture typical of sponge borings is well preserved on this ichnospecies (Plate 3B).

Entobia paradoxa (Fischer, 1868)

(Plates 2B; 3E, G)

1868 *Cliona paradoxa* Fischer, p. 169, pl. 25, fig. 8.
1984 *Entobia paradoxa* - Bromley and D'Alessandro, p.259, pl. 20, fig. 2; pl. 26, fig. 4; pl. 29, figs. 1, 4.

Material: UA 7798, UA 7804, UA 7806, also several other specimens.

Occurrence: Relatively common in branching coral molds, especially at HRQ and CK. It is commonly associated with other entobians, *Trypanites* and small new ichnogenus A.

Diagnosis: A camerate entobian composed, in mature stages, of a network of very irregular chambers, somewhat amoeboid in shape, usually arranged in two tiers parallel to the substrate surface. Each chamber is connected to several others; the shape becomes extremely irregular owing to tapering as necks of varying lengths, before the constrictions that separate each chamber from its neighbours. In gerontic forms, a partial fusion among the chambers leads to the development of non-camerate galleries, usually variable in diameter and lacking diagnostic character. Apertures circular in shape, usually relatively small, uncrowded.

Remarks: This ichnospecies commonly has large well developed chambers (up to 1 cm in diameter) which are irregularly shaped (Plate 3E).

Entobia volzi Bromley and D'Alessandro, 1984

(Plate 3F)

1984 *Entobia volzi* - Bromley and D'Alessandro, p. 261, pl. 22, figs. 3, 4; pl. 25, fig. 3; pl. 28, figs.1, 2.

Material: UA 7805, also several other well preserved specimens.

Occurrence: Commonly fills small branching coral molds, especially abundant at HRQ.

Diagnosis: Diminutively camerate entobian consisting, in phase D, of chambers connected by wide intercameral canals or partially fused, taking a form resembling an irregular, close framework. This system is crossed in all directions by relatively wide, subcylindrical canals that connect with the substrate surface through large apertures. Growth front compact. Phases B and C are considerably reduced, characterized by appearance of irregular chambers or clusters of chambers as small swellings on the walls of the wide canals. Phase A comprises long, slender canals arranged irregularly and branched as a boxwork, having palmate expansions at nodal points. Apertures of two sizes, circular to oval, very irregularly distributed.

Remarks: The cylindrical boxwork of small spherical chambers develops from a non-camerate axial branch (Plate 3F). When this branch, or several branches are obscured by the chambers surrounding them, it can be confused with *E. ovula*.

New Ichnogenus A

(Plate 4A-G; Fig. 7)

1851 *Clionites* Morris.
1984 *Entobia* - Bromley and D'Alessandro.

Ichnospecies: New Ichnogenus A *glomerata*.

Material: UA 7807, UA 7808, UA 7809, UA 7810, also numerous other specimens.

Occurrence: Common in massive and branching coral molds of the Bluff Formation, especially at HRQ, CK, and less common at EEQ, BR, PCQ and QR.

Stratigraphic range: Cretaceous to Recent.

Diagnosis: Borings in carbonate substrates, comprising a single chamber, connected to substrate surface by an aperture, or one or more apertural canals. Progressive increase in size of chamber, and fusion of apertural canals may occur during ontogeny.

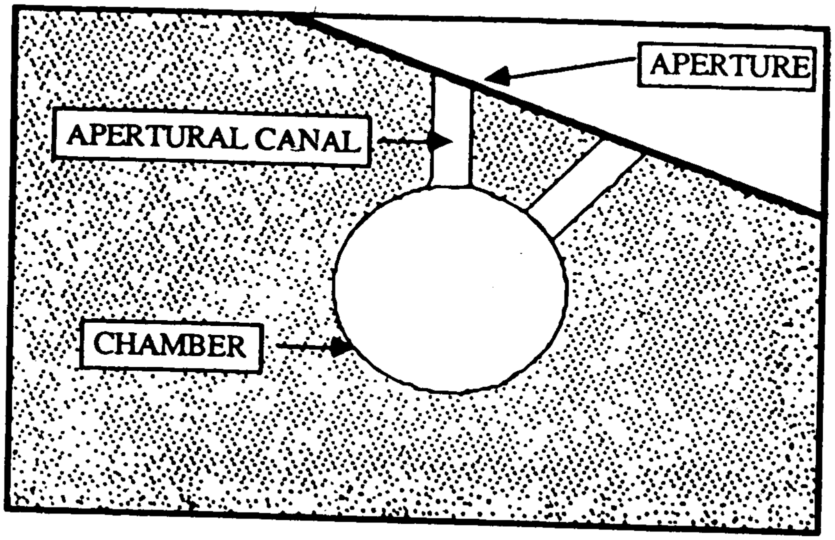


Figure 7. Terminology of New Ichnogenus A.

Remarks: New ichnogenus A may be comparable to the Palaeozoic ichnogenus *Topsentopsis* de Laubenfels, which is common in many Devonian stromatoporoids. *Topsentopsis* has a central cavity, about 10 mm in diameter, with numerous simple or branching tubes radiating from this. The apertural canals of new ichnogenus A are unbranched and extend from the chamber only in areas closest to the substrate surface. In accord with the arguments referring to Palaeozoic sponge-type borings, made by Bromley and D'Alessandro (1984), *Topsentopsis* will not be related to the new ichnogenus. It must await detailed re-examination of the type specimen. At present, *Topsentopsis* has a confused nomenclature and synonomy, and even its affinity to boring sponges is doubted (see Teichert, 1945, p. 200; Cameron, 1969b, p. 189; Häntzschel, 1975, p. W135).

Ichnogenus A differs from *Entobia* in being single chambered. It differs from *Gastrochaenolites* by the common development of more than one apertural canal. Those samples of ichnogenus A having only one apertural canal can be distinguished from *Gastrochaenolites* by the elongation of the chamber parallel to the substrate, rather than perpendicular to it (Table 1). Those samples of ichnogenus A, having only one apertural canal and a spherical chamber can be confused with *G. ornatus*: however, the latter has a neck that tapers from the main body chamber towards the substrate surface, whereas the apertural canal of ichnogenus A is generally cylindrical and distinct from the chamber. In addition, *G. ornatus* bears distinct bioglyphic structures on its base (Kelly and Bromley, 1984). Ichnogenus A is similar to recent borings of an Adociid sponge, *Siphonodictyon* and a clionid sponge, *Cliona amplicavata* (Table 2).

All boring sponges bore in a similar manner, producing a pitted microsculpture on the boring wall (Bergquist, 1978; Simpson, 1984). The negative form of this microsculpture has not been observed on the borings of the new ichnogenus because of the presence of either a calcite cement coating or xenoglyphic structures.

An ontogenetic sequence of the borings in the Bluff Formation is comparable with the sequence described for *Siphonodictyon coralliphagum* by Rützler (1971, fig. 9, p. 10).

Specimen	Chamber Dimensions mm	Canal Length mm	Canal Diameter mm	Depth of Boring mm	Number of Canals	Chamber Shape
UA7807	40 x 34 x 24	10 - 15	5 - 10	45	7	flattened ovoid
UA7808	18 x 12 x 15	10 - 13	2 - 3	36	4	spherical - irregular
UA7809	29 x 17 x 16	<11	1.5 - 3	27	4+	ovoid
UA7810	14 x 11 x 6	2	1.5	8	3+	flattened ovoid
1213a	33 x 42 x 16	<5	2 - 5	19	4+	very flattened ovoid
1213b	18 x 18 x 20	5	1.5 - 2	25	2	spherical
1213c	44 x 28 x 24	7	25 (fused mass)	29	at least 5 fused	ovoid
1213d	19 x 18 x 16	2 - 3	2	21	2 - 3	spherical
1213e	16 x 14 x 11	1	2	12	2	flattened spheroid

Table 1. Dimensions of New Ichnogenus A from the Bluff Formation.

Sponge Name	Chamber Dimensions mm	Canal Length mm	Canal Diameter mm	Depth of Boring mm	Number of Canals mm	Chamber Shape	Reference
Cliona amplicavata	5x16x12	1.1-4.8	0.8-2	<17	1-5	ovoid to equi-dimensional	Rützler, 1974
Siphonodictyon coralliphagum - obruta	5x5x5 to 20x20x20	5-10	0.5-1.8	25	1-2	spherical to ovoid	Rützler, 1971
S. coralliphagum - typica	10x20* 20-130 cm2	20-30	10-15	40	2*	ovoid to irregular	Rützler, 1971
S. coralliphagum - tubulosa	<2 litres	-----	-----	-----	-----	-----	Rützler, 1971
S. cachacrouense	500 cm3	-----	-----	60	-----	-----	Rützler, 1971
S. mucosum	250	-----	-----	-----	-----	spherical	Rützler, 1971
S. brevitulatum	10x10x10 to 50x50x50	3.8-9.8	narrow	60	-----	spherical	Pang, 1973

Table 2. Dimensions of Recent boring sponges comparable to New Ichnogenus A. (* indicates size measuerd from illustration)

He described four forms of *S. coralliphagum* that may represent growth phases. A single small chamber, commonly with two long apertural canals (tubes) increases in size, enveloping the apertural canals as it grows in all directions The apertural canals are thus shortened with increasing size of the chamber, reaching the final form of a single large chamber with one large indistinct aperture at the substrate surface. In the fossil molds of the Bluff Formation, new ichnogenus A with chambers that are about 1.5 cm in diameter, generally have the longest apertural canals, whereas larger chambers have relatively short canals or are indistinct blobs (Table 1). In contradiction to this sequence are some small ichnogenus A with 2 to 4 relatively short apertural canals compared to their chamber size of 5 to 10 mm (Table 1). These are also found in association with entobians in the branching coral molds (Plate 4F, G), while the larger ichnogenus A are commonly found in the massive coral molds associated with *Gastrochaenolites*. *Entobia glomerata* fits into this category of small chambered forms. These are analogous to the borings of *Cliona amplicavata* (Rützler, 1974), which are similar to those of *Siphonodictyon* except their chambers are smaller and their canals shorter (Table 2).

New Ichnogenus A *glomerata*

(Plate 4A-G)

1851 Clionites glomerata Morris, p. 89, pl. 4, figs. 11
1984 Entobia glomerata - Bromley and D'Alessandro, p. 234.

Material: As for ichnogenus.

Occurrence: As for ichnogenus.

Diagnosis: Single chamber may be spherical to irregularly elongate, with elongation parallel to substrate surface. One to seven apertural canals radiate out from chamber to connect to substrate surface. Chamber up to 5 cm maximum dimension.

Description: Chambers typically from 15 mm to 40 mm in diameter, with the apertural canals, up to 15 mm long and 3 mm to 5 mm in diameter (Table 1). Elongation ratios (length : thickness/width) of chambers less than 4 : 1. Smallest specimens have spherical forms; in many cases, deviations from this shape become increasingly pronounced during growth (Table 1). Xenoglyphic ornament representing the internal structure of the host coral common. Rarely two chambers may fuse.

The best specimen from the Caymanian examples (Plate 4A) is a slightly flattened ovoid with the dimensions 4 x 3.4 x 2.4 cm, with flattening and elongation normal to substrate. It has 7 pronounced apertural canals, 1-1.5 cm in length and 0.5-1 cm in diameter, radiating out from chamber in areas closest to substrate surface (Plate 4A). A number of small apophyses, averaging 2 mm in diameter and 5 mm in length, are present on chamber. A xenoglyphic sculpture covers the boring.

Remarks: New ichnogenus A *glomerata* differs from all ichnospecies of *Entobia* in being single chambered. *Siphonodictyon* and *Cliona amplicavata* produce similar borings in the Recent.

Ichnogenus *Gastrochaenolites* Leymerie, 1842

(Plate 5A-E)

1842 *Gastrochaenolites* Leymerie.
1972 *Trypanites* - Bromley.
1976 *Paleolithophaga* Chiplonkar and Ghare.
1980 *Teredolites* - Bradshaw.
1980 *Gastrochaenolites* - Kelly.
1980 *Teredolites* - Kelly (in Balson, 1980).
1984 *Gastrochaenolites* - Kelly and Bromley.

Type ichnospecies: *Gastrochaenolites lapidicus* Kelly and Bromley, 1984

Ichnospecies : *Gastrochaenolites* includes the ichnospecies *ampullatus, cluniformis, dijugus, lapidicus , orbicularis, ornatus, torpedo, turbinatus* and *longissimus* (Kelly and Bromley, 1984).

Stratigraphic range: Jurassic to Recent

Diagnosis. Clavate borings in lithic substrates. The apertural region of the boring is narrower than the main chamber and may be circular, oval, or dumb bell shaped. The aperture may be separated from the main chamber by a neck region which in some cases may be widely flared. The main chamber may vary from subspherical to elongate, having a parabolic to rounded truncated base and a circular to oval cross section, modified in some forms by a longitudinal ridge or grooves to produce an almond- or heart-shaped section.

Remarks: These are clavate borings in lithic substrates, and may be formed by bivalves, gastropods or worms.

Gastrochaenolites torpedo Kelly and Bromley, 1984

(Plate 5B-E)

1984 *Gastrochaenolites torpedo* -Kelly and Bromley, p. 802, figs. 3I, 8a, b.

Material: UA 7812, UA 7813.

Occurrence: Occurs in completely leached molds of massive corals. Locally abundant at HRQ and QR. Less common at other localities.

Diagnosis: Elongate smooth boring, widest point close to mid-line with the base acutely parabolic. The neck region is markedly compressed but the aperture itself is oval or approaches a figure-of-eight shape.

Remarks: This is an elongate clavate boring, with the widest point in the middle of the boring (Plate 5B-D). The neck is compressed in length and thickness, and the aperture is oval. *G. torpedo* is commonly produced by *Gastrochaena* and *Lithophaga* in the Recent.

Gastrochaenolites turbinatus Kelly and Bromley, 1984

(Plate 5A)

1984 *Gastrochaenolites turbinatus* Kelly and Bromley, p. 803, figs. 3g, 8c

Material: UA 7811.

Occurrence: Not abundant and occurs in both branching and massive coral molds, at HRQ and EEQ.

Diagnosis: Smooth *Gastrochaenolites*, acutely conical, having evenly tapered body and neck, the widest point close to the short rounded base, rounded cross section throughout length

Remarks: This is a smooth, conical, evenly tapered boring, with the widest point close to the rounded base. It maintains a rounded cross section throughout its length, although the aperture may be slightly ovate. A typical specimen is 45 mm in length and 3.5 mm-5 mm in diameter. The body is commonly curved along its length. Evans (1970) described borings of *Penitella* that resemble this ichnospecies.

New Ichnogenus B

(Plate 5E-H; Fig. 8)

Ichnospecies: New Ichnogenus B new ichnospecies A.

Material: UA 7814, UA 7815, UA 7816, also several other specimens.

Occurrence: This ichnospecies is common in completely leached massive coral molds at all localities except BH, and is associated with new ichnogenus A, *Gastrochaenolites* and *Trypanites*. The larger borings of this ichnospecies are the most common, i.e. those with an aperture diameter of 1.5 to 2 cm and a length of 6 to 7 cm.

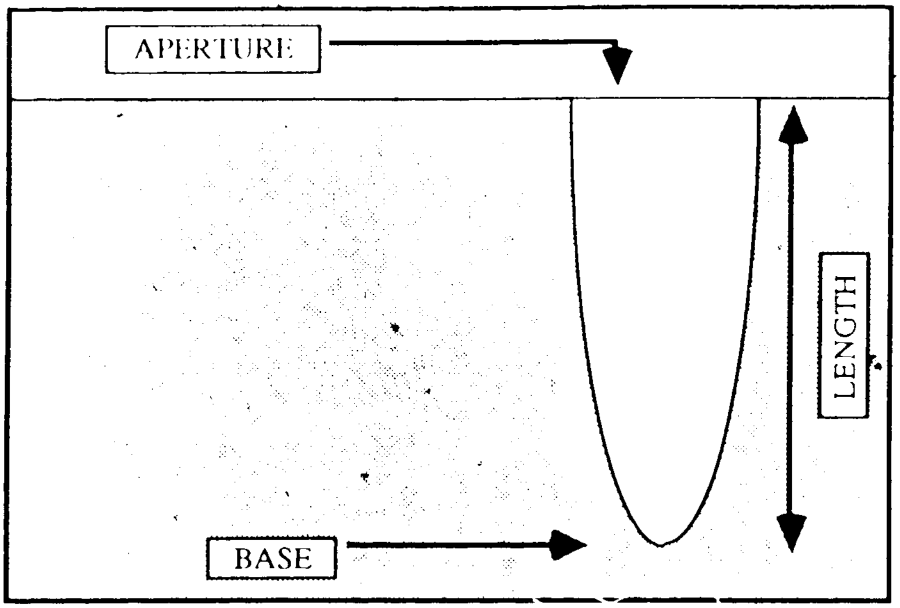


Figure 8. Terminology of New Ichnogenus B

Stratigraphic Range: Tertiary to Recent.

Diagnosis: Single entrance boring in lithic substates, with apertural region wider than middle and distal portions. Boring may be straight or curved about long axis. Apertures are slightly ovate to circular.

Remarks: New ichnogenus B tapers from its aperture to its distal end, and this serves to differentiate it from clavate *Gastrochaenolites* and the cylindrical *Trypanites*. The affinities of the new ichnogenus are unknown. The borings resemble *Lithotrya* 'borings' illustrated by James (1970). This barnacle does not bore, but allows the coral to envelope it with hard material. It is thus an embedment structure. Bromley (1978), however, noted that James had incorrectly identified this embedder, suggesting that it was in fact *Ceratoconcha domingensis*. These taper distally and commonly have longitudinal ribs on the 'boring wall'. In view of the fact that many embedders can also enlarge their crypts, they are treated as borings for taxonomic purposes (Ekdale *et al.*, 1984). *Lithotrya* physically bores into hard substrates producing cylindrical borings, with oval apertures up to 8 mm in diameter and lengths up to 7 cm (Ahr and Stanton, 1972). These may also have produced some new ichnogenus B.

New Ichnogenus B new ichnospecies A

(Plate 5E-H)

Material: As for ichnogenus.

Occurrence: As for ichnogenus.

Diagnosis: Single entrance boring which tapers evenly and gradually from aperture to base. Boring is straight or only slightly curved along length.

Desoription: Axis of boring straight or only slightly curved, with apertural diameter of 4 to 21 mm. Length ranges from 7 to 78 mm (Table 3). Surface of boring smooth or with xenoglyphic sculpture. Ichnogenus B generally occurs perpendicular to the substrate surface. The largest specimen found is 12 mm in diameter and 78 mm in length (Plate 5G). Aperture is slightly ovate. Boring is straight and covered with xenoglyphic ornament.

Remarks: Distinguished from *Gastrochaenolites torpedo* by its even tapering from the widest point, the aperture, to the base, and by its slightly ovate to round aperture. *G. torpedo* is a clavate boring with and oval or figure-of-eight aperture.

Ichnogenus *Trypanites* Mägdefrau, 1932

(Plates 1B; 2D, G; 5A; 6A-C, E; F-H)

1840 *Talpina* von Hagenow.
1932 *Trypanites* Mägdefrau.
1937 *Nygmites* - Mägdefrau.
1952 *Specus* Stephenson.
1967 *Vermiforichnus* Cameron.
1968 *Conchifora* - Müller.
1972 *Trypanites* - Bromley. -
1977 *Trypanites* - Kobluk *et al.*
1977 *Spiracavities* Chiplonkar and Ghare.

Specimen	Length mm	Diameter of Aperture mm	Shape of Aperture	Diameter of Base mm	Remarks
UA7814	55	15 x 15	circular	4 x 3	slightly curved along length
UA7815	18	10 x 10	circular	3 x 3	short, stout
UA7816	78	12 x 16	slightly ovate	3 x 3	straight, evenly tapered
1196a	32	13 x 12	ovate to triangular	4 x 3	slightly triangular in cross section
1196b	41	12 x 14	sub-rectangular	5 x 3	maintains rectangular x-section to base
1196c	20	8 x 8	circular	4 x 4	slightly curved along length
1196d	18	11 x 12	circular	3 x 3	tapering very pronounced
1196e	7	4 x 4	circular	2 x 2	tapers gently
1196f	12	7 x 5	ovate	3 x 3	short, stout

Table 3. Dimensions of New Ichnogenus B from the Bluff Formation.

1980 *Trypanites* - Pemberton *et al*
1982 *Cylindrocavities* Ghare.
1987 *Trypanites* - Pemberton *et al*

Type ichnospecies: *Trypanites weisei* Mägdefrau, 1932

Ichnospecies: *Trypanites* Mägdefrau consists of three ichnospecies, *fimbriatus*, *solitarius* and *weisei* .

Stratigraphic Range: Cambrian to Recent.

Diagnosis: Simple unbranched vertical to sinuous borings with a single opening to the surface; with or without a flared entrance. Circular in cross-section; generally isodiametric throughout the entire length but may terminate in a small cupulate chamber.

Remarks: The borings are generally isodiametric, have a circular cross section and are single entranced (see Pemberton *et al.*, in press). They are slender and elongate and commonly less than 2 mm in diameter. The borings may be produced by spionid polychaetes, sipunculids, phoronids and thoraciacan cirripedes (Cameron, 1969a; Bromley; 1970; Pemberton *et al.*, 1980).

Kelly and Bromley (1984) removed *Gastrochaenolites* and *Teredolites* from *Trypanites*, restricting these to clavate or club shaped borings in lithic and lignic substrates respectively. These are generally produced by bivalves and are thus larger than *Trypanites*. Kelly and Bromley (1984) acknowledged that slender forms similar in shape to *G. torpedo* and *G. turbinatus* existed. These slender forms resemble borings produced by sipunculids and polychaetes in recent environments. Kelly and Bromley (1984) did not make it clear if these slender forms should be assigned to *Gastrochaenolites*. These slender forms are obviously gradational between *Trypanites* and *Gastrochaenolites*, being similar in size and dimensions to the former, but similar in shape to the latter.

These slender forms, similar to *G. torpedo* and *G. turbinatus*, are present in the Bluff Formation. Additionally, there are slender borings which taper distally (comparable to the knitting needle borings of Palmer and Palmer, 1977), which are comparable in shape to new ichnogenus B, but in size to *Trypanites*. Pemberton *et al.* (in press) recognised that some *Trypanites* may taper slightly towards the base, may have a cupulate chamber at the base (*T. frimbiatus*), or may flare at the aperture. Thus, *Trypanites* is not strictly isodiametric. In view of the apparently transitional borings between *Trypanites* and *Gastrochaenolites*, and between *Trypanites* and new ichnogenus B, a boundary must be drawn.

The most practical division is to place all the slender borings, less than about 2 mm in diameter, into *Trypanites*. The slender borings are generally cylindrical and do not have pronounced morphologies (Fig. 9). This also overcomes any problems encountered when viewing these slender borings in thin section. Commonly, due to the orientation of the thin section, the borings appear to taper, even though they are really isodiametric. Those borings larger than 2 mm and having morphologies consistent with *Gastrochaenolites* or new ichnogenus B, are assigned to these ichnogenera. This distinction is consistent with the phyletic relationships of the ichnogenera. *Gastrochaenolites* and new ichnogenus B are probably produced mainly by bivalves (gastropods and ?barnacles), whereas *Trypanites* is produced mainly by worms. This scheme is employed in classifying the Caymanian borings.

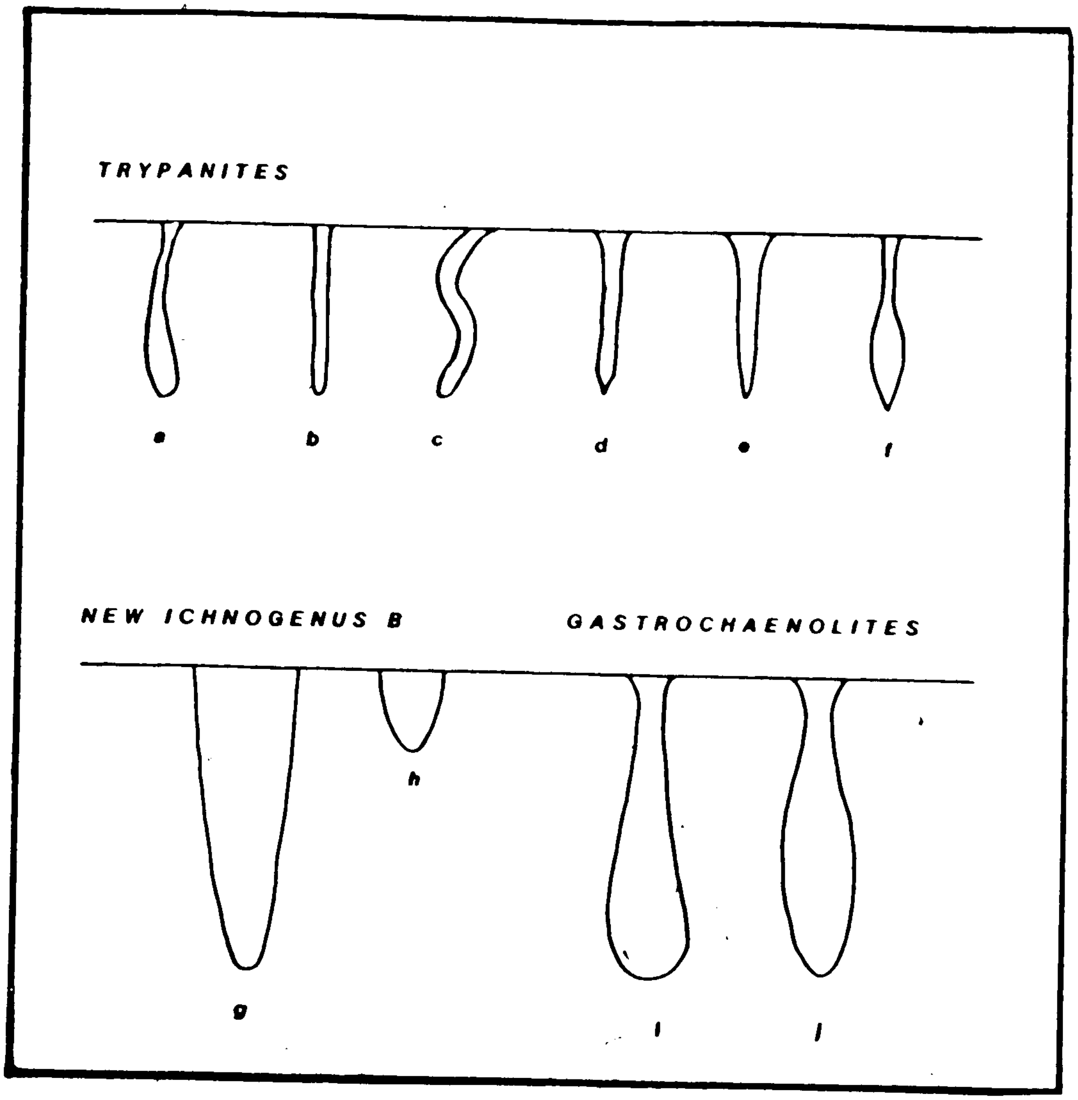
Trypanites solitarius **(von Hagenow, 1840)**

(Plates 5A; 6A, C, E)

1840 *Talpina solitaria* von Hagenow, p. 671.
1849 *Talpina solitaria* - Quendstedt, p. 470, pl. 30, fig. 37.
1937 *Nygmites solitarius* Mägdefrau, p. 56.
1972 *Trypanites solitarius*-Bromley, p. 96, fig. 1h.

Material: UA 7805, UA 7811, UA 7817, also many other specimens.

Occurrence: These are common in both the massive and branching corals at all localities.



Trypanites

a - frimbiatus
b - weisei
c - solitarius
d - weisei
e - sp.
f - sp.

New ichnogenus B

g - new ichnosp. A
h - new ichnosp. A

Gastrochaenolites

i - turbinatus
j - torpedo

Figure 9. A comparison of *Trypanites*, *Gastrochaenolites* and new ichnogenus B morphologies.

Diagnosis: Simple, more or less cylindrical Trypanites with straight or gently curving course generally following close beneath the substrate surface. Diameter ca. 1 mm, length up to ca. 3 cm.

Remarks: The isodiametric (average 2 mm diameter) borings are sinuous, commonly meandering from wall to wall of the skeletal cavities.

Trypanites weisei **Mägdefrau, 1932**

(Plates 2D, G; 6C, E)

1932 *Trypanites weisei* Mägdefrau, p. 152.
1987 *Trypanites weisei* - Pemberton *et al.* in press.

Material: UA 7805, UA 7810, also many other specimens.

Occurrence: This ichnospecies is less common than *T. solitarius*, although it is commonly associated with it and also occurs at all localities.

Diagnosis: Simple, vertical, more or less straight borings with a single opening to the surface; circular in cross-section and isodiametric throughout their length.

Remarks: This boring is cylindrical and straight, and ranges in diameter from 1 to 2 mm.

Trypanites **sp.**

(Plate 6B)

Material: UA 7818, also several other specimens.

Occurrence: Occurs in branching and massive corals, associated with other *Trypanites* ichnospecies at HRQ.

Remarks: Slender borings that are torpedo shaped, comparable to recent sipunculid borings. These taper towards the aperture, and towards the base. On the basis of their slender morphologies (less than 2 mm in diameter) they are included in *Trypanites* (Plate 6B). Also many other *Trypanites* that cannot be assigned to an ichnospecies due to their incomplete borings or slightly variable shapes.

Ichnogenus *Maeandropolydora* Voigt, 1965

(Plate 6D, F, H)

1965 *Maeandropolydora* Voigt.
1983 *Maeandropolydora* - Bromley and D'Alessandro.

Type ichnospecies: *Maeandropolydora decipiens*.

Ichnospecies: *Maeandropolydora* has three ichnospecies: *decipiens* , *elegans* and *sulcans*.

Stratigraphic Range: Cretaceous to Recent.

Diagnosis: Long cylindrical galleries having two or more apertures, running through the substrate sinuously or in irregular contortions. Galleries may run parallel in contact with each other in pairs, with or wothout fusion. Loose or tight loops may occur; the limbs of these may be connected by a vane or form a pouch.

Remarks: These are long cylindrical galleries, with two or more apertures. They are intermediate forms between *Trypanites* and *Caulostrepsis* (Bromley and D'Alessandro, 1983) and thus presumably also formed by worms.

Maeandropolydora cf. *sulcans* Voigt, 1965

(Plate 6D, F, H)

1901 *Dodekaceria* (?) sp. Rovereto, p. 228, pl. 28, fig. 1, 4c.
1965 *Maeandropolydora sulcans* Voigt, p. 204, pl. 26, fig. 5, 6; pl. 27, fig. 3, 5 (?).

Material: UA 7819, UA 7821, also several other specimens.

Occurrence: Common in branching coral molds, especially at HRQ.

Diagnosis: Cylindrical gallery having at least two apertures, irregularly contoted, commonly bent in loops, never showing fusion where walls are in mutual contact; vane absent.

Remarks: In the Bluff Formation, U-shaped borings are present. The borings have a constant diameter of 1-2 mm, but they are broken or obscured by other borings. Loops, however, are well developed, sometimes running in pairs and then diverging in opposite directions. These are therefore incomplete borings of *Maeandropolydora*, and can be tentatively assigned to *M. sulcans*, as no vanes are present, and the galleries do not fuse.

Ichnogenus *Talpina* von Hagenow, 1840

(Plate 6G)

Selected Synonymy:

1840 *Talpina* von Hagenow.
1922 *Bascomella* Morningstar.
1944 *Heteronema* Condra and Elias
1945 *Conchotrema* Teichert.
1952 *Graysonia* Stephenson.
1961 *Graysonia* - Casey

Type Ichnospecies: *Talpina ramosa.*

Ichnospecies: *Talpina* now has five ichnospecies: *annulata*, *eduliformis*, *gruberi*, *hirsuta* and *ramosa* (Voigt, 1975).

Stratigraphic Range: Devonian to Recent.

Remarks; *Talpina* has been attributed to the work of annelids, sponges and bryozoa (e.g. Morris, 1851; Bromley, 1970), but Voigt (1972; 1975) referred it to the phylum Phoronidea. Phoronids are lophophorate worms that commonly bore, producing pseudocolonies (Voigt, 1972).

In the Bluff Formation, delicately branching networks of galleries are present (Plate 6G, UA 7820). These have numerous apertures connecting to the substrate surface. The galleries are on average 200 µm in diameter and are comparable to *T. eduliformis*.

Ichnogenus *Caulostrepsis* Clarke, 1908

(Plate 7A, C)

Selected Synonymy:

1908 *Polydorites* Douvillé.
1908 *Caulostrepsis* Clarke.
1971 *Dodecaceria* Voigt.
1973 *Ramosulcichnus* Hillmer and Schulz.
1983 *Caulostrepsis* - Bromley and D'Alessandro.
Also numerous *Polydora* borings.

Type Ichnospecies: *Caulostrepsis taeniola* Clarke, 1908.

Ichnospecies: *Caulostrepsis* has six ichnospecies: *avipes*, *biforans*, *contorta*, *cretacea*, *taeniola* and ichnospecies.1 (Bromley and D'Alessandro, 1983).

Stratigraphic Range: Devonian to Recent.

Diagnosis: Single-entrance borings or embedment structures having a pouch shape or ear shape produced by a gallery bent in a U. More complex structures can be produced by development of multiple lobes on the same basic U-plan. The limbs may be clearly visible throughout their length and connected by a vane, or they may be fused to produce an oval or flattened pouch lacking a vane. All intermediate states, involving an axial depression, occur. At the distal end the width is at least double the thickness; the cross sectional area shape here varies from flat-oval, elliptical or constricted to dumbbell-shaped. At the apertural end the width is normally noticeably less than at the distal end, but the shape of the section may be more or less the same, or subcircular. In some cases symmetrical rows of deep pits may be developed towards the apertural end. The aperture itself may have the same form as the proximal cross section, or it may be modified by the development of superficial branches or apertural grooves, normally 2 to 4 in number, radiating out from it.

Remarks: These are single entrance borings, with a pouch formed by a modified U-shaped gallery. They are distinguished from *Trypanites* by their flattened cross sections.

Caulostrepsis cf. *cretacea* Voigt, 1971

(Plate 7A, C)

1970 *Dodecaceria* (?) sp. Voigt, p. 375, pl. 6, fig. 2-5.
1971 *Dodecaceria cretacea* Voigt, p. 150, pl. 15, fig. 12; pl. 16, fig. 1-9.
1972 *Trypanites cretacea* - Bromley, p. 96, fig. 1i.
1983 *Caulostrepsis cretacea* - Bromley and D'Alessandro, p. 291, pl. 21, figs. 1, 3; pl. 22.

Material: None.

Occurrence: Rare and occurs at QR and HRQ in massive coral molds.

Diagnosis: Galleries bent in a long, narrow U-form with the inward-facing walls of the limbs fused by complete removal; the original position of the median wall is sometimes indicated by a very shallow axial depression along the structure. Vane absent. Transverse section always flattened-elliptical but showing gradual decrease in width toward the aperture. Shape of aperture flattened-oval.

Remarks: Average dimensions are 6 mm in width and 8 cm in length, but are flattened, being only about 2 mm in thickness. A shallow axial depression is just visible. The borings curves along their axis, but there is no axial tortion. These are comparable to *C. cretacea*, although somewhat longer than typical specimens.

Problematic Ichnotaxa

(Plate 7B, D, E)

Material: UA 7822, UA 7823.

Occurrence: Common in massive coral molds especially, but can occur in larger branching coral ~~molds~~. Present at HRQ, QR, EEQ, PCQ.

Remarks: Large tunnel-like borings are common in the coral molds. These generally run parallel to the substrate surface, and branch at right angles to connect with the surface. The diameter of the main tunnel in the best specimen is 1-1.5 cm, and the length is over 7 cm. The boring is incomplete as the tunnel is broken off at this length. The 'apertural canals' that run perpendicular to the main tunnel are 4 mm in diameter. The tunnels are cylindrical except where two run along side and partially fuse. Only segments of this boring have been found, and these are commonly associated with new ichnogenus B.

These are comparable to borings of the Spirastrellid sponge, *Spheciospongia*. This produces a very large non-camerate network composed of tunnels commonly running parallel to, and just below, the substrate surface. It can penetrate up to 12-15 cm into the substrate and commonly branches (MacGeachy and Stearn, 1976; Bromley, 1978).

Microborings

(Plates 11E, F; 7F, G)

Microorganisms, like macroorganisms, cannot readily be identified on the basis of their borings. Many microorganisms produce borings that are too simple or too variable to allow precise identification of the trace maker (Golubic *et al.*, 1975). In addition, unrelated species may produce morphologically similar borings. However, there are some species that produce highly characteristic borings, and it is on this basis that present studies concentrate on comparing fossil with recent borings, in the hope that the latter are diagnostic of biological taxa (Ekdale *et al.*, 1984). On account of this, no good ichnotaxa have been erected. More promising advances have been made in the recognition of distinct boring patterns that change with depth, climate and/or geographic location (e.g. Rooney and Perkins, 1972; Edwards and Perkins, 1974; Golubic *et al.*, 1975; Budd and Perkins, 1980). The erection of ichnotaxa based on these boring patterns, without trying to relate them to biotaxa, would provide a useful ichnological nomenclature, and would be consistent with normal ichnological taxonomy procedures. This would also provide a meaningful palaeoecological framework.

Even attempts at differentiating between borings of fungi and those of algae are equivocal (Bromley, 1970; Klappa, 1979, Table 1, p. 958). Algae are not restricted to the photic zone as commonly believed. Like fungi, some algae are heterotrophic, deriving carbon and energy by degrading organic compounds, not by photosynthesis. Algae have been found in dark caves (Høeg, 1946; Palik, 1960; Jones, 1965; Nagy, 1965; Mason-Williams, 1967; Dragonov, 1977; Jones and Motyka, in press) and in deep waters (Riding, 1975). Even photosynthesizing algae can occur in deep waters, as the photic zone may vary considerably in thickness depending on the clarity of the water etc. (Ekdale *et al.*, 1984). Both fungi and algae can occur in subaerial, freshwater and marine environments, from temperate regions to the tropics (Schroeder, 1972).

The microborings of the Bluff Formation may be both algae and fungi, and probably bacteria. The diameter of the borings are from 2 μm to 300 μm. Some of these larger ones may even be bryozoan, worm or sponge borings. Their branching characteristics are inconsistent and commonly vary over the length of the boring. Therefore the criteria of Bromley (1970) cannot be applied to these with any certainty. Although earlier comparison was made to the green alga, *Ostreobium*, (average diameter of 2-5 μm), other algae can produce similar boring patterns e.g. the red alga, *Porphyra* (Golubic *et al.*, 1975). Therefore, the microborings will be assigned no biological affinities.

C. SIGNIFICANCE OF THE BLUFF FORMATION BORINGS

Amount of Boring

Percentage total volume bored of the corals of the Bluff Formation ranges from 0 to 75% (visual estimation). On the basis of counts made on 10 rock slabs, an average of 38% bored is obtained (Table 4). This is in reasonable agreement with visual observations. The range of percentage coral bored is apparent in adjacent molds and in any one mold. This is especially apparent in the branching coral molds, where one end of the branch, or one branch of a colony, contains no borings, and the other areas of the same mold are heavily bored.

Entobian borings are responsible for about 75% of the total bored areas, *Trypanites* about 7%, *Gastrochaenolites* 1%, ichnogenus A 8%, ichnogenus B 4%, *Maeandropolydora* <1%, *Talpina* <1% and *Caulostrepsis* <1% (Table 5). The large tunnel-like borings represent about 3%, and it is indeterminable how significant the microborings are.

The boring assemblage (Table 5) is dominated by the ichnogenus *Entobia*, which is represented by 9 ichnospecies. *Trypanites* sp. and new ichnogenus A are locally abundant. The most abundant ichnospecies incude *E. ovula* phases C and D, *E. laquea*, and *T. solitarius*.

The branching Stylophorid coral molds (average 2 x 8 cm) are the most common and most extensively bored skeletons and contain all the ichnospecies of *Entobia*, *Trypanites*, *Maeandropolydora*, new ichnogenus A and *Talpina* that occur in the Bluff Formation (Table 6). In any one mold, there may be from one to five other ichnospecies. Where the borings are crowded in a small space, they commonly become stenomorphic.

The massive corals range in preservation from completely hollow, structureless molds, to molds with a high degree of structure preservation due to 'inverse porosity'. Ichnogenus A, ichnogenus B, *Gastrochaenolites*, *Trypanites* and *Caulostrepsis* are found in the hollow molds (average size 20 x 20cm). Unlike the molds of *Stylophora*, these do

SLAB	CORAL AREA cm^2	% CORAL BORED
1	14	31
2	5	73
3	14	25
4	6	21
5	4	61
6	3.5	49
7	3	25
8	3.5	64
9	2.5	21
10	1	14
average	▨	38

Table 4. Percentage boring in 10 corals from the Bluff Formation. Rocks were impregnated with fibreglassing resin and slabbed at 1.5 cm intervals. Slabs cutting through the middle of coral molds were chosen for quantification.

ICHNOGENERA	ICHNOSPECIES	AFFINITIES	% OF TOTAL BORINGS
Entobia	*cateniformis*	Clionids and Spirastrellid sponges	75
	geometrica		
	laquea		
	mammillata		
	megastoma		
	ovula		
	paradoxa		
	volzi		
	new ichnosp. A		
New ichnogenus A	new ichnosp. A	Adociid & Clionid sponges	8
Gastrochaenolites	*torpedo*	Bivalves, Gastropods	1
	turbinatus		
New ichnogenus B	new ichnosp. A	Molluscs?, Barnacles?	4
Trypanites	*solitarius*	Annelids, Cirrepedes, Phoronids	7
	weisei		
	sp.	Sipunculids	
Maeandropolydora	cf. *sulcans*	Polychaetes	<1
Talpina	sp.	Phoronids, bryozoa?	<1
Caulostrepsis	cf. *cretacea*	Polychaetes	<1
Problematica	----------------	Spirastrellid sponges	3

Table 5. Ichnogenera and ichnospecies present in the Bluff Formation and their relative importance as indicated by percentage of the total borings. The microborings are not included on the table.

CORAL TYPE	ICHNOTAXA PRESENT
Stylophora sp.	*E. cateniformis, E. geometrica,* *E. laquea, E. mammillata, E. megastoma, E. ovula, E. paradoxa,* *E. volzi,* E. new ichnospecies A, *T. solitarius, T. weisei, T.* sp. *G. turbinatus,* *M.* cf. *sulcans,* *Talpina* sp. New ichnogenus A.
Completely leached massive coral molds	New ichnogenus A New ichnogenus B, *T. solitarius, T. weisei, T.* sp. *G. torpedo,* *Caulostrepsis* cf. *cretacea*
Montastrea sp.	*E. ovula,* *Talpina* sp.
Diploria sp.	*E. ovula*
Dichocoenia sp.	*E. ovula*
Agaricia sp.	*E. geometrica, E. laquea*
Solitary corals	none

Table 6. Ichnotaxa present in the most common coral types of the Bluff Formation.

not preserve the negative impressions of the corallites and the xenoglyphic sculptures on the borings are not distinct enough to aid in the recognition of the coral. Clusters of ichnogenus A and ichnogenus B, and the tunnel-like borings, are common in these molds (Plates 4C; 5E; 7E). Entobian borings are absent, possibly as a result of destruction by subsequent generations of borings.

Entobian borings and *Talpina* sp. are present in the massive corals that exhibit preservation as a result of 'inverse porosity'. *E. geometrica* and *E. laquea* are commonly the only macroboring found in *Agaricia* sp. *E.ovula* phase C is common in *Diploria* sp., *Talpina* and *E.ovula* phase C are present in *Montastrea* sp. and *E. ovula* phase C in *Dichocoenia* sp. These corals are characterized by an absence of ichnogenus A, ichnogenus B, *Caulostrepsis* and *Gastrochaenolites* (Table 6).

No borings have been observed in the solitary corals (which have a high degree of 'inverse' preservation).

E. ovula and *E. cateniformis* occur on the bivalve casts, while only the former has been found in the gastropods (Plate 8A, B). *Entobia* sp. and possibly *Trypanites* sp? occur in the coralline algal crusts.

Microborings are common in the branching and massive corals, and the coralline algal crusts, but due to the difficulty of preserving these during thin section preparation and the consequent restriction of their study to the SEM, their abundance cannot be related to that of the macroborings.

The borings are not distributed evenly in geographic terms. Borings of all types are common and abundant at localities HRQ, EEQ, BR (Fig. 1B). All the boring types have been observed at localities QR and PCQ, but they are less abundant than at the former three localities. Entobians, small chambered forms of new ichnogenus A, *Trypanites* and *Maeandropolydora* are present at CK, the former (entobians) being especially abundant. The entobians found at CK are also generally larger and more robust forms than at the other localities. Only entobians were found at PBQ, but this is most probably a result of very limited outcrop exposure. Entobians, *Gastrochaenolites*, *Trypanites* have been found at BH , however, these are not common. The latter locality exhibits the least amount of borings of the localities studied. This locality is also significantly different from the other ones, in that it comprises a linear trending series of large coral heads. This may represent a series of patchreefs, and the reduced amount of boring may be due to the lack of dead corals or dead areas on the corals. Corals are generally fragmented at the other localities, and may therefore have been bored after death. *Siphonodictyon* is able to penetrate live corals, and so new ichnogenus A would be expected at BH. It has not however, been found at BH. *Siphonodictyon* has not been found to bore in waters less than 1.5 m deep (Rützler, 1971). The lack of new ichnogenus A at BH may thus indicate very shallow waters.

Coral Preservation versus Amount of Boring

The corals are preserved to varying degrees, generally as molds with the negative impression of the original structure preserved. This type of preservation has been termed 'inverse porosity' (Buchbinder, 1975) or 'reversal' phenomenon (Supko, 1977) whereby the corals are represented in negative form because of dissolution of the original skeleton. The variation in extent of sediment fill in the original interseptal spaces of the corals produces the variable amounts of 'apparent' preservation. The branching corals are the least well preserved and the most extensively bored of the coral types. The solitary corals on the other hand, exhibit the best preservation and the least amount of boring. With respect to the massive corals, the hollow structureless molds are more extensively bored than the molds preserving the negative impressions and casts of the corallites. There is thus a strong correlation between the amount of boring and the degree of coral preservation.

Colonies of branching stylophorids are completely leached in the Miocene Ziqlag Formation of Israel (Buchbinder, 1977), while massive faviids and poritids are preserved to varying degrees. It was tentatively suggested that the branching stylophorids may have been dissolved due to the destruction of the corals' internal structure by intense microboring activity. The other types of corals may have been less susceptible to such bioerosion. (Buchbinder, 1977). However, no evidence of boring was found in any of the corals.

The differential preservation of the corals may also be due to the differences in the structure of their internal elements. The branching stylophorids of the Bluff Formation have small, widely spaced corallites, while the corallites of the massive corals are much more closely packed and generally larger. The solitary corals and *Diploria* sp. have large interseptal areas open to the surface. The larger size of the corallites, and interseptal spaces, may have allowed more sediment to infill the coral. This sediment was lithified preserving the coral structure in negative form, so that when differential dissolution removed the aragonitic coral skeleton, the lithified internal sediment was left preserving the negative form of the coral.

Macintyre (1977) suggested that the presence of submarine crusts around corals influences the degree of cementation of the coral and hence its preservational potential. Submarine algal crusts are best developed around the branching corals of the Bluff Formation. These crusts may have prevented sediment fill of the corals, and therefore when the aragonite skeleton was preferentially leached, there was no internal sediment to preserve the coral structure.

Factors Influencing Amount of Boring

The size (and shape) of the coral and the density of the coral skeleton are the two major factors influencing the degree of boring in any particular coral, assuming the environmental settings are the same for the host corals. These factors can explain why the branching corals are more extensively bored than the other coral types.

The branching corals have a greater surface area per volume than the more equidimensional massive corals and thus have a relatively greater area exposed to boring activity. *Entobia*, the dominant boring in the corals of the Bluff Formation, is not found at a depth greater than 1.5 cm in the substrate. In recent environments, *Cliona* does not bore to a depth greater than 2 cm as it must maintain good water circulation through its gallery system (Highsmith, 1981a). As sponge borings commonly dominate modern bioerosion, most boring is concentrated around 2 cm of the coral perimeter (Neumann, 1966; Highsmith, 1981a). The small branching corals of the Bluff Formation are therefore, more extensively bored than the massive corals, which are larger than the branching corals and coral fragments. As Highsmith (1981a, p. 335) noted "Massive corals have a potential escape in size from catastrophic bioerosion".

It has also been suggested that boring activity is increased in corals with denser skeletons (Highsmith, 1981b) since they offer greater protection to the borers against predatory organisms and are less adept at recolonising dead areas (Highsmith, 1981b). Most boring organisms will only settle onto, and bore into dead areas of the skeleton, because they are unable to protect themselves against the live polyps. It is not known which are the densest corals in the Bluff Formation.

Boring Styles

Slender borings in *Diploria* sp. tend to meander in between the coral septa casts (Plate 8C, D). They are therefore present in what was originally the aragonitic skeleton. The borings have thus avoided the original voids.

Similarly, in the *Montastrea* corals, the borings tend to encircle the corallite structures, again avoiding the original voids which would have been in the middle of the corallites (Plate 8E). Ward and Risk (1977) found that *Cliona vermifera* colonies in *Montastrea annularis* took on a string-of-beads appearance due to the concentration of chambers in the corallites. The borings had followed the zones of least resistance, expanding their chambers in the voids of the corallites. This pattern of boring is not observed in the corals of the Bluff Formation, where the borings have tended to avoid the corallites.

In contrast, the borings in the branching corals may either branch around the corallites to avoid them, or penetrate straight through the middle (Plate 8F). Where they penetrate through, a xenoglyphic sculpture is preserved on the borings (Plate 3A).

D. MODERN REEF BIOEROSION

Introduction

The term bioerosion was introduced by Neumann (1966, p. 92) to cover the process of "...removal of consolidated mineral or lithic substrate by the direct action of organisms...". To this must be added the removal of organic and man-made substrates. The end product of bioerosion is a 'bioerosion fabric' (Ekdale *et al.*, 1984), that may be recognised in the fossil record. This fabric results from a combination of boring, scraping, biting, gnawing, drilling and crushing by bioeroding organisms, and may be modified by diagenesis.

In carbonate environments, bioerosion is ubiquitous on the seafloor (Warme, 1977) and is especially common in rocky shore environments, hardgrounds and reefs. Bioerosion in reefs is well documented (e.g. Goreau and Hartman, 1963; James, 1970; MacGeachy and Stearn, 1976; Warme, 1977; Bromley, 1978; Risk and MacGeachy, 1978; Highsmith, 1981a; Highsmith *et al.*,1983). The rate of growth of reefs is dependent at least in part on the rate of both physical and biological destruction (MacGeachy and Stearn, 1976). Biological destruction may facilitate extensive physical destruction, by weakening the substrate; whereas physical breakdown in turn produces more substrate surfaces for biological degradation.

In order for a comparison to be made between the borings observed in the fossils of the Bluff Formation and modern coral reef bioerosion, a brief review of the most important recent borers and those thought to be represented in the Bluff Formation is necessary: sponges, bivalves, polychaete annelids, sipunculids, barnacles, phoronids and thallophytes.

Sponges

There are 34 species of boring sponges belonging to 5 families that are described in the literature: Adociidae, Clathriidae, Clionidae, Spirastrellidae and family uncertain (Pang, 1973; Hartman, 1957, 1958; Bergquist, 1968; Rützler, 1971, 1974; Pomponi, 1976; Rützler and Bromley, 1981). Of these, the clionids and adociids are the most significant in terms of bioerosion.

All sponges bore in a similar manner, producing characteristic silt-sized chips of substrate by mainly chemical means, and leaving a pitted microsculpture on all surfaces bored (Nassonov, 1883; Warburton, 1958; Cobb, 1969; Rützler and Rieger, 1973; Hatch, 1980; Pomponi, 1980). Only 2 to 3% of the total material removed is in solution (Rützler and Rieger, 1973). The other 97 to 98% is removed in the form of chips. In all species of sponges, the chips produced are of about the same size, from 16 to 85 µm (Rützler and Rieger, 1973), even though the size of the excavated tunnels and cavities may vary considerably.

These chips can be distinguished in the adjacent sediment (Fütterer, 1974; Halley *et al*, 1977; Acker and Risk, 1985)). Fütterer (1974) recognised that sponge chips comprised 2-3% of the sediment from the Persian Gulf and northern Adriatic Sea, and 30% of the sediment of Fanning Island lagoon. 40% of silt sized sediment in patch reefs of Belize can also be attributed to sponge chips (Halley *et al.*, 1977).

The amount of material removed by boring sponges is high. Estimates range from 0.9% to 60% of skeletal material removed from coral heads (Table 7). Neumann (1966) estimated that *Cliona lampa* in Bermuda reefs was capable of removing 22-25 kg $CaCO_3$ $m^{-2}yr^{-1}$ (Table 7). However, Rützler (1975) confirmed Neumann's fears that short term experiments (those under 6 months) resulted in unusually high initial penetration rates. After 6 months, initial stimulation of the substrate is surpassed by competition for food and space, which slows down the boring activity. Using experiments of up to 12 months in Bermuda, Rützler (1975) showed that the boring rate curve flattened after 6 months, and

Location	Substrate	average % Excavated	% Excav. by Sponges	Sponge Bioerosion Rate *	References
Bermuda	several	-----	-----	22 -25	Neumann, 1966
Bermuda	several	-----	-----	0.26	Rützler, 1975
Florida	coral heads (8)	28	12.3	2340**	Hein & Risk, 1975
Curaçao	coral heads	1.5	1.4	2.5 -3.3	Bak, 1976
Barbados	*M. annularis* (36)	12.5	11.2	-----	MacGeachy & Stearn, 1976
Barbados	*M. annularis* (45) *S. siderea* (35) *P. astreoides*(39)	-----	8.4 4.5 4.7	-----	MacGeachy, 1977
St. Croix	core	-----	35.6	0.2 - 3.3	Moore & Shedd, 1977
Barbados	-----	-----	-----	0.1 - 0.5	Stearn & Scoffin, 1977
Enewetak	*G. retiformis* (18) *P. lutea* (54) *F. pallida* (13) coral rubble (11)	7.9 2.5 1.2 5.5	6.0 2.1 0.9 3.8	-----	Highsmith, 1981a
Belize	*M. annularis* (19) *M. cavernosa* (23) *P. astreoides* (23)	7.9 5.3 4.8	7.4 4.6 4.1	-----	Highsmith *et al*, 1983
Grand Cayman	coral heads	-----	20	8	Acker & Risk, 1985

Table 7. Table showing the importance of boring sponges in modern reef bioerosion.
* is equal to kg $CaCO_3$ $m^{-2}yr^{-1}$.
** indicates $mm^3 yr^{-1}$.

that boring potential is only about 0.26 kg $CaCO_3$ $m^{-2}yr^{-1}$, somewhat lower and more realistic than the figure of Neumann (Table 7). Bak (1976) calculated a rate of 2.5-3.3 kg $CaCO_3$ $m^{-2}yr^{-1}$ for *Cliona peponaca* in Curaçao, Moore and Shedd (1977), a rate of 0.19-3.29 kg $CaCO_3$ $m^{-2}yr^{-1}$ from core from St. Croix, Stearn and Scoffin (1977), a rate of 0.08-0.38 kg $CaCO_3$ $m^{-2}yr^{-1}$ in Barbados and Acker and Risk (1985) calculated a value of 8 kg $CaCO_3$ $m^{-2}yr^{-1}$ for *Cliona caribbaea* in Grand Cayman Island (Table 7). The actual long term rates of sponge boring on coral reefs will depend on the rate new substrate surfaces become available (Highsmith, 1981a), the type of substrate, the environmental conditions and the particular species of sponge (Hutchings, 1986).

No bioerosion rates are known for *Siphonodictyon*, however, it is apparent that they are locally abundant and can produce large cavities (Rützler, 1971; MacGeachy and Stearn, 1976; MacGeachy, 1977; Bromley, 1978; Risk and MacGeachy, 1978). Unlike most clionids they can attack live corals (Rützler, 1971).

Spirastrellids can also produce large excavations, with wide tunnels extending up to 15 cm into the sustrate (Bromley, 1978). MacGeachy and Stearn (1976) reported that spirastrellids were responsible for 0.8 to 13.2% of the total bored volume of corals from Barbados.

Bivalves

Boring bivalves play a fundamental role in the bioerosion of coral reefs. Members of the superfamilies Myacea, Adesmacea, Veneracea, Saxicavacea, Gastrochaenacea, Cardiacea and Mytilacea have developed the boring habit (Yonge, 1963).

Boring bivalves are generally less important in coral reef bioerosion than sponges or worms. Few quantitative estimates on the amount or rates of bioerosion are available. Hamner and Jones (1976) calculated that bioerosion by *Tridacna crocea* on the Great Barrier Reef amounted to 0.14 kg $CaCO_3$ $m^{-2}yr^{-1}$. In addition, the bivalve produces about 0.06 kg $CaCO_3$ $m^{-2}yr^{-1}$ by increasing its shell size, thus, giving a sediment production rate of 0.2 kg $CaCO_3$ $m^{-2}yr^{-1}$. Locally this value may rise to 4.5 kg $CaCO_3$ $m^{-2}yr^{-1}$.

MacGeachy and Stearn (1976) documented boring bivalves from Barbados, where they were responsible for 0.2 to 1.9% of the total bored volume of coral heads.

Polychaete Annelids

Polychaetes may be the most abundant and diverse infauna on coral reefs, comprising over 70 families (Highsmith, 1981a). They were found to be the principle borers on *Porites lutea* blocks during a one and a half year experiment at Lizard island, Great Barrier Reef (Davies and Hutchings, 1983). During this time, boring sponges were insignificant. Three families are important in bioerosion: the Eunicidae, Spionidae and Sabellidae. The borings range in morphology from single tubes to U-shaped borings, to irregular, branching networks.

Highsmith (1981a) found that the polychaetes accounted for over 80% of the endolithic organisms in the coral heads studied at Enewetak, though he could not be certain how many of these were euendoliths as opposed to cryptoendoliths. In reef limestones of the East Indian Ocean, polychaete densities are as high as 49000 m^{-2} (Kohn and Lloyd, 1973). Davies and Hutchings (1983) calculated that annual erosion rates by boring polychaetes, at Lizard island, Great Barrier Reef, were 0.7 kg $CaCO_3$ $m^{-2}yr^{-1}$ (on the reef front), 0.8 kg $CaCO_3$ $m^{-2}yr^{-1}$ (on the reef flat), and 1.8 kg $CaCO_3$ $m^{-2}yr^{-1}$ (on patch reefs), giving an average value of 1.11 kg $CaCO_3$ $m^{-2}yr^{-1}$ (my calculation). Hutchings and Bamber (1985) found that the rates over a greater variety of reef environments at the same locality ranged from 0.33-4.82 kg $CaCO_3$ $m^{-2}yr^{-1}$. This shows that the bioerosion rates vary according to substrate and environment similar to the variations in the densities and species composition of the polychaetes themselves (Hutchings, 1986).

Sipunculids

Sipunculids may bore into calcareous or non-calcareous substrates. These worms can produce borings up to 1cm in diameter, which are single entranced, straight or sinuous, and which commonly taper both towards the base and entrance ((MacGeachy and Stearn, 1976).

Boring sipunculid densities in reef limestones have been estimated at 500 m^{-2} to 800 m^{-2} (Kohn, 1970; Rice, 1976). They are less abundant than polychaetes at Enewetak, being most abundant in the rubble (Highsmith, 1981a). At Barbados, they are responsible for 0.2 to 2.0% of the total bored volume of corals (MacGeachy and Stearn, 1976). Although they may be locally numerous in certain coral heads (Warme, 1977), they are of only minor importance in reef bioerosion compared to other endolithic organisms.

Barnacles

Barnacles may be borers (e.g. *Lithotrya dorsalis*) or embedders (*Ceratoconcha domingensis*). They are responsible for 3.7 to 11.3% of the total bored volume of corals at Barbados (MacGeachy and Stearn, 1976). Trudgill (1976) reported a bioerosion rate of 0.844 cm y^{-1} at Aldabra, Indian Ocean.

Phoronid Worms

Phoronids are lophophorate worms, and probably bore by chemical means (Bromley, 1970). Marcus (1949) observed over 150 borings cm^{-2} in one shell fragment. The borings are generally parallel to the surface and are typically 0.2-0.3 mm in diameter, fitting the animal closely. The worm can be up to 6 mm in length and can form pseudocolonies by budding (Marcus, 1949; Bromley, 1970).

Microboring Organisms

Algae, fungi and bacteria are pervasive in coral reefs. Kobluk and Risk (1977a) observed 100% infestation of iceland spar surfaces by boring algae, and removal of the outer 30 μm of carbonate in about 213 days, in Jamaica. Bathurst (1966) described how microborers could breakdown skeletal grains by centripetal boring to form a micrite envelope. Micrite cement is also precipitated in vacated borings (Bathurst, 1966; Kobluk and Risk, 1977b; Tudhope and Risk, 1985), though the latter authors did not feel that this completely offsets the dissolution of carbonate by the boring activity. Tudhope and Risk (1985) calculated a bioerosion rate of 0.35 kg $CaCO_3\,m^{-2}yr^{-1}$ by algae on the lagoon floor of Davies reef, Great Barrier Reef. During the boring process, the endolithic filaments may project out into the porespace betwwen grains and become calcified (Kobluk and Risk, 1977b). These calcified filaments may break off due to agitation of the waters and thus would become part of the fine grained sediment.

Bacteria may also be important in skeletal breakdown by preconditioning the substrate for other borers (Risk and MacGeachy, 1978). DiSalvo (1969) found that weakened areas of the corallum of *Porites* contained from 10^4 to 10^5 bacteria gm^{-1} dry weight. They also found that the bacteria were capable of digesting chitin, which would weaken the substrate for other borers.

Discussion of Modern Bioerosion

Boring sponges are the most important bioeroders in most areas in terms of relative skeletal damage (Table 8). In Florida, sponges and polychaetes are the most important (43% and 47% respectively), while bivalves are less significant (Hein and Risk, 1975). At Enewetak sponges account for 75-83% of skeletal damage (Highsmith, 1981a); in Barbados they are responsible for over 90% (MacGeachy and Stearn, 1976; MacGeachy, 1977); in Curaçao, 97% (Bak, 1976) and in Belize, 85-94% (Highsmith *et al.*, 1983). Total skeletal damage is estimated at 28%, 13%, 6%, 4% and 1.5% in Florida, Barbados, Belize, Enewetak and Curaçao respectively (Hein and Risk, 1975; Bak, 1976; MacGeachy and Stearn, 1976; Highsmith, 1981a; Highsmith *et al.*, 1983). Polychaetes are generally more important than bivalves at these localities, although both are much less important than sponges (with the exception of Florida). Polychaetes are abundant but do not remove large amounts of hard material. The importance of sponges stems from the morphology of their borings, as one sponge can excavate a ramifying network of tunnels, compared to the discrete single borings of bivalves.

A comparison of the rates of bioerosion have important implications for coral reef budgets (Table 9). Sponges may remove from 0.26-25 kg $CaCO_3$ m^{-2}yr^{-1}, polychaetes 0.3-4.8 kg $CaCO_3$ m^{-2}yr^{-1}, bivalves from 0.14-4.5 kg $CaCO_3$ m^{-2}yr^{-1} and algae 0.35 kg $CaCO_3$ m^{-2}yr^{-1}. Rates of reef calcification range from 0.3 to about 12 kg $CaCO_3$ m^{-2}yr^{-1}

LOCATION	% EXCAVATED	RELATIVE IMPORTANCE OF BORERS % SPONGES	WORMS	BIVALVES	REFERENCE
FLORIDA	28	43	47	10	Hein & Risk, 1975
CURAÇAO	1.5	97	<3	<3	Bak, 1976
BARBADOS	12.5	90+	<4	locally abundant	MacGeachy & Stearn, 1976
ENEWETAK	4.3	75-83	0.5-25	-----	Highsmith, 1981a
BELIZE	6	85-94	2.8-15	0-11	Highsmith *et al*., 1983

Table 8. Table showing the relative importance of boring sponges, worms and bivalves in modern reef bioerosion.

	kg $CaCO_3$ m^{-2} yr^{-1}	REFERENCES
Sponge Bioerosion	0.3 25	Neumann, 1966; Rützler, 1975; Bak, 1976; Moore & Shedd, 1977; Stearn & Scoffin, 1977; Acker & Risk, 1985.
Polychaete Bioerosion	0.3 - 4.8	Davies & Hutchings, 1983; Hutchings & Bamber, 1985.
Bivalve Bioerosion	0.1 - 4.5	Hamner & Jones, 1976. 6
Algal Bioerosion	0.4	Tudhope & Risk, 1985.
Coral Calcification	0.3 - 12	Davies & Hopley, 1983; Smith, 1983.

Table 9. Bioerosion rates of selected boring organisms from modern reefs, compared to rates of coral calcification.

(as summarized by Davies and Hopley, 1983, Smith, 1983). Smith (1983) estimated an average rate of 1-1.2 kg CaCO$_3$ m^{-2}yr^{-1}. Thus, bioerosion may sometimes exceed reef calcification, and indeed it has been suggested that rates of bioerosion may be greater than rates of skeletogenesis in Florida (Hein and Risk, 1975). However, it is apparent that rates of bioerosion are not constant over time or space (Hutchings, 1986). Thus, the estimated rates of bioerosion must be used with caution.

Most bioerosion is concentrated within 2 cm of a dead surface . Therefore, massive corals may escape intensive bioerosion due to their size and shape, compared to the susceptibility of branching corals. In addition, denser corals may be more susceptible to bioerosion than less dense corals, as they offer greater protection against grazing predators and are less adept at colonizing dead areas (Highsmith, 1981b). Most dead areas, and thus borings, are concentrated at the base of the coral (Goreau and Hartman, 1963). This reduces the stability of the colony, especially on slopes, and renders it less resistant to wave attack (MacGeachy, 1975). Once toppled the corals are subject to further bioerosion or physical degradation. In addition, the endolithic organisms attract other bioeroders as they are an important food source for many grazers.

E. COMPARISON OF BIOEROSION IN THE BLUFF FORMATION WITH MODERN REEF BIOEROSION

The bioerosion assemblage of the Bluff Formation is represented only by the borings of organisms. There is an absence of grazing, rasping, biting and drilling structures, common in modern bioerosion fabrics. This is probably due to their destruction during diagenesis, rather than their absence in the original environment.

The boring assemblage found in the corals of the Bluff Formation is dominated by the borings of sponges (*Entobia*, ichnogenus A, problematica), which are responsible for about 85[illegible] of the borings (Table 5), and in this respect is comparable to the relative abundance of boring sponges in modern reefs. Next in abundance are the borings of worms (*Trypanites*, some *Gastrochaenolites*, *Talpina*, *Maeandropolydora*, *Caulostrepsis*), which are responsible for about 9% of the borings observed in the corals of the Bluff Formation, again comparable to their relative importance in the modern. However, the worm borings do not outnumber the sponge borings as they commonly do in recent coral reefs. In the corals of the Bluff Formation, bivalve borings (*Gastrochaenolites*, ichnogenus B?) are locally abundant (responsible for up to 5% of the borings) but far less numerous than the sponge and worm borings. Bivalve borings in the Recent are similarly only locally abundant and are less important than worms and sponges in most coral heads.

The amount of total bioerosion in the coral molds of the Bluff Formation is somewhat higher than in recent corals (compare Table 4 with Tables 7 and 8). The average percentage boring in the Bluff Formation is 38%. This high figure is probably a result of the small size of many of the corals. Most of the borings in the Bluff Formation are concentrated around the edge of corals. *Entobia* penetrates to a depth of 1.5 cm, commonly less, whereas *Trypanites* is concentrated within about 2.5 cm of the surface. *Talpina*, *Maeandropolydora* and *Caulostrepsis* are generally within 1.5 cm of the substrate surface. Ichnogenus A is observed down to about 5 cm into the substrate, whereas ichnogenus B and *Gastrochaenolites* are found down to about 7 cm. In recent corals, bioerosion is concentrated within 2 cm of a dead surface, and generally towards the base of corals (Goreau and Hartman, 1963; Hein and Risk, 1975).

The irregular distribution of the borings in the molds of the Bluff Formation is also similar to the irregular variations observed in different modern reef environments. The abundance of borers in recent reefs varies not only from one reef to another due to differing environmental parameters, but also in the same reef (Hutchings, 1986). This variation is apparent in the skeletal molds of the Bluff Formation, where abundance and type of borings varies from one locality to another, and even at the same locality. Certain borings may be locally abundant in one particular coral mold, but rarely found elsewhere.

It is apparent that the bioerosion in the Bluff Formation is similar in terms of organism composition, relative abundance and localised distribution, to present day bioerosion on coral reefs. Although rates of bioerosion cannot be determined for these ancient borings, the quantitative data presented does suggest that the importance of bioerosion in the middle Tertiary was similar to that of the present day.

In recent reefs, the products of sponge boring activity are highly characteristic and potentially preservable. The chips have been recognised in recent sediments (Fütterer, 1974; Halley *et al.*, 1977; Acker and Risk, 1985) and in ancient rocks (Kobluk, 1981a; 1981b). These chips have not been found in the Bluff Formation rocks, even though sponge boring activity was so high. This absence may be due to pervasive dolomitization or a result of their absence in the original sediment. Acker and Risk (1985) noted that most of the recent sponge chips in their study area in Grand Cayman, were riddled with algal and fungal borings. This would eventually reduce distinctive chips into lime mud. They also noted that the abundance of sponge chips was very low compared to the high rate of chip production. The abundance increases below 25 m (the depth of maximum sponge abundance) and they suggested that the chips are being transported downslope, possibly ending up at the bottom of the Cayman Trench. The chips of the Bluff Formation may have suffered a similar fate, with those not transported downslope, being degraded by microorganisms (microborings are in the Bluff Formation). Any that did survive, may have then been obscured by later diagenesis.

The similarity of the boring assemblage of the Bluff Formation and recent coral reef environments indicates that similar interactions between reef destruction and construction were operative by Oligocene-Miocene times. In addition, this bioerosive activity was occurring on a similar scale to that of the present day. Boring organisms not previously recognised in the fossil record have distinct borings which are well preserved in the Bluff Formation, and may be common in other rocks. The presumed affinities of these are an important aspect of recent bioerosion (especially *Siphonodictyon*), and their borings are at least locally abundant in the Bluff Formation.

F. BIOEROSION IN THE FOSSIL RECORD

Examples of ancient bioerosion assemblages in reefs are rare, but there are enough documented examples of isolated borers to produce a general trend of borers through the geologic record.

Trypanites and fossil sponge borings have been found in archaeocyathid patch reefs of the lower Cambrian Forteau Formation of southern Labrador (James *et al.*, 1977; Kobluk, 1981a). In this Formation *Trypanites* densities are up to 5720 m ^{2} (Kobluk *et al.*, 1978). The extinction of archaeocyathids at the end of the early-middle Cambrian left no skeletal metazoans to form reefs. Therefore, from the middle Cambrian to the middle Ordovician, the only extensive niche available for bioeroding organisms were hardgrounds (James *et al.*, 1977). Diversity of hardground fauna reached a maximum in the middle Ordovician and remained at this level until the late Cretaceous (Palmer, 1982). Reefs were re-established in the middle Ordovician, providing suitable niches for macroboring organisms, which began to radiate and diversify. Scattered examples of macroborings exist from the middle Ordovician to the Mesozoic (see Kobluk *et al.*, 1978; Pickerill and Harland, 1984). Pemberton *et al.* (in press) documented *Trypanites* from Devonian stromatoporoids from north east Alberta, which had bored from 6 to 24% of these heads. It has been suggested that macroborings played only a minor role in the reef system during the Paleozoic (Ekdale *et al.*, 1984).

By the Triassic, bioerosion may have played a more significant role in reef destruction (Zankl, 1971). Examples of bioerosion in the Jurassic of Morocco, show that boring bivalves and worms had become very important, while boring sponges were rare (Warme, 1977). Sponges had increased in importance by the Cretaceous, especially in rudist buildups (Kauffman and Sohl, 1974). Boring worms were also significant aspects of bioerosion at this time (Voigt, 1970, 1971, 1972). Evidence from the middle Tertiary Bluff Formation presented here, indicates that by this time, sponges had come to dominate reef bioerosion, surpassing both the boring worms and bivalves in numbers as well as importance. At about the same time, boring sponges had become important in other marine environments, e.g. in the Miocene, Pliocene and Pleistocene infralittoral limestones of southern Italy (Bromley and D'Alessandro, 1984). The relative importance of micro- to macroborings in the Bluff Formation has not been established, but the microborings are common. Microborings are common from the lower Cambrian to the present (James and Kobluk, 1978).

G. SYNOPSIS

1. Borings are preserved as natural casts in fossil molds of the Bluff Formation.
2. The borings have occluded much of the moldic porosity.
3. The fossil molds are on average 38% bored (0 to 75%); the highest amount of boring being in the stylophorid corals.
4. Entobians dominate the boring assemblage, being responsible for about 75% of the borings. *Trypanites, Gastrochaenolites, Maeandropolydora, Talpina, Caulostrepsis,* new ichnogenera A and B, and microborings are also present.
5. Using the phyletic relationships of the ancient borings, it is apparent that this ancient boring assemblage is comparable in terms of composition and scale of damage to that of modern coral reefs. Boring sponges dominate in both. They are responsible for up to 90% of the bioerosion in modern reefs and up to 85% in the molds of the Bluff Formation. Boring worms are the next most important in both. Bivalves and barnacles may be locally abundant. Boring worms in the Caymanian examples are not as numerically abundant as they tend to be in modern reefs.
6. Microborings are abundant in the molds of the Bluff Formation.

III. PETROGRAPHY OF THE ROCKS OF THE BLUFF FORMATION

A. BLUFF FORMATION DOLOSTONE

The Bluff Formation is composed of hard, white to beige, finely crystalline dolostone, in which the precursor textures are commonly preserved. It is generally over 90% dolomitized, but locally grades into a dolomitic limestone. The dolostones have a wide lateral extent, cropping out on all three Cayman islands. Recent borehole data confirms that the dolostones extend to a depth of at least 20 m. Earlier boreholes drilled by K. E. Merren and D. P. Hamilton (1956) on Grand Cayman Island recorded limestones to a depth of 401 m (see Emery and Milliman, 1980). However, it is not known whether these are limestones or dolostones, since the surface outcrops of the Bluff Formation were considered at that time to be limestone rather than dolostone.

The dolostone is finely crystalline, with average crystal sizes of about 5-10 μm. The fine dolomite crystals are cloudy and appear brown in thin section, probably due to minute inclusions from the precursor carbonate. They have a hypidiotopic to xenotopic texture, and some of the scattered euhedral crystals in the groundmass are zoned with cloudy centres and clear rims. There is little intercrystalline porosity in the groundmass dolomite. Irregular patches of coarser dolomite crystals (30-40 μm) also occur. These coarser crystals have an idiotopic texture and are much clearer than the finer crystals.

B. SUBMARINE TEXTURES

Several features are indicative of submarine cementation, though these features are not common. These include algal crusts, peloids, micrite cement and isopachous cements. All these features have been dolomitized.

Many of the coral molds (at all localities) are coated and bound by a laminated, smooth to knobbly crust, up to 1 cm thick (Plate 10F). Although composed mainly of coralline algae, it also contains encrusting foraminifera, bryozoa, dolomicrite and calcite spar. The skeletal allochems are much darker than the dolomicrite in the crusts. *Entobia* borings cut through the laminated crust, indicating that the crust was lithified on the seafloor. The borings do not penetrate the calcite spar, implying that this calcite was a later diagenetic product.

Peloids are common in some coral cavities (BH 1227, BH 1229). The elongate to spherical peloids are about 50 µm in diameter and are composed of structureless dolomicrite. They are commonly embedded in a fine dolomitic groundmass (Plate 10B). They are distinguishable in this groundmass due to their darker colouration. Peloids similar to these are common features in coral cavities in recent reef environments (Shinn, 1969; Macintyre *et al.*, 1968; James *et al.*, 1976; Macintyre, 1977). Some peloids occur in skeletal molds associated with dolomitized microborings (HRQ 1208). These peloids vary in shape from spheroids to rod-shaped structures.

Thin isopachous dolomitic rinds, which commonly fringe primary voids in replaced corals (BH 1229), are composed of bladed crystals, about 30 µm in length (Plate 10B-E). The internal structure of the coral is represented only by internal sediment similar to that of the groundmass. A dark outline is apparent in the dolomite at the base of the isopachous fringes, which probably represents the original outline of the coral (Plate 10D, E). The isopachous crystals are syntaxially overgrown on this coral substrate (the dark outline). This type of optically continuous cementation is typical of acicular aragonite cements in recent reefs (Lighty, 1985). Some of the more equidimensional crystals possibly resulted from a HMC precursor cement.

Dolomitized internal sediment and/or dolomitized micritic cement occur in the same cavities as the isopachous fringe cements. This sediment or cement is commonly geopetal, and is overlain by, or interlayered with, the isopachous to more equidimensional dolomitized cements (Plate 10D). It was therefore cemented contemporaneously with deposition. Both micritic cement and micrite infills are common within recent reef cavities (Friedman, 1985).

C. FOSSIL PRESERVATION

The dolostones are highly fossiliferous, with locally abundant coral, red algae, molluscs, foraminifera, echinoids and bryozoa. Numerous borings preserved as natural casts in the [illegible]tal molds attest to the abundance and diversity of the original endolithic community. The originally HMC components have generally been preserved (red algae, echinoid plates, some foraminifera, bryozoa), whereas the originally aragonitic components have been preferentially dissolved (corals, molluscs, some foraminifera).

Corals are the most conspicuous element of the fauna. Branching coral fragments of *Stylophora* are the most common type, but massive corals (e.g. *Montastrea* and *Diploria*) and solitary corals (e.g. *Antillia* and *Antillocyathus*) are also locally abundant. These are commonly only preserved as molds. The branching stylophorids are totally dissolved, with only the outer parts of the corallites preserved as a negative mold or imprint (Plates 1F; 3C). This is comparable to the preservation of the Miocene stylophorids of Israel (Buchbinder, 1977). The massive and solitary corals on the other hand exhibit more variable types of preservation. These are most commonly preserved as negative molds, where fine sediment has filled the interseptal voids, thus supporting the coral structure (Plates 9F; 8D). This type of preservation has been termed 'inverse porosity' (Buchbinder, 1975) or 'reversal' phenomenon (Supko, 1977) whereby the corals are represented in negative form because of dissolution of the original skeleton. The variation in extent of sediment fill produces the variable amounts of 'apparent' preservation. The solitary corals and *Diploria* have the most filling, and thus appear the best preserved. Rare corals have been dolomitized, generally by coarse (30-40 µm) clear rhombs. Another common feature is chalkification, with large patches of the coral being composed of soft, white dolomitic chalk.

The bivalves and gastropods have also been leached and are thus preserved as internal and external molds (Plate 8A, B). Bivalves, which may be articulated, are up to 5 cm in diameter. Gastropods range in size from about 100 μm to 10 cm. Although the molluscs are ubiquitous, they are not numerically abundant.

Red (coralline) algae are very abundant in the Bluff Formation, although they are commonly only visible in thin sections. They occur as encrustations around coral molds (Plate 10F), or as discrete particles (up to 1 cm long) in the matrix. At one locality (QR), spheroidal to elliptical rhodoliths up to 10 cm in diameter are common. Although dolomitized, the red algae have retained their original microstructure (Plate 9A, B). They are composed of 1-2 μm dolomite rhombs that are aligned, while the conceptacles are filled with slightly coarser rhombs (Plate 9D). The red algae appear much darker than the dolomitic groundmass. Many have been partially leached (Plate 9B), and this must have occurred prior to or during dolomitization since dolomite cement may line the secondary cavities (Plate 9C).

The ubiquitous foraminifera are locally very abundant (e.g. PCQ 1243, QR 1182). These may be dolomitized or leached, and examples of both are apparent in a single thin section (Plate 9E). Dolomitization preserves their original microstructure (Plate 9F). This is especially apparent in the *Amphistegina* where the crystallographic orientation of the wall structure has been retained (Plate 10A). Foraminifera may also encrust the coral molds with the red algae.

Dolomitized bryozoa and echinoid plates and spines are present, but are not as common as the other components. Each echinoid plate or spine has been replaced by a single dolomite crystal, although the original skeletal voids may be lined or filled with small dolomite rhombs (Plate 9G).

Fossil Distribution

Most of the surface outcrop of the Bluff Formation (HRQ, EEQ, BR, PCQ, CK, PBQ, Fig. 1B) is a skeletal wackestone to packestone, composed of abundant branching stylophorids, which have been extensively bored, abundant red algal particles and foraminifera. Foraminifera may be locally very abundant producing a foraminiferal packestone to grainstone. Molluscs and echinoid particles are ubiquitous but not numerically significant. Massive equidimensional to hemispherical corals (20 x 20 cm) and small solitary corals are less common than the branching coral types (i.e. the stylophorids). At QR, the massive and solitary corals become more significant at the expense of the branching stylophorids, producing a more mixed coral fauna. The massive corals are larger (50 x 50 cm) and/or more tabular in shape. Rhodoliths dominate the fauna in a bed 1.5 m thick in this area. The outcrop at BH, is a coastal section about 50 m in width and about 200 m in length. Much of this area is dominated by large massive coral heads, up to 2 m in height (e.g. *Montastrea, Agaricia, Diploria, Agathiphyllia*, poritids and faviids). Evidence of boring is reduced at this locality compared to all the others. In size and shape, this is comparable to a series of patch reefs. The other parts of the Bluff Formation probably represent a backreef- lagoonal area, dominated by thickets of *Stylophora*. Such a scenario is apparent in the modern lagoons of Grand Cayman Island.

D. POROSITY

The porosity which can be up to 25%, is moldic, intracrystalline, microvuggy (especially HRQ 1212) and intraparticle. Moldic porosity, which is the most common porosity type, can produce cavities up to 1 m in diameter, where large coral heads have been dissolved (e.g. at BH). On a larger scale, joints, solution cavities and caves also contribute significantly to the porosity and permeability of the Bluff Formation. The borings in the fossil molds have significantly occluded much of the moldic porosity.

E. DOLOMITE CEMENTS

Pores and cavities may be lined with coarse, euhedral dolomite rhombs up to 100 µm in length (Folk *et al.*, 1973). A distinct boundary exists between these and the groundmass dolomite and they are thus interpreted to be cements (Plate 18F). These cements represent about 5% of the dolomite of the Bluff Formation. They are analogous to the limpid dolomite crystals described by Folk and Land (1972; 1975) and Folk and Siedlecka (1974). These cements are of three types.

The first type of dolomite cement lines the pores and is composed of complexly interlocking crystals with c-axes perpendicular to the substrate (Plate 11A-C). These are up to 30 μm in length and may be zoned or may rarely contain hollow bands. The second type of dolomite cement consists of zoned, commonly cloudy-centred clear-rimmed rhombs (Plate 13A) up to 100 μm (average 50 μm). These are irregularly spaced in the pores, with c-axes parallel to subparallel to the substrate (Plate 11D). Where both types occur in the same pore, the second type is rooted on the first type. They are therefore different generation cements. These cements are comparable to the first generation cements documented by Jones *et al.* (1984), from dolostone and caymanite cavities of the Bluff Formation.

The dolomite rhombs commonly appear to be floating in the blue epoxy that fills the pores in the thin sections. SEM examination shows that this is either due to (1) an artifact caused by the angle at which the thin section was cut and/or (2) the presence of dolomite coated algal/fungal filaments in the pore (e.g. HRQ 801). The algal/fungal filaments have been coated with both generations of dolomite cements (Plate 11E, F). The first generation cements are 3 to 5 μm long, whereas the irregularly spaced second generation crystals are up to 40 μm long. Thin sections cutting parallel to obliquely to such filaments yield dolomite rhombs that appear to be floating in the pore.

A third, less common (CK 1177, QR 1184) type of dolomite cement, which is not observed in the same pores as the first two, consists of alternating zones of syntaxial dolomite and calcite towards the edge of the rhomb (Plate 18B). These zones are about 5 μm in thickness and are laterally continuous from crystal to crystal. The calcite zones commonly grade laterally into hollow bands. These are comparable to the third type of dolomite cements described by Jones *et al.* (1984).

Dolomite Spar

Rarely, large, subhedral dolomite crystals (up to 150 μm) occude cavities (PBQ 1060, PBQ 1062). These are rooted on the sparry calcite cement, and are thus later than the limpid dolomite and sparry calcite cements

F. DOLOMITE DISSOLUTION

Many of the limpid dolomite crystals show evidence of dissolution. The first generation limpid dolomite cements rarely exhibit hollow zones in the crystals. The second generation dolomite cements are commonly hollow or hollow zoned, even though adjacent zoned rhombs may be complete (Plate 12A-C; 13A, B). It is apparent from this juxtaposition that the hollow centres of the rhombs are a result of the preferential dissolution of the cloudy centres. Similar hollow dolomite rhombs have been reported by Evamy (1967), Folk and Siedlecka (1974), Land *et al.* (1975), Longman and Mench (1978), Ward and Halley (1985) and Rosen and Holdren (1986).

In places, the dolostone is highly porous and soft, giving rise to chalky patches that are up to 4 cm long (most notably in HRQ samples, especially 1212). Ng (1985a) documented this soft porous dolomite from around the joint systems in the Bluff Formation. When viewed in thin section it is apparent that many of these soft patches are a result of dolomite dissolution. The dolomite rhombs are hollow or completely degraded, and commonly appear to be floating in the blue epoxy. The porosity, as well as being intracrystalline, is also microvuggy in appearance (Plate 13E). This may be due to more extensive dissolution of several complete dolomite rhombs. This soft dolostone is comparable to the pulverulent chalk of Blank and Tynes (1965), and the pulverulite of Rose (1972) and Chafetz and Butler (1980). The latter believed their pulverulite to be a surface weathering phenomenon, characterized by the dissolution and disintegration of the dolomite. The soft dolostone in the Bluff Formation is related to permeability conduits.

G. CALCITE CEMENTS

Sparry and Poikilotopic Calcite

The spar calcite, which is from 40 μm to 2 mm long, either lines cavities or completely occludes pores. The crystals are patchfly distributed and are more common at localities HRQ (especially), PCQ, PBQ, EEQ and BR than at CK or QR localities. The calcites are typically equant, display abundant enfacial junctions, do not preserve any depositional textures, commonly exhibit drusy growth from a dolomitic substrate and line and occlude pores (e.g. Plate 12D). They are therefore cements. These are comparable to the second generation cements described by Jones *et al.*(1984).

Many of the crystals, especially the larger ones, are poikilotopic (Folk *et al.*, 1973), encasing and filling the hollow dolomite rhombs (Plates 10G; 12D-F; 13D-F). Some of the poikitopically enclosed dolomite rhombs appear to be 'floating' in the calcite (Plate 13D, F). However, closer inspection on the SEM reveals that this is either a function of the thin section angle or a result of the crystals being supported on endolithic dolomitized filaments (HRQ 801). Patches of dolomite rhombs apparently suspended in the blue epoxy (the soft patches), lie adjacent to the patches of dolomite suspended in calcite (HRQ 1212). This attests to the ability of the rhombs to exist *in situ* without needing the calcite to support them. Thus, the calcite has not necessarily replaced the dolomite. Many of the hollow rhombs are also filled by the calcite (Plate 13F). The calcite in the centres of these rhombs is in optical continuity with the calcite surrounding them, but not usually with the dolomite rims. The calcite therefore did not replace the dolomite, but passively filled and surrounded it. This is in good agreement with the petrographic features of the sparry calcite crystals themselves, which are characteristic of cements. The calcites are thus a later phase than both the dolomitization event and dolomite dissolution.

Spiky Calcite

Some of the calcite spar crystals that line cavities may be dissected with spikes aligned (Plate 14A) in the direction of the c-axis of the host crystal (PBQ 1060, PBQ 1063). These are comparable to the 'spiky calcite' described by Folk *et al.* (1985) and to the spiky calcite produced by fungi on iceland spar crystals (Jones and Pemberton, 1987a).

In the spiky calcite from the Bluff Formation, it appears that the spikes may have formed along dislocation or basal cleavage planes in the crystals (Plate 14B, C). Along each microcrack the spikes are developed on both upper and lower surfaces, thus pointing into each other. Rarely some are connected across this plane. The gradational development of the spikes is apparent across this same microcrack, with well developed sharp spikes at one end and blunt, stunted pillar-like structures at the other (Plate 14B). One calcite crystal is extensively dissected with the spikes (Plate 14D).

Knobbly Textures

Knobbly textures are apparent on the same spar crystals as the spikes. In areas of the crystal away from the spiky plane, a low relief, rough, knobbly texture is present (Plate 14E, F). A gradation is visible from a rough surface with etch pits and scattered knobbles, to a well developed knobbly texture, where the knobbles grade into each other. The knobbles are elongate, almond shaped bumps (6 x 3 µm) with variable sized attachment areas (Plate 15A), and are developed on an uneven irregular surface (at this magnification). These knobbles are aligned on any one crystal face (Plate 14E, F). This is especially apparent in the well developed knobbly texture. The knobbles are comparable to 'monticuli' described by Jones and Pemberton (1987b). Loose particles resting on the crystal face may be monticuli that have become unattached.

Columnar Calcite

These crystals are rare in the Bluff Formation (PBQ 1060) and occur in cavities in association with spiky calcite and calcified filaments. In transverse section, these crystals are triangular and resemble spikes (Plate 15B, C), being up to 60 µm long and up to 40 µm in width at the base. They are, however, made up of a number of (about six) crystallites, arranged in a concentric bundle (Plate 15D). A central cavity is visible in the centre of these bundles. One crystallite may extend further than the others, giving the crystal a very pointed termination. A distinct boundary exists between the base of these columnar crystals and the substrate (Plate 15B). Comparable columnar crystals have been described by Kendall and Broughton (1978) and Chafetz and Butler (1980).

Calcified Filaments

Calcified filaments (PBQ 1060, PBQ 1061) are found with the columnar crystals and spiky calcite. The filaments, which are about 3 μm in diameter and over 10 μm in length, are covered with acicular crystals of variable lengths (Plate 15E, F). Most of the crystals are about 0.5 μm in diameter and 1.5 μm in length, whereas a few reach about 4 μm in length. The crystals have pointed terminations and are therefore most likely calcite rather than aragonite. The c-axes of these crystals are perpendicular to the substrate. The small diameter and dichotomous branching suggests that the filaments are of fungal origin. They are comparable to the calcified filaments documented by Klappa (1979) from calcrete horizons.

Acicular Crystals

Clusters of acicular $CaCO_3$ crystals are common in coatings developed around boring casts and in the coral molds (QR 1182, CK 1174, CK 1176, HRQ 1213). They range in diameter from 1 to 6 μm, and in length from 5 to 40 μm. When viewed in thin section (QR 1182, Plate 18A), they are comparable to the needle-fibres described by Ward (1970), James (1972), Steinen (1973) and Harrison (1977), and to the whisker crystals of Supko (1971). The crystals are rooted on both dolomite and calcite.

SEM examination shows them to be of several types (Plates 16A-F; 17A-C). One group consists of blades of calcite stacked *en echelon*, with their c-axes mutually parallel (Plate 16A-C). These occur in an Fe and Mn rich coating on borings and lining the coral mold (CK 1176). These are 1 to 4 μm wide and up to 20 μm in length. They are irregularly clustered and appear to have no preferred orientation. Some appear to be breaking down into regularly shaped micrite sized particles (Plate 16C). These blades are comparable to the lublinite crystals found in caves (described by Stoops, 1976; Bögli, 1980). Associated with these are randomly oriented to parallel aligned, dense needle-fibres (Plate 16D, E). These are each composed of at least two fibres connected by a depressed axial plane. This gives the crystals a bladed appearance. The terminations of the fibres are pointed and extend beyond the axial plane, giving a ragged appearance to each crystal.

In another coating on the borings (HRQ 1213), tangentially oriented needle fibres are common, 0.5 to 1 µm wide and up to 30 µm in length. Each crystal is composed of a bundle of 4 fibres (Plates 16F; 17A, B). The terminations of the crystals are blunt because all the fibres end at the same point. The crystals are equidimensional in cross section. Irregular patches of fine micrite appears to be binding these needle fibre crystals together (Plate 17A, B).

In another coating examined (CK 1174), more stout crystals are common. These are about 1 µm in width and 4-5 µm in length. They are irregularly shaped with their c-axes perpendicular to the substrate (Plate 17C).

Microstalactitic Cement

The microstalactitic cement is composed predominantly of calcite and has an asymmetrical growth form (Plates 17D, E; 18B). It is developed only on the cavity roof (CK 1171, CK 1172). The cement is rooted on the dolomite substrate, which is composed of both dolomitic groundmass, dolomite type 1, 2 and 3 cements, calcite spar and micrite (Plate 18B). The microstalactitic cement is composed of alternating bands of light and dark calcite (a few bands of dolomite are apparent towards the outer edge of the cements), with a long axis up to 350 µm. The light bands are thicker than the dark ones. It has been suggested that the dark bands may be due to the presence of iron and/or organic inclusions (Ng, 1985b). The laminae are generally continuous and follow the contours of the cavity.

Bulbous terminations develop adjacent to protuberances on the cavity wall. Under crossed-nichols, a sweeping radial extinction is apparent (Plate 17E).

Jones *et al.* (1984) also documented microstalactitic cement from cavities in the Bluff Formation, consisting of alternating bands of calcite and dolomite. This compositional banding was interpreted to be an original feature, reflecting chemically fluctuating porewaters.

H. DARK MICRITE

Irregular shaped patches of dark coloured micrite are common in the dolostone of the Bluff Formation, especially at HRQ (HRQ 1208, 1215, 1195, 1194, 1209). They occur as small micron sized patches only visible in thin section, to larger bodies of micrite, up to 4 cm in diameter (Plate 17F). In 3-D, these are elongate bodies of micrite. These larger bodies of micrite may be homogeneous or laminated. Some of the laminated bodies have hollow centres, some of which contain Recent plant roots. Examination of the role of these plant roots in the diagenesis of the Bluff Formation is in progress (Squair, 1987 pers. comm.). The laminated micrite thus encases the root and the cavity around it. The micrite in these laminated bodies varies in colour from tan to orange to brown-black. Much of the black areas are composed of Mn and Fe dendrites and spheroids.

The small patches of micrite occur in the same samples as the larger bodies, and the patches generally decrease away from these larger bodies in any one thin section. The micrite in both is petrographically similar. It is much finer grained than the dolomitic host rock, the individual crystals being unresolvable under the petrographic microscope (Plate 18C-E). The micrite patches in the dolomitic groundmass are homogeneous. However, where it occurs adjacent to and within calcite cemented areas, it may develop a clotted texture (Plate 18E). The contact between this calcite spar and the micrite is fuzzy and irregular (Plate 18C, E).

The boundary between the dolomite and the micrite is irregular and gradational, especially at the contact between the larger bodies of micrite and the dolomitic groundmass (Plate 18D). At this boundary, dolomite rhombs appear to be encased in the micrite. Some have micrite cores, while others are embayed by the micrite around the crystal edges (Plate 18D). Relic dolomite rhombs are also visible in the smaller patches of micrite scattered around the larger micritic bodies. Where this micrite occurs in the skeletal allochems, e.g. in red algae, the texture has been obliterated. Dolomite in the same allochem preserves the original skeletal textures.

I. CAYMANITE

Caymanite is a fine grained, colour-banded, dolomitic, cavity fill in the Bluff Formation (Folk and McBride, 1976; Rigby and Roberts, 1976; Lockhart, 1986). It is associated with joints, caves, solution cavities and coral molds, where it can be geopetal or can completely occlude porosity. The fine, anhedral dolomite in the caymanite (Plate 13C) is interpreted to be detrital (Lockhart, 1986) and thus has a different origin to the dolomite of the host rock. Caymanite may have been formed from material derived from swamps at the surface of the island, and then to have been washed into the Bluff Formation via joints and solution cavities during seasonal storms (Lockhart, 1986).

Caymanite sedimentation has operated continuously, probably since the island's emergence. Cut and fill channels in caymanite, different orientations of laminae in the same cavity, caymanite phases separated by cement generations and caymanite clasts within caymanite laminae, support this idea. Limpid dolomite and sparry calcite cement may occur in cavities in caymanite (Plate 13C).

J. TERRA ROSSA

Terra rossa, a weathered carbonate product, is a common fill in cavities in the Bluff Formation at all localities, and ranges from a loosely packed soil to a well consolidated rock. It varies in colour from pale brown to deep red, and is composed of calcium carbonate (and clay minerals). The differences in consolidation and colour reflect different stages of terra rossa fill. Like the caymanite, the terra rossa represents a long continuous process, and thus cannot be dated.

Terra rossa fills joints, caves, solution cavities and moldic porosity, and may interfinger with caymanite. It commonly enhances the coral mold morphologies, filling them or staining the walls red. The extensive fill by this terra rossa attests to the high permeability of the Bluff Formation.

K. FLOWSTONE

Beige to dark brown laminated flowstone is common in the Bluff Formation (Smith, in prep). It occurs in joints, caves, solution cavities and moldic porosity, and may be interlayered with terra rossa and caymanite.

IV. STABLE ISOTOPE GEOCHEMISTRY

Diagenetic minerals such as calcite and dolomite, precipitated in equilibrium with porewaters, should exhibit oxygen isotopic compositions characteristic of these waters, as well as the temperature at which the minerals formed. This assumption is based on the premise that the minerals have remained more or less unmodified since crystallization. In this respect, dolomite should be more stable (less susceptible to isotopic exchange) than calcite at sedimentary temperatures, unless later mineral dissolution and reprecipitation has occurred (Epstein *et al.*, 1964). Carbon isotopic compositions are particularly useful in determining porewater evolution during diagenesis in near-surface meteoric environments, where inputs of soil CO_2 produce a wide variability of $\delta^{13}C$ values (Hudson, 1975; Allan and Matthews, 1977, 1982). Thus, the isotopic compositions should record the evolution of porewaters during diagenesis of the Bluff Formation.

A. ISOTOPIC SIGNATURES OF DOLOMITE IN THE BLUFF FORMATION.

The dolomites of the Bluff Formation have average $\delta^{18}O$ (PDB) values of +2.4‰ and average $\delta^{13}C$ (PDB) values of +2.0‰ (Fig. 10, Table 10); the values range from +1.4 to +3.3‰ and from -0.3 to +3.4‰, respectively.

Oxygen Isotopes

It is assumed that the $\delta^{18}O$ value of dolomite is indicative of the dolomitizing fluids rather than those of the original carbonates it replaced (Land, 1980). Even if this assumption is valid, it is uncertain to what extent later diagenesis will affect the isotopic signature of the dolomite. It has been suggested that dolomite is less susceptible to oxygen isotopic exchange with diagenetic fluids than calcite (Epstein *et al.*, 1964) and that the

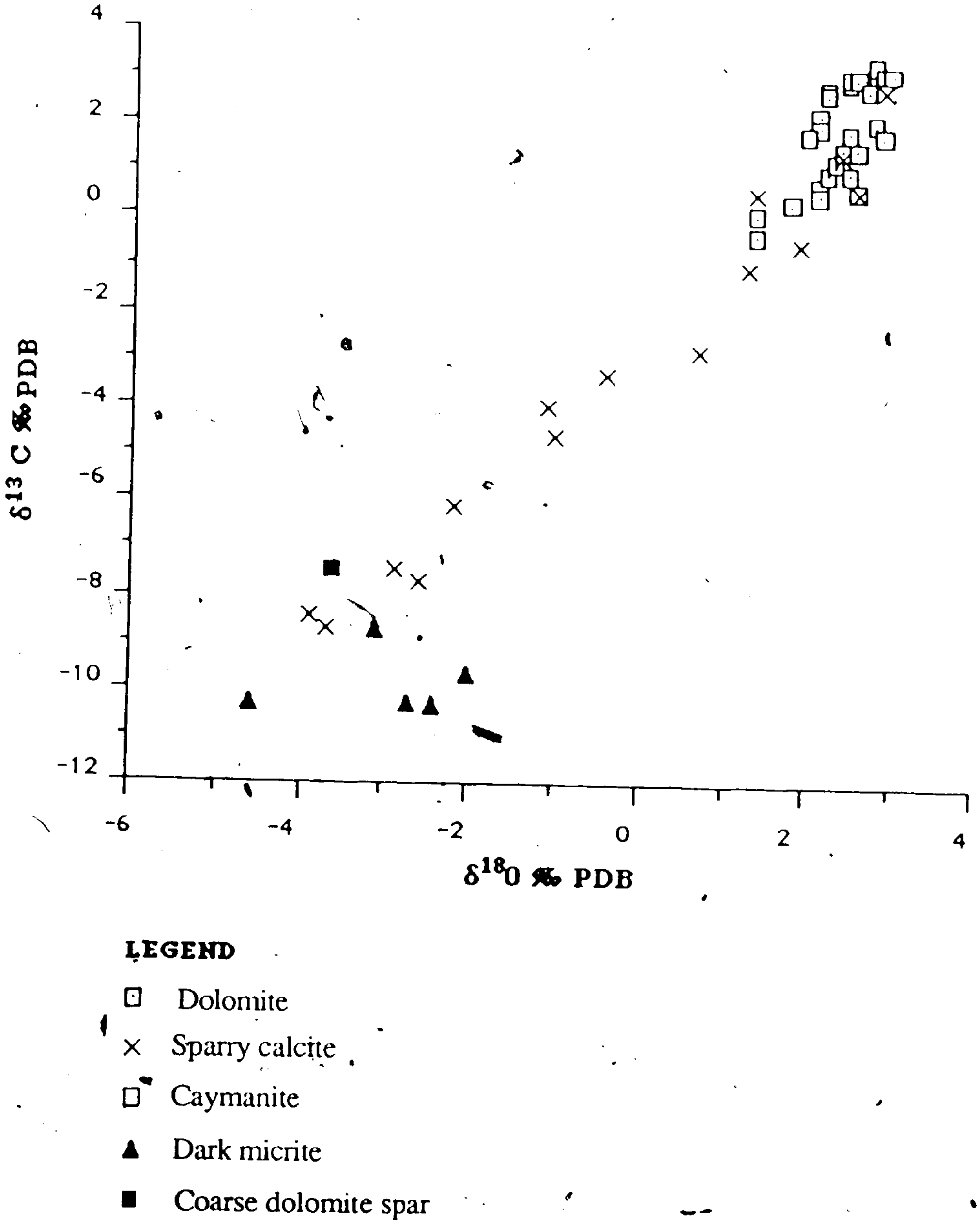


Figure 10. Oxygen and carbon isotope compositions of dolomite and calcite from the Bluff Formation.

SAMPLE	DESCRIPTION	dolomite $\delta^{18}O$ ‰ PDB	dolomite $\delta^{13}C$ ‰ PDB	calcite $\delta^{18}O$ ‰ PDB	calcite $\delta^{13}C$ ‰ PDB	% dolomite	mole % $CaCO_3$	AGE
801a	dolostone	2.3	1.3	——	——	100	——	OLIGOCENE to MIOCENE
801b	soft dolostone	2.1	0.8	——	——	100	——	
1209a	dolostone	2.8	2.1	——	——	100	52	
1062b	"	2.1	2.3	1.3	-1.0	77	50	
1063	"	2.6	0.7	2.6	0.7	99	——	
1195	"	2.1	0.6	——	——	100	——	
1223b	"	1.4	-0.3	-2.2	-6.2	62	——	
1212	"	1.8	0.4	-3.9	-8.5	47	52	
1184a	"	2.5	3.0	1.4	0.6	96	55	
842	"	2.9	3.2	——	——	100	——	
1242b	"	2.6	3.1	——	——	100	51	
1243a	"	3.0	3.2	2.4	1.4	85	——	
1243b	"	2.8	3.2	——	——	100	——	
1171	"	2.8	3.4	2.9	2.8	98	52	
1215	"	2.9	1.8	1.9	-0.5	87	51	
1240	"	2.2	2.8	——	——	100	——	
75	"	2.5	1.0	-2.6	-7.7	64	——	
1247	"	2.5	3.1	——	——	99	——	
1230	"	2.2	2.7	——	——	100	——	
1213	"	——	——	0.7	-2.7	94	——	
1185	"	2.7	2.8	——	——	100	——	
1193	"	2.1	2.0	——	——	100	——	
1201a	" around mold	2.4	1.6	-1.0	-4.6	64	——	
1201b	" " "	2.2	1.0	-3.7	-8.7	29	——	
92	" " "	2.6	1.6	-2.9	-7.5	56	——	
52	algal crust	1.4	0.1	-1.1	-4.0	64	——	
1209d	dolostone	2.5	1.9	-0.4	-3.3	77	——	
1184b	caymanite	2.0	1.8	——	——	100	——	
1062a	coarse spar	-3.6	-7.5	-1.2	-2.5	10	——	
283	dark dolostone	——	——	-2.9	-6.2	56	——	
1223a	dark calcite spar	——	——	-5.1	-8.9	0	——	
843	"	——	——	-5.6	-6.3	0	——	
1208a	porous, grey fill	——	——	-2.7	-10.3	0	——	
1208b	laminated micrite	——	——	-2.0	-9.7	0	——	
1208c	grey fill	——	——	-2.4	-10.4	17	——	
1209b	laminated micrite	——	——	-3.1	-8.8	45	——	
1209c	brown micrite	——	——	-4.6	-10.3	19	——	
1297c	dolostones from	2.8	3.3	——	——	100	——	Olig.- Mioc
1297d	Cayman Brac	2.8	2.7	——	——	100	——	

Table 10. Oxygen and carbon isotope compositions, and dolomite nonstoichiometry data for the Bluff Formation dolostones of Grand Cayman Island and Cayman Brac.

original isotopic composition of the dolomite should be preserved (Berner, 1965). Land (1980, 1983a, 1983b) however, argued that recrystallization during burial diagenesis could influence the isotopic composition. In the Bluff Formation, there is no evidence of any burial of the rocks, and thus the dolomites probably have not been extensively recrystallized. The oxygen isotopic values of these dolomites could, therefore, reflect the original composition of the dolomite.

Carbon Isotopes

Dolomites may retain the $\delta^{13}C$ signature of the precursor sediment, rather than establishing a new one from the diagenetic solutions (Degens and Epstein, 1964; Land *et al.*, 1975; Land, 1985). However, Sears and Lucia (1980) argued that the carbon signature comes from the diagenetic fluids that formed the dolomites. Magaritz (1985, p. 122) concluded that the "...carbon isotope record of dolostones is preserved even in cases where large quantities of meteoric water were flushed through the rocks..." Therefore, once formed, the dolomites retain their compositions during shallow diagenesis, although it is unknown what these values represent. It is noteworthy that few authors have attempted an interpretation of the carbon values of dolomite (Land, 1980).

Comparison of Caymanian Dolomite Values with Modern Dolomites and $CaCO_3$ Sediments

The Caymanian dolomites are enriched in both ^{18}O and ^{13}C compared to Holocene mixing zone dolomites. Mixing zone dolomites from Jamaica have average oxygen and carbon values of -1.0 ‰ and -8.4 ‰, respectively (Land, 1973b); those from Israel have average oxygen and carbon values of -1.3‰ and -9.6‰, respectively (Magaritz *et al.*, 1980). Both t[illegible]^{18}O and $\delta^{13}C$ values of the Caymanian dolomites are slightly lower than Holocene hypersaline dolomites of Abu Dhabi (McKenzie, 1981), which have an average $\delta^{18}O$ value of +1.8‰ and an average $\delta^{13}C$ value of +2.8‰. The $\delta^{18}O$ values of the Caymanian dolomites are also lower than Holocene hypersaline dolomites of Baffin Bay, whereas the $\delta^{13}C$ values are equal to or slightly higher than these: The Baffin Bay dolomites have an average $\delta^{18}O$ value of +4.8‰ and $\delta^{13}C$ value of 0.9‰ (Behrens and Land, 1972) and range from $\delta^{18}O$ values of +4.8 to +5.4‰ and $\delta^{13}C$ values of 1.6 to -7.5‰ (Kier, 1973). The Caymanian dolomite values thus fall between mixing and hypersaline dolomites, with respect to their $\delta^{18}O$ and $\delta^{13}C$ values.

Extrapolation of high temperature experimental data on dolomite - calcite equilibrium fractionations predict that dolomite should be enriched in ^{18}O and ^{13}C compared to co-existing calcite at low temperatures (Northrop and Clayton, 1966; O'Neil and Epstein, 1966; Fritz and Smith, 1970; Sheppard and Schwarcz, 1970). The highest estimate of ^{18}O enrichment is 7 permill and of ^{13}C is 2.5 permill. It has also been documented that many co-existing dolomite - calcite pairs exhibit no fractionation (Degens and Epstein, 1964). Dickson and Coleman (1980, p. 117) argued that "Dolomitization with textural preservation leads to isotopic preservation and hence no fractionation". However, this idea has been challenged and a more reasonable $\Delta^{18}O$ ($\delta^{18}O_{dolomite}$ - $\delta^{13}C_{calcite}$) of 2 to 4 permill has been suggested (Land, 1980). Dolomite may only be slightly enriched in ^{13}C compared to co-existing calcite (Land, 1980) or there may be no fractionation at all (McKenzie, 1981).

Minimum and maximum values of $\delta^{18}O$ for the Caymanian dolomites are +1.4‰ and +3.3‰, respectively. Considering a fractionation value of about 3 permill, the dolomites should be in equilibrium with calcite having a $\delta^{18}O$ of -1.6 to +0.3‰. Calcites having this composition are in equilibrium with seawater (Hudson, 1977, fig. 1, p. 638). Holocene $CaCO_3$ sediments from lagoons of normal salinity around Grand Cayman Island have compositions which range from $\delta^{18}O$ of -1.9 to +1.0‰ and $\delta^{13}C$ of +1.6 to +3.6‰ (Fig. 11, Table 11). The $\delta^{18}O$ values of the Caymanian dolomites, assuming a fractionation of 3‰ (-1.6 to +0.3‰), are thus, similar to the normal marine sediments of

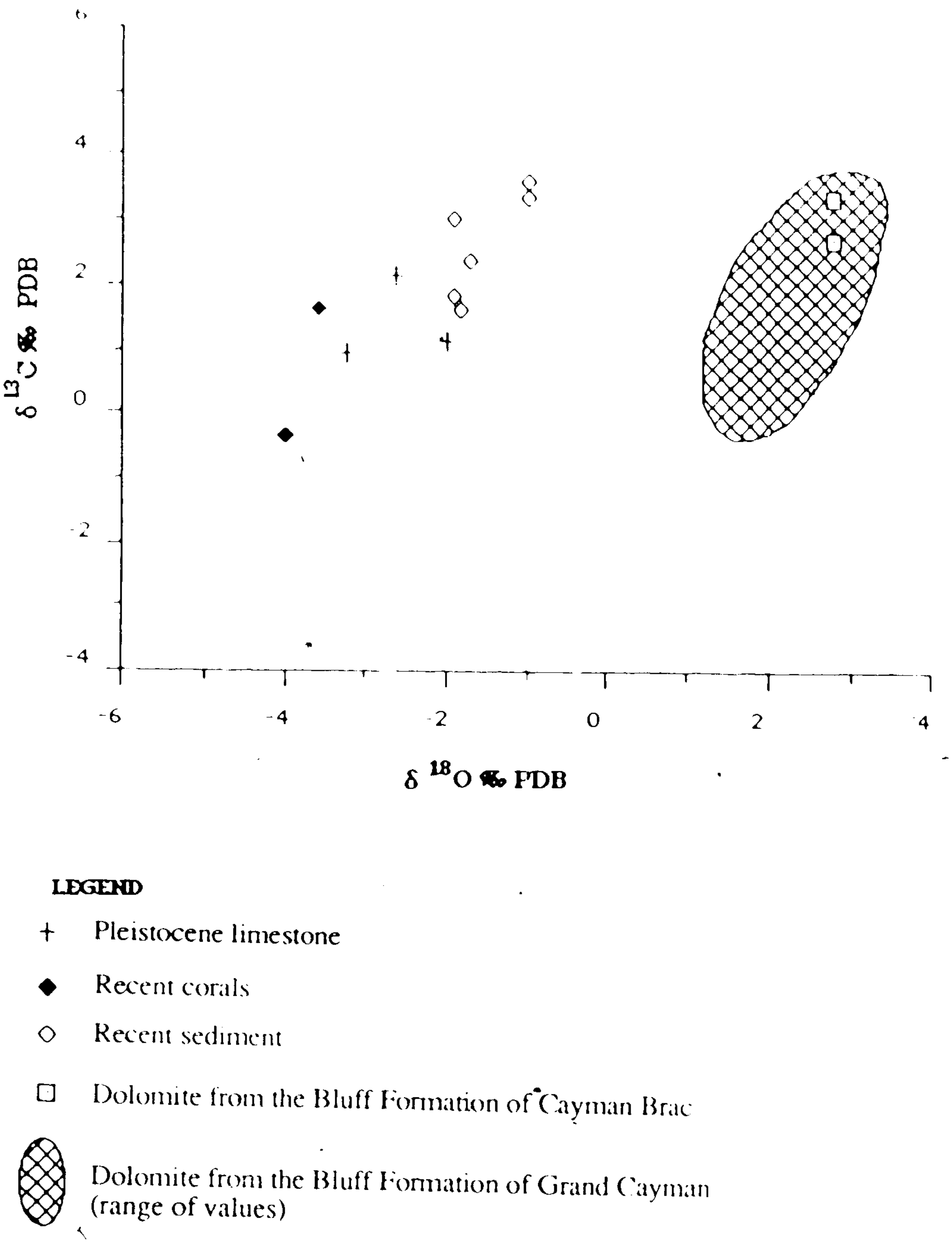


Figure 11. Oxygen and carbon isotope values of Pleistocene and Recent carbonates of Grand Cayman Island, compared with the Bluff Formation dolomite values of both Grand Cayman Island and Cayman Brac.

SAMPLE	DESCRIPTION	calcite $\delta^{18}O$ ‰ PDB	calcite $\delta^{13}C$ ‰ PDB	% dolomite	AGE
817	oolitic limestone	-2.6	2.1	0	Pleistocene
1342a	rhyzolith	-2.0	1.1	0	
1343	rhyzolith sediment	3.2	0.9	0	
01	*M annularis*	-4.0	-0.4	0	RECENT
02	*D. strigosa*	-3.6	1.6	0	
1401	patch reef sediment	-1.7	2.4	0	
1402	sediment behind reef	-1.9	1.8	0	
1408	sed. around *Porites*	-1.8	1.6	0	
1409	*Callianassa* mound	-1.0	3.6	0	
1415	groove sediment	1.0	3.3	0	
1416	*Halimeda* sand	-1.0	3.6	0	
660	beachrock	-1.9	3.0	0	

Table 11. Oxygen and carbon stable isotope data for the Pleistocene and Recent carbonates from Grand Cayman Island.

Grand Cayman (Fig. 11). The $\delta^{13}C$ values of the Bluff Formation dolomites are typical of normal marine carbonates and fall within the range of marine sediments of Grand Cayman (Fig. 11). The range in the carbon values is small (-0.3 to +3.4‰); Fritz and Smith (1970) have argued that a narrow range of ^{13}C values precludes the participation of freshwaters in the formation of secondary dolomites. Certainly, had the dolomites formed in the mixing zone, a greater range in values would be expected due to the combined influence of the dissolving precursor limestone, rainwater, soil horizons and seawater. The $\delta^{18}O$ and $\delta^{13}C$ values of the Caymanian dolomites are therefore compatible with normal marine waters.

Isotopic Comparison with Ancient Dolomites

Mixed seawater-freshwater, normal marine, and hypersaline waters have all been inferred to account for ancient dolomites not obviously associated with evaporites. Many of these ancient examples have similar $\delta^{18}O$ and $\delta^{13}C$ values to those of the Bluff Formation (Table 12).

Salinity and Temperature of the Dolomitizing Solutions

Comparing the oxygen and carbon isotopes of the Caymanian dolomite to both modern and ancient examples of dolomite highlights the uncertainties of interpreting such geochemical data. Waters of all salinity have been regarded as possible dolomitizing solutions for ancient dolomites. Comparing the isotopic compositions of the Bluff [illegible]rmation dolomites with these ancient examples is therefore of little value in determining the salinity of the dolomitizing solution.

More meaningful comparisons can be made with modern occurrences of dolomite. The $\delta^{18}O$ and $\delta^{13}C$ values of the Caymanian dolomites fall into the range of normal marine $CaCO_3$ (assuming an oxygen fractionation factor between co-existing calcite and dolomite of about 3 permill) and have slightly lower $\delta^{18}O$ values than hypersaline dolomite. The oxygen and carbon isotopic compositions of the Caymanian dolomites are thus most

LOCATION & AGE	average $\delta^{18}O$ ‰ PDB	average $\delta^{13}C$ ‰ PDB	INTERPRETATION	REFERENCE
Jamaica - Pleistocene	+2.2	+1.1	mixing zone dolomite	Land, 1973a
Bonaire - Pliocene	+2.0 - +4.1 (range)	-----	mixing zone dolomite	Sibley, 1980
Yucatan - Pleistocene	+1.7	+2.1	mixing zone dolomite	Ward & Halley,1985
Bahamas - Neogene	+2.1	+1.8	unknown - mixing ?	Supko, 1977
Bikini and Eniwetok Atolls - Tertiary	+2.0	+1.7	hypersaline	Gross & Tracey, 1966
Illinois Basin - Mississippian	+1.9	+2.5	hypersaline	Choquette, 1968
Enewetak Atoll - Eocene	+2.5	+2.3	deep, normal marine	Saller, 1984
Niue, S. Pacific - Tertiary	+2.8	+2.0	deep, normal marine	Aharon *et al.*, 1987
Israel - Cretaceous	-0.3	-0.5	shallow, normal marine	Sass & Katz, 1982
Grand Cayman - Oligocene-Miocene	+2.4	+2.0	shallow, normal marine - slightly hypersaline	This Study

Table 12. Table showing the similarities of oxygen and carbon stable isotope data of some ancient dolomites not associated with evaporites and the dolomites of the Bluff Formation.

compatible with solutions having strong affinities to seawater. The oxygen isotopic composition of Caribbean seawater has been determined to be from 0‰ SMOW (Schuffert, 1985) to +1.4 to +1.5‰ (this study).

Using the temperature equation $10^3 \ln \alpha$ dol - $H_2O = 3.2\ (10^6)\ T^{-2} - 3.3$ (Land, 1983b) to predict the relationship between dolomite, water and the temperature of formation, a number of possibilities are evident (Fig. 12):

1. Using the minimum and maximum dolomite values in the equation, the dolomite could have formed from normal marine water (0 to +1.5‰ SMOW) at temperatures of 21° to 35°C. Using the average dolomite value, the temperature range would be 25° to 31°C.
2. Using the minimum and maximum dolomite values, the dolomite could have formed from freshwater or mixed waters (-4‰ SMOW, this study) at temperatures of 6° to 13°C. Using the average dolomite value, the dolomite could have formed at 9°C.
3. Using the minimum and maximum dolomite values, the dolomite could have formed from hypersaline waters (+2 to +4‰ SMOW) at temperatures of 29 to 48°C. Using the average dolomite value, the dolomite could have formed at 33° to 43°C.

Seawater temperatures in the lagoons around Grand Cayman are about 27° to 32°C; temperatures taken of the mixing zones beneath the island range from 27° to 31°C (Ng, pers. comm. 1987). Thus, freshwaters can be precluded in the formation of the dolomite, as the temperatures required are too low (Fig. 12). Cold, deep, seawater can also be precluded, as most of the dolomites would fall into the range of warmer temperatures at this salinity. The lack of evaporites in the Bluff Formation precludes very hypersaline waters and high temperatures.

The only possible dolomitizing waters are, therefore, warm (shallow) normal saline to slightly hypersaline waters: (1) Using the minimum and maximum dolomite values, the dolomite could have formed from normal saline to slightly hypersaline waters, assuming 0

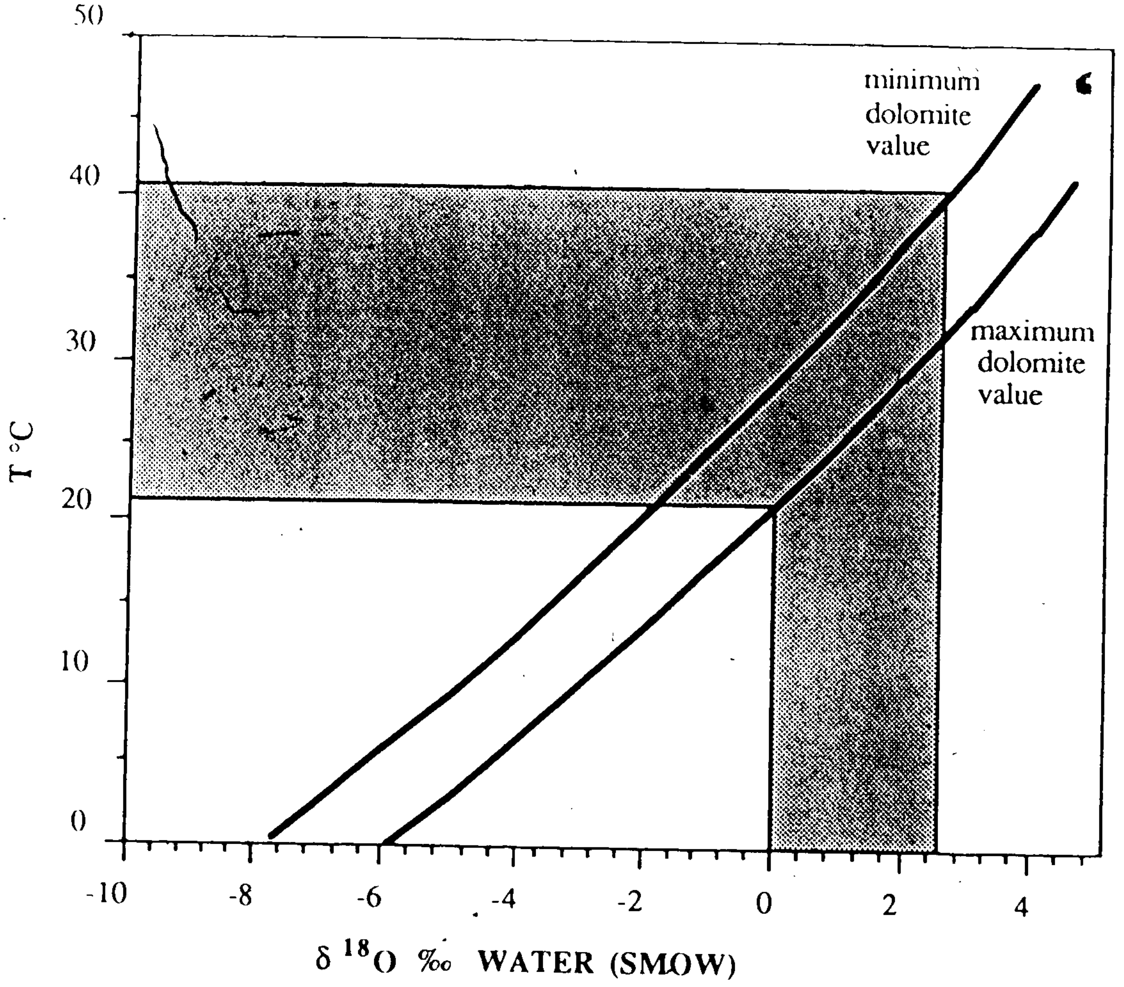


Figure 12. Relationship between dolomite oxygen isotopic compositions, and water composition and temperature in which the dolomites may have formed. The temperature equation of Land (1983b) was used (appendix II).The stippled area indicates possible temperatures and water compositions responsible for dolomitization of the Bluff Formation, using the minimum and maximum dolomite (SMOW) values.

to +2.5‰ SMOW for the waters as possible end members (any higher than +2.5‰ would be in the range of sabkha waters), at temperatures ranging from 21° to 41°C. (2) Using the average dolomite value (+33.3‰ SMOW), the dolomite could have formed at 25° to 36°C. This is in good agreement with the temperatures recorded from the island, and with comparisons made to the isotopic compositions of recent carbonates.

Range of Isotopic Compositions of the Dolomites

The most ^{18}O and ^{13}C enriched dolomites occur in those Caymanian dolostones that are hard, less porous and have less calcite cement (Table 10). These dolostones have both the cloudy centres and clear rims of the zoned dolomite preserved. The dolomites that have low $\delta^{18}O$ and $\delta^{13}C$ values occur in those dolostones from localities exhibiting an abundance of soft, porous patches, but locally have more calcite cement (Table 10). The soft, porous patches contain only the clear rims of the dolomite rhombs, as a result of the preferential dissolution of the cloudy centres. This behaviour could be explained by two processes, although it is not possible to distinguish which:

(1) The presence of cloudy centres *and* clear rims preserved in the dolomites with the highest $\delta^{18}O$ and $\delta^{13}C$ values. A higher proportion of cloudy centres may have made the $\delta^{18}O$ and $\delta^{13}C$ values higher, i.e. the cloudy centres precipitated from more saline waters than the clear rims of the dolomite rhombs. Conversely, the least isotopically enriched dolomites occur in those rocks exhibiting extensive selective dissolution of the cloudy cores. The lighter values thus reflect the higher proportion of clear rims that precipitated in more dilute solutions. Both Folk and Siedlecka (1974) and Folk *et al.*(1975) argued that in zoned dolomite rhombs, the clear rims precipitated from more dilute solutions than the cloudy cores, and were thus more stable in fresher waters.

(2) Contamination of dolomite by calcite. The dolomites with the highest $\delta^{18}O$ and $\delta^{13}C$ values occur in rocks with the least amount of calcite cement; the lowest values

are found in those with the most calcite cement. This calcite is isotopically lighter than the dolomites (Fig. 10). Due to the procedure followed for physically inseparable dolomites and calcites (Walters *et al.*, 1972), some contamination is likely.

Late Diagenetic Dolomite Cement

One sample of dolomite cement (PBQ 1062a) was large enough to be independently analysed (from the matrix). This dolomite is larger and clearer than the type 1 and 2 dolomite cements, and petrographic evidence suggests it is a later diagenetic product, as it is rooted on calcite spar. Both the oxygen and carbon values (-3.6‰ and -7.5‰, respectively) are much lower than any of the other dolomites sampled (Table 10, Fig. 10). Contamination from the calcite spar in this sample is precluded as this calcite (PBQ 1062a) is isotopically more enriched than the dolomite. This dolomite cement falls well into the field of freshwater carbonates (Hudson, 1977), especially if a fractionation in oxygen of about 3 permill is considered. The large crystal size (100 μm) and clear nature indicates slow precipitation in dilute solutions (Folk and Land, 1972, 1975). This therefore supports the idea of freshening porewaters during the evolution of the dolomite and calcite cements.

B. ISOTOPIC SIGNATURES OF CALCITE IN THE BLUFF FORMATION

The calcite values exhibit a covariant trend in $\delta^{18}O$ and $\delta^{13}C$ toward more negative values (Fig. 10). This positive covariance is characteristic of meteoric diagenesis (Allan and Matthews, 1977; 1982; Allan *et al.*, 1978; James and Choquette, 1984). The progressively less enriched $\delta^{18}O$ values reflect the increasing involvement of freshwaters derived from ^{16}O enriched rainwater. The $\delta^{13}C$ values show a much greater range in values due to the influence of isotopically light carbon from soil gas CO_2 toward the vadose zone, and the rapid equilibration of the carbon towards the carbon composition of the host rock away from this zone (Lohmann, 1982).

The calcite values in any given dolomite-calcite sample pair, are consistently lighter than the dolomite values. An overlap between the calcite and dolomite values on Figure 10 is due to either some cross contamination, or to the fact that some of the dolomites and calcites precipitated from isotopically similar solutions.

Sparry Calcite Cements

The poikilotopic and sparry calcite cements have a wide range of isotopic values ($\delta^{18}O$ of +2.9 to -3.9‰ and $\delta^{13}C$ of +2.8 to -8.6‰). It is extremely unlikely that temperature variations alone could produce the observed range in $\delta^{18}O$ values. If a constant H_2O composition of 0 permill (SMOW) is assumed, a range in temperature of more than 31°C is required to account for all the calcite values (using the equation of Friedman and O'Neil, 1977). This variation is unrealistic in view of the shallow diagenetic environments in which the Bluff Formation rocks have evolved.

The poikilotopic and sparry calcite cements thus record porewaters with progressively lower $\delta^{18}O$ and $\delta^{13}C$ values. The strong covariance of the carbon and oxygen isotopes is characteristic of mixing zone diagenesis, in which the $\delta^{18}O$ and $\delta^{13}C$ values become isotopically depleted as salinity decreases across the mixing zone (Moore, 1973; Allan and Matthews, 1977, 1982; Allan *et al.*, 1978). Moore (1973) documented a positive oxygen and carbon isotopic covariance for carbonate cements from a mixing zone beneath a beach on Grand Cayman Island. A similar trend would be expected in a mixing zone beneath a freshwater lens (Allan and Matthews, 1982).

The most isotopically enriched sparry calcites tend to be the smaller crystals (50-100 µm), while the least enriched are the largest cement crystals (up to 2 mm). Large equant crystals are typical of slow precipitation in dilute solutions (Folk and Land, 1972, 1975; Folk, 1974). The light values of these large crystals are typical of freshwater limestones (Gross, 1964; Keith and Weber 1964; Hudson, 1977).

The calcite cements that are not demonstrably poikilotopic probably reflect the irregular distribution of dolomite dissolution. Where the calcites line a cavity surrounded by tightly bound dolomite crystals, then the calcite is not able to take on a poikilotopic texture, even though this calcite may be intimately related (i.e. precipitated from the same solution as) the other poikilotopic calcites.

Petrographic evidence indicates that these calcites are later than both dolomitization and dolomite dissolution. It has been suggested that the process of dolomitization records porewaters becoming increasingly less enriched in ^{18}O and ^{13}C. Calcite precipitation began in waters isotopically similar to those that precipitated some of the dolomites, and continued as waters became increasingly fresh. Thus, calcite cementation began in waters of mixed meteoric and marine origin (mostly marine), after most of the pervasive dolomitization, and continued through the mixing zone into the freshwater phreatic zone.

Micrite Patches and Bodies

The lowest ^{18}O values for calcite are represented by the micrite patches and laminated nodules and elongate bodies ($\delta^{18}O$ of -2.0 to -4.5‰, and $\delta^{13}C$ of -8.7 to -10.4‰). These fall into the range of flowstone compositions in the Bluff Formation (Smith pers. comm. 1987) and vadose calcites of elsewhere (e.g. Allan and Matthews, 1977; James and Choquette, 1984). Petrographic evidence also suggests that these are vadose features. Some of the micrite patches show cross-cutting relationships between them, which indicate a progressive freshening of waters.

The low $\delta^{13}C$ values are derived from atmospheric carbon dioxide brought in by rainwater, and ^{12}C enriched soil carbon dioxide. The ^{16}O enrichment of fresh rainwater causes the low $\delta^{18}O$ values. If, as is suspected, the micrite has replaced the host dolostone, then the isotopic signatures of the former dolostone has been lost.

C. SYNOPSIS

1. The dolomites have average PDB values of $\delta^{18}O$ of +2.4‰ and $\delta^{13}C$ of +2.0‰.
2. The dolomites have higher $\delta^{18}O$ and $\delta^{13}C$ values than modern mixing zone dolomites.
3. The dolomite $\delta^{18}O$ values are only slightly lower than those of modern hypersaline dolomites. The $\delta^{13}C$ values of the Caymanian dolomites are about the same as these.
4. Accepting a fractionation value of 2 to 4 permill for $\Delta^{18}O_{(\text{dolomite}-\text{calcite})}$, the dolomite isotopic compositions are typical of normal marine carbonates.
5. The oxygen isotope data for the dolomites indicate that warm (shallow) normal marine to slightly hypersaline marine waters were responsible for dolomitization.
6. Porewaters may have become increasingly depleted in both $\delta^{18}O$ and $\delta^{13}C$ as the precipitation of dolomite rhombs proceeded from the cloudy centres to the clear rims.
7. The calcites (which formed after dolomitization) exhibit a covariant trend towards increasingly ^{18}O depleted porewaters. The sparry calcite isotopic compositions are characteristic of marine-meteoric mixing to freshwater phreatic diagenesis, while the $\delta^{18}O$ and $\delta^{13}C$ values of the micrite patches and nodules are typical of vadose diagenesis.

V. DIAGENETIC INTERPRETATION

A. SUBMARINE CEMENTATION

Examples of submarine diagenesis, which are rare in the dolostones of the Bluff Formation, consist of peloids, micrite cement, isopachous cements, and early lithification of algal coatings. The paucity of submarine cements may be a function of (1) the original sparsity of submarine cementation, or (2) an artifact produced by extensive recrystallization of the Bluff Formation. The former seems the most probable since the observed submarine cements are very well preserved. This paucity of submarine cements has important implications regarding the depositional environment of the Bluff Formation.

Peloids and Micritic Cement

Peloids can form by (1) chemical precipitation by repeated nucleation around centres of growth (Macintyre, 1977; 1984; 1985), (2) spontaneous chemical precipitation (Marshall, 1983), (3) the pelletizing action of organisms (Macintyre *et al.*, 1968), (4) replacement of aragonite faecal pellets (Shinn, 1969), (5) fragmentation of calcified algal filaments (Schroeder, 1972), (6) passive infill of interstices by detrital lime silt grains (James *et al* .,1976) and/or (7) precipitation induced by bacteria (Chafetz, 1986).

The peloids in the Bluff Formation are now composed of dolomite (Plate 10B). Some at least are a result of the breakdown of calcified filaments. Dolomitized filaments, presumably once calcified, associated with peloids are apparent in some cavities (HRQ 1208). Most of the peloids however, occur in coral cavities, where they are embedded in a finely crystalline dolomite matrix. The peloids are much denser than the ones associated with the filaments. The dolomicrite may represent an original submarine cement. Fine dolomicrite also occurs in coral cavities without peloids. The dolomicrite is commonly geopetal and may be capped by isopachous cement (Plate 10D). Both micritic cement and micrite are ~~common cavity fill~~ers in recent reefs (Friedman, 1985). HMC peloids in recent reefs commonly occur in coral interstices associated with micritic cement. Whatever the origin of the peloids in cavities in the Caymanian dolostones, it is probable that they were cemented in place in the submarine environment.

Isopachous Cement

Isopachous cements are well developed on replaced coral structures (Plate 10B-E). These cements have been dolomitized, but the syntaxial orientation of the cements and coral substrate have been preserved. The original coral surface is defined by the presence of a dark dolomicritic outline (Plate 10D). Talbot (1972) recognised that a replaced coral skeleton may be outlined by minute inclusions, which appear as a dark micritic rim. In the Caymanian examples, the cements extend into and beyond this dolomicritic outline, especially when viewed under crossed nichols. This type of optically continuous cementation is typical of acicular aragonite cementation (Macintyre, 1977; Lighty, 1985) which is contemporaneous with active coral growth (Hubbard, 1975; Macintyre, 1977). Some of the isopachous cement lines the coral cavities as well as coats them. This cement is also interlayered with internal sediment in the cavities, all of which has subsequently been dolomitized. The cements can thus be interpreted to represent submarine cementation.

Algal Crusts

The red algal drapes coating many of the coral molds are commonly thickened on one side of the mold (Plate 10F). Entobian borings cut through the algal laminations, as well as the dolomicrite in the crusts. The crusts were therefore lithified in the submarine environment. Similar submarine coatings have been documented by Land and Goreau (1972) and Macintyre (1977). The borings do not cut through the calcite spar which occurs in the crusts. The calcite is therefore younger than the borings.

Implications

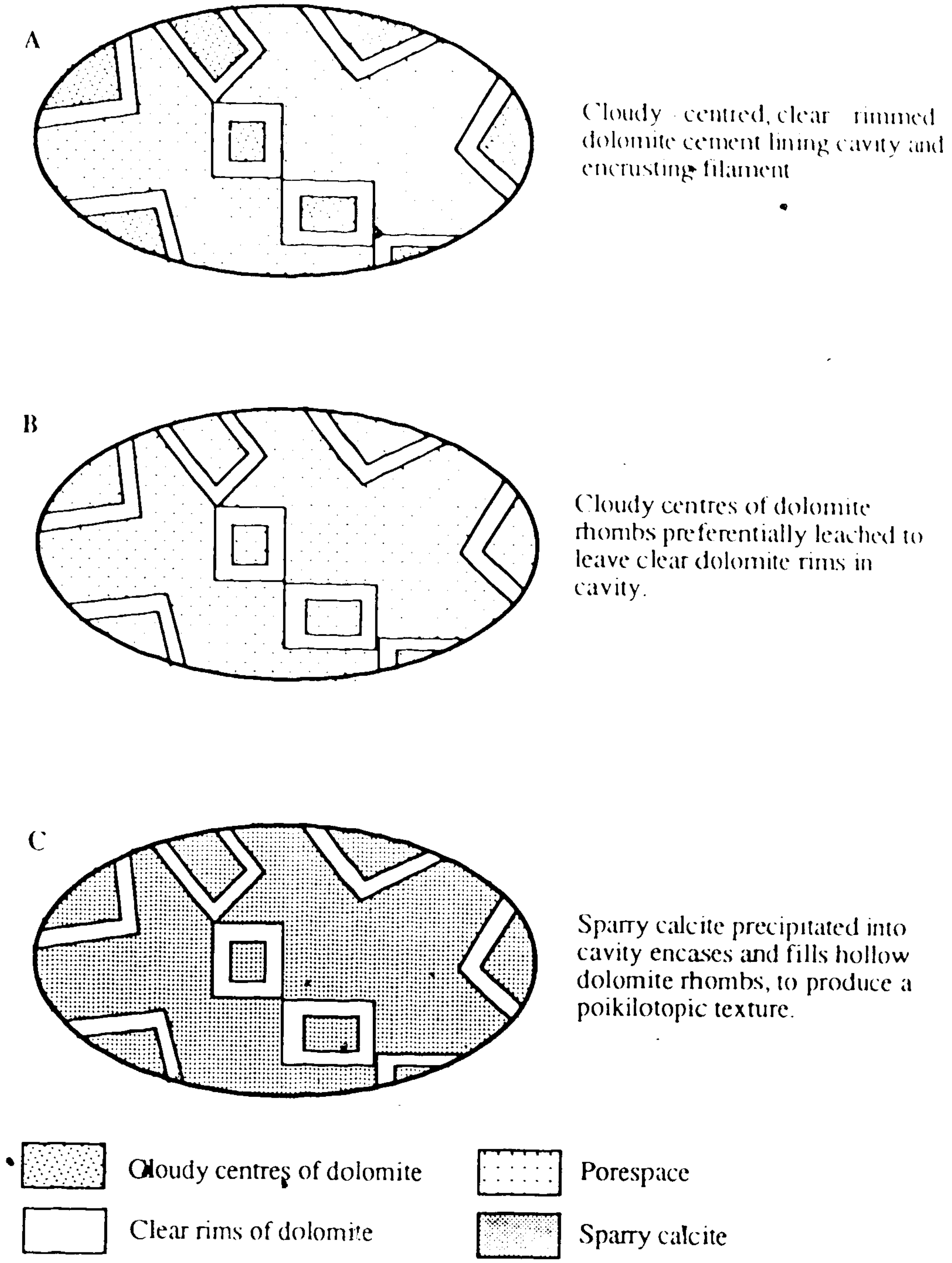
Submarine cementation, especially HMC cementation, is much reduced in areas of low agitation and fine grained sediments (Ginsburg *et al.*, 1971; James *et al.*, 1976). The sparsity of submarine cements in the Bluff Formation therefore could attest to the low energy conditions during deposition. The absence and sparsity of submarine cements has been noted in protected back reef areas of fringing and barrier reefs (James *et al.*, 1976; MacIntyre, 1977), in contrast to the high abundance of cementation on the seaward margins of shelf-edge reefs. However, Marshall (1986) expressed reservations on the use of submarine cements as palaeoecological tools.

B. GENESIS OF POIKILOTOPIC TEXTURES

The development of the poikilotopic textures, wherein calcite spar encases and fills hollow dolomite rhombs, has involved temporally distinct diagenetic stages (Fig. 13). The growth of cloudy-centred clear-rimmed dolomite was a prerequisite for the evolution of this texture. The cloudy centres preferentially dissolved leaving hollow dolomite rhombs (Plate 11A, B; 12A-C; 13A, B). In some areas extensive dissolution of the dolomite occurred resulting in a soft, friable dolostone (Plate 13E). The distribution of this soft dolostone appears to have been generated by circulating groundwaters, and thus intimately linked to the distribution of the permeability conduits, in particular, the joint systems (Ng, 1985a).

After some dissolution of the dolomite, calcite cementation began. The presence of rhomb shaped voids in calcite crystals, however, indicates that dolomite dissolution continued after calcite precipitation had begun. Calcite spar filled the pore spaces around and in some of the hollow dolomite rhombs, thereby generating poikilotopic textures (Plate 10G; 12D-F; 13D-F). Less calcite cementation occurred than dolomite dissolution , resulting in many porous dolomite areas devoid of calcite cement (Plate 13E).

Folk and Siedlecka (1974, p. 11) suggested that cloudy-centred clear-rimmed dolomite crystals were due to different types of dolomite forming from waters of differing salinity: "...an initial dolomite crystal formed under conditions of higher salinity and as time passed the water freshened and an overgrowth of limpid dolomite was formed on the



Cloudy - centred, clear - rimmed dolomite cement lining cavity and encrusting filament

Cloudy centres of dolomite rhombs preferentially leached to leave clear dolomite rims in cavity.

Sparry calcite precipitated into cavity encases and fills hollow dolomite rhombs, to produce a poikilotopic texture.

Figure 13. Evolution of the poikilotopic textures of the Bluff Formation.

dirtier nucleus." Since the limpid rims are more stable than the cloudy cores, in waters of lower salinity, the unstable cores may be preferentially leached. Land *et al.*(1975) also argued that in zoned dolomite rhombs, the clear rims precipitated from more dilute solutions than the cloudy cores. Such fluctuating salinities are characteristic of the freshwater-seawater mixing zone. In addition, the type 3 dolomite calcite alternating cements also indicates fluctuating water chemistry, that is also indicative of the mixing zone (Jones *et al.*, 1984).

The largest of the calcite cements are typical of crystals formed in pores completely filled with water in the freshwater phreatic zone (Folk *et al.*, 1973). These large crystals (up to 2 mm) are associated with areas of greatest dolomite dissolution, and are thus considered the latest cements in the evolution of the poikilotopic textures. Therefore, the final stage in the development of these textures occurred in the freshwater phreatic zone (Fig. 13).

Isotopic evidence further elucidates this diagenetic sequence. The calcites show a positive covariant trend in the oxygen and carbon values, which is typical of a mixing zone trend, in which waters became increasingly fresh. Those calcites most enriched in ^{18}O were precipitated in waters similar to those in which the dolomites formed, probably of mostly marine origin. The less ^{18}O-rich calcite precipitated from progressively less saline waters of the mixing zone. The isotopically lightest calcites are also the largest crystals, and have isotopic compositions typical of freshwater calcites. The dolomite with isotopically very light values (PBQ 1062a), rooted on the calcite spar, is indicative of the freshest waters of the mixing zone or of the freshwater phreatic zone.

The evolution of the poikilotopic textures, therefore, records a trend of progressively freshening porewaters, which marks the rocks relative movement through the mixed to freshwater phreatic diagenetic environments. This is a regressive cement sequence, which could reflect the gradual emergence of the island above sea level.

Jones *et al.* (1984) showed that many of the cavities in the dolostone and in some of the caymanite in the Bluff Formation were filled or partly filled with three generations of cement. The first generation limpid dolomite cement probably formed in the mixing zone, the second generation sparry calcite probably formed in the freshwater phreatic zone (Folk *et al.*, 1973; Jones *et al.*, 1984). The third generation microstalactitic cement formed in the vadose zone. This succession of cements is similar to the sequence proposed for the evolution of the poikilotopic textures. Thus, it can be suggested that the two sequences correlate. Indeed, Jones *et al.* (1984) also interpreted their regressive sequence of cements to represent the net uplift of the island. It is apparent then, that caymanite sedimentation must have begun very soon after the island began to emerge, as this also hosts the regressive sequence of cements in some examples (Plate 13C). In the examples documented by Jones *et al.* (1984) a period of internal sedimentation separated the formation of limpid dolomite from that of the sparry calcite. There is, however, no evidence of internal sedimentation in the poikilotopic textures described herein. Instead, a period of dolomite dissolution separated the formation of dolomite from the formation of calcite.

Implications

It has been suggested that: (1) textures in which dolomite rhombs appear to be floating in patches of calcite, and (2) dolomite rhombs that have a core of calcite, are indicative of dedolomitization (Shearman *et al.*, 1961; Folkman, 1969; Warrak, 1974). The calcite filling and encasing the hollow dolomite rhombs in the Bluff Formation can therefore be regarded as dedolomitization textures. Dedolomitization has generally been regarded as a single process involving the concomitant dissolution of dolomite and precipitation of calcite (e.g. von Morlot, 1847; Yanat'eva, 1955; Chilingar, 1956; de Groot, 1967; Al-Hashimi and Hemingway, 1973; Longman and Mench, 1978). Smit and Swett (1969) however, argued that dedolomitization involves any process(es) that result(s) in dolomite being partly dissolved or replaced by any other mineral. Teall (1909) used the term for dolomites that had been transformed to Mg silicates during metamorphism. Similarly, Faust (1949) used the term for the dissociation of dolomite to periclase and calcite.

The dedolomites in the poikilotopic textures of the Bluff Formation evolved through a series of temporally separated processes: (1) the precipitation of zoned dolomite rhombs, (2) partial dissolution of the rhombs to produce hollow dolomite rhombs, and (3) precipitation of calcite which fills and encases these hollow rhombs (Fig. 13). Thus, textures in which dolomite rhombs appear to be floating in calcite, and dolomite rhombs that have a core of calcite, are not solely the product of concomitant dolomite dissolution and calcite precipitation. The former texture can be produced by dolomite crystals attached to filaments in cavities especially if the thin sections are cut obliquely to these filaments. The latter texture can be produced by passive cementation within hollow dolomite rhombs.

This demonstrates that dedolomites can form by more than one process. Calcitization (Smit and Swett, 1969) is a better term and would restrict 'dedolomites' to those that had evolved through the concomitant replacement of dolomite by calcite. However, dedolomitization continues to be used as a matter of convenience, because of its widespread usage in the literature (Evamy, 1967; Zenger, 1973; Warrak, 1974).

Dedolomites formed by the concomitant dissolution of dolomite and precipitation of calcite may be recognized by the presence of (1) rhombic calcite crystals (but see Perkins, 1968), (2) palimpsest textures where the outlines of the original dolomite rhombs are highlighted by iron oxides and other substances (Shearman *et al.*, 1961; Folkman, 1969), and (3) polycrystalline rhombic pseudomorphs of calcite after dolomite, or polycrystalline calcite cores surrounded by a rim of dolomite (Shearman *et al.*, 1961; Evamy, 1967; Folkman, 1969).

The dedolomites in the poikilotopic textures of the Bluff Formation are thus produced by time separated processes, not by the concomitant replacement of dolomite by calcite. They are the result of groundwaters becoming progressively less saline through time, which could represent the rocks passing relatively through the marine to meteoric diagenetic zones. This regessive sequence is recorded from other parts of the Bluff Formation (Folk *et al.*, 1973; Jones *et al.*, 1984), and probably represents the initial (and only) emergence of the island.

It has been suggested that dedolomitization is caused by the reaction of dolomite with solutions containing sulphate (e.g. von Morlot, 1847; Yanat'eva, 1955; Evamy, 1963, 1967; de Groot, 1967; Folkman, 1969; Warrak, 1974; Hanshaw and Back, 1979) or sulphate and calcium (Back *et al.*, 1983). There is no evidence to suggest that sulphate was involved in the dedolomitization process of the Bluff Formation. The dedolomites formed as a response to groundwaters becoming increasingly fresh, as the rocks passed relatively from the marine to the meteoric zones. It has also been suggested that dedolomitization is a near-surface phenomenon (Braun and Friedman, 1970), however, Budai *et al.* 1984 demonstrated that dedolomites could result from deep burial diagenesis. These ideas were based on the assumption that dedolomites were a result of concomitant dolomite dissolution and calcite precipitation. The dedolomites described herein are not a result of this process. Dedolomite formed by time separated processes can form simply by moving the rocks from the marine to meteoric diagenetic environments. This type of dedolomite may therefore be indicative of proximity to unconformities.

C. VADOSE DIAGENESIS

Examples of dissolution features, cements and replacement products formed in the vadose zone, presumably after the development of the poikilotopic textures, are present in the Bluff Formation. These represent the final stages of the regressive sequence of diagenetic features. Since the island's emergence in the middle Tertiary, it has remained largely subaerial (Emery, 1981). Hence, since this time most diagenesis involving the surface outcrops of the Bluff Formation described herein, has taken place in the vadose zone.

Spiky Calcite

The spiky crystals are comparable to those described by Folk *et al* (1985) and Jones and Pemberton (1987a). The spikes of the Bluff Formation calcite crystals are up to 60 µm in length (Plate 14A). Those described by Jones and Pemberton (1987a) and Folk *et al.*, (1985) are only up to 10 µm long. Folk *et al.* (1985) suggested that spiky crystals resulted from both precipitation and dissolution. Jones and Pemberton (1987a) however, argued that they were produced by organically mediated dissolution, largely by the action of fungi. Experiments showed that this organic dissolution occurred even in the absence of large amounts of fluids.

The following observations indicate that the Bluff Formation spiky crystals are a result of dissolution:

1. The development of spikes along cleavage planes, with the spikes developed on either side of the cracks, and which may be interconnected across these cracks (Plate 14B,C) is indicative of dissolution. Similar features (called 'teeth') resulting from dissolution have been found in augite, feldspar, hypersthene and diopside (Berner *et al.*, 1980; Berner and Schott, 1982).
2. The transition from spikes developed along a narrow dislocation zone to a spar crystal completely dissected by spikes (Plate 14B) suggests dissolution rather than precipitation.
3. The irregular and variable sizes and shapes of the spikes on the same dislocation plane is also most likely a dissolution phenomenon.

4. There is no distinct boundary between the spikes and the calcite spar crystal, thus the spikes did not grow off the spar substrate (Plate 14D).
5. The spikes are commonly parallel to the c-axis of the calcite crystal, and developed along a basal cleavage. Calcite dissolves more readily perpendicular to the basal cleavage, i.e. in the direction of the c-axis (Folk *et al.*, 1985).

In dislocation or cleavage planes, etching may be restricted, and concentrated in the direction of the c-axis of the crystal. Prolonged dissolution could produce extensively spiked crystals.

Jones and Pemberton (1987a, 1987b) have emphasized the role of fungi in the dissolution of calcite spar crystals. Although the association of calcified filaments (fungi ?) with the spiky calcite in the Bluff Formation does not preclude this organic role in dissolution, it cannot confirm it. Spiky crystals are inferred to be a vadose feature (Jones and Pemberton, 1987a) and have been found in association with travertines (Folk *et al.*, 1985).

Knobbly Textures

The knobbly textures found on the surfaces of the same spar crystals that have the spikes, are also probably a dissolution feature (Plates 14E, F; 15A). The knobbles (comparable to the 'monticuli' described by Jones and Pemberton, 1987b) are aligned, and this alignment pattern is different on adjacent faces of the rhomb (Plate 14E). The rough knobbly texture may have been produced by discontinuous dissolution over the grain surface. Berner and Schott (1982) suggested that a rough texture produced on augite, was a result of etching attack of the grain surfaces at points of excess energy, e.g. dislocation outcrops. A gradation in degrees of roughness is apparent on the same crystal face (Plate 14E, F). There is also a gradation between those monticuli with large attachment areas to micrite particles on the spar surface. Alexandersson (1972) referred to 'shell-residue micrite' as the cryptocrystalline material remaining after dissolution of larger crystals in shells. The micrite on the spar described herein, is interpreted to originate from the calcite spar crystal, and it is therefore a residue micrite in the sense of Alexandersson. This represents the progressive breakdown of the spar crystal into micrite particles, a process termed 'sparmicritization' (Kahle, 1977) or '*in situ* micritization' (Harrison, 1977).

Etching may produce aligned knobbles or monticuli, which become increasingly pronounced as dissolution proceeds. As the attachment areas are eroded, the monticuli become unstable, and are eventually detached from the host crystal. This produces residue micrite on the crystal surface, which may remain in place or be washed away.

Spikes may develop from the etching of this rough surface of the knobbly textures. Triangular shaped etching patterns are apparent on some of the knobbles (Plate 14F). Indeed, Berner and Schott (1982) suggested that etch pitting could produce large scale features such as 'teeth'. In areas where etching is more restricted, e.g. dislocation cracks, dissolution may produce blunt and pointed spikes in the crack. On the other surfaces of the spar crystal, etching of the whole crystal face produces the rough knobbly texture. Thus, both spikes and knobbly textures are present on the same spar crystal, both of which are a result of dissolution of the host calcite crystal.

Columnar Calcite

The columnar calcite crystals, found in cavities in the Bluff Formation (Plate 15B-D) rooted on a calcite substrate, are comparable to the columnar crystals of Chafetz and Butler (1980) and Kendall and Broughton (1978), which are also formed of aggregates of fibres or crystallites. These crystallites have triangular cross sections, however, the crystals from the Bluff Formation have only been observed in SEM so this feature cannot readily be compared. The crystallites do not appear to have a distinctive triangular shape in cross section (when viewed obliquely on the SEM).

The distinct boundary between the substrate from which these crystals extend, and the base of the crystals themselves (Plate 15B), indicates that these are growth features.

The columnar crystals described by Kendall and Broughton (1978) and Chafetz and Butler (1980) are interpreted to have formed from freshwater, deep in the vadose zone. The columnar crystals of the Bluff Formation may therefore also be interpreted as vadose features.

Calcified Filaments

Calcified filaments have been described from marine, meteoric phreatic and terrestial vadose environments (Schroeder, 1972; Golubic *et al.*, 1975; Kobluk and Risk, 1977; Klappa, 1979a, 1979b). The calcified filaments of the Bluff Formation are encrusted with acicular $CaCO_3$ (Plate 15E, F) and are similar to those described by Klappa (1979) from calcrete horizons. However, similar calcified filaments may also form in other environments. The association of the calcified filaments described herein with other vadose features, e.g. spiky and columnar calcite, supports the idea that these are probably also vadose features. The small size of the filaments and the dichotomous mode of branching are characteristic of those filaments produced by fungi (Bromley, 1970; Klappa, 1979).

Acicular Crystals

Acicular crystals found in cavities in the Bluff Formation consist of lublinite-type crystals, at least two types of needle-fibres, one of which is intimately associated with fine micrite, and more stout acicular crystals that have grown perpendicular to the substrate (Plates 16A-F; 17A-C; 18A). All of these acicular crystals occur in coatings on the borings and in linings in the coral molds.

Buckley (1951) suggested that needle crystals would precipitate if solutions were highly supersaturated. Harrison (1977) suggested that fine micrite would also precipitate readily from supersaturated solutions. He also argued that a localised microenvironment associated with plant roots could provide highly supersaturated solutions conducive to the precipitation of needle fibres. Supersaturated solutions were also thought to be a fundamental requirement for dendritic calcite formation on algal filaments in cavities in a karst breccia from Grand Cayman Island (Jones and Kahle, 1985). It was postulated that the algae themselves may have the ability to extract CO_2 from the water, thus raising the saturation level of the solution. Thus the presence of algae (and/ or fungi) in cavities may also influence the microenvironment quite substantially. Ward (1970, 1974) postulated that needle fibres were actually calcified fungal hyphae. Certainly the size of the needle fibre bundles found in cavities in the Bluff Formation is consistent with the size expected for these.

Needle fibre crystals have been documented largely from vadose environments (Ward, 1970; James, 1972; Steinen, 1973; Harrison, 1977), in which both plant roots, and algae and fungi are common. However, an inorganic origin cannot be totally precluded. The lublinite-type crystals and the irregularly shaped stout crystals are less easily explained by a direct organic origin, although it is not known to what extent organic activity may have modified the microenvironments in which the crystals formed. Lublinite is a cave mineral (Stoops, 1976; Bögli, 1980) and the similarities of the lublinite-type crystals with these, as well as their association with Fe and Mn, suggests that these too may be vadose features. Fe and Mn may be fixed by bacteria in cave environments (Moore, 1981). Mn coatings on karst breccia clasts have been documented from Grand Cayman (Jones and Kahle, 1985).

The acicular crystals are rooted on both calcite spar, micrite and dolomite, and therefore postdate these phases. Although the origin of acicular crystals is problematic, they are most likely a vadose diagenetic feature.

Microstalactitic Cement

The pendulous morphology of the microstalactitic cement (Plates 17D, E; 18B) indicates that this cement formed in cavities not completely filled with water. In the vadose diagenetic zone, water commonly hangs from the bases of grains, and CO_2 loss in this environment can induce calcite precipitation (Müller, 1971). The alternating calcite-dolomite bands of the microstalactites described by Jones *et al.* (1984) were interpreted to represent fluctuating water chemistries in the vadose environment. The microstalactitic cements described herein are rooted on both dolomitic groundmass and limpid dolomite cements. They therefore postdate these phases.

Micritization

The dark micrite occurs in irregular patches and as the main component of laminated, elongate bodies in the dolostone. Isotopic analyses of the micrite patches, nodules and elongate bodies reveals that the micrite is enriched in ^{16}O and ^{12}C (Fig. 10, Table10), i.e. it is isotopically light. These isotopic compositions are similar to the compositions of flowstone from Grand Cayman Island and are typical of vadose diagenetic features documented elsewhere. The association of the micritic elongate bodies with roots penetrating the Bluff Formation rocks, also supports the idea of a vadose origin for this micrite.

The features observed in the micrite indicate that the micrite is a replacive phenomenon, replacing both calcite spar and dolomite. These processes are sparmicritization and dedolomitization respectively, and in this instance are examples of vadose diagenesis.

Dedolomitization: Dedolomitization by the process of concomitant dolomite dissolution and calcite precipitation (von Morlot, 1847) is indicated by the following observations:

1. In dolomitized red algae that contains some dark micrite, the dolomite retains the skeletal texture, whereas the micrite does not. The micrite is thus later than the dolomite.
2. Relic dolomite rhombs appear to be floating in the micrite of the laminated nodules and elongate bodies, especially towards the outer edge of the bodies (Plate 18D). This criterion is however, equivocal.

3. Many of these relic dolomite rhombs have a core of micrite (Plate 18D). This is good evidence of dolomite being replaced by calcite as discussed previously.
4. Many of the dolomite rhombs in the micrite are apparently corroded and embayed by the micrite around the edges of the rhombs (Plate 18D).
5. The contact between the micrite and the dolomitic groundmass is gradational. This is more likely a result of replacement than of passive infill of cavities by micrite.

The micrite is thus later than the dolomite, and at least some of it has replaced the dolomite.

Sparmicritization: Sparmicritization is the process of reducing a sparry calcite crystal to micrite (Kahle, 1977). Several observations indicate that calcite spar has been degraded to micrite:

1. In areas of sparry calcite cementation, irregular patches of micrite may occur in the cement (Plate 18C). The contact between the two is indistinct.
2. The micrite may outline the individual spar crystals. The boundary between the two is fuzzy indicating that the micrite was not passively precipitated onto the spar surface. Evidence has already been presented to indicate that the spar is a cement.
3. In places, the micrite may pseudomorph the spar calcite crystals. This results in a clotted texture, with the individual clots about the size of the spar crystals replaced (Plate 18E). The development of a clotted texture resulting from sparmicritization was also recognized by Kahle (1977) in calcrete profiles of Florida.
4. Micrite that has relaced dolomite may grade into micrite that has replaced calcite spar.

5. SEM examination has revealed the presence of residue micrite on spar surfaces, thought to be a result of spar etching. This process of a spar crystal degrading to micrite by dissolution can also be considered sparmicritization, in contrast to the sparmicritization process of micrite replacing calcite as described in the first four points.

Sparmicritization may therefore cover both processes of (1) spar degrading to micrite by dissolution of the host crystal, and (2) spar being concomitantly replaced by micrite. Kahle (1977) recognized only the latter process. In the Bluff Formation dolostones, both processes are operative, but the latter is much more common.

D. SYNOPSIS

Diagenesis of the Bluff Formation records progressively freshening porewaters, as the rocks were moved relatively from the marine, through the mixing and freshwater phreatic to the vadose diagenetic zones. This probably records the initial net uplift of the island above sea level, in the middle Tertiary. Since this time, the rocks studied herein have remained largely in the vadose environment. The following points summarize the main diagenetic features discussed in this chapter:

1. The Bluff Formation was deposited in a low energy marine environment, probably a back reef setting.
2. The growth of cloudy-centred clear-rimmed dolomite rhombs occurred after pervasive dolomitization, and from progressively freshening groundwaters, after at least part of the island had become emergent.
3. In porewaters conducive to the precipitation of the clear rims, the cloudy centres of the rhombs were leached, as the cloudy centres were more unstable than the clear rims in dilute solutions.

4. Extensive dolomite dissolution occurred in areas adjacent to permeability conduits (e.g. joints).
5. Sparry calcite cementation infilled and surrounded the hollow dolomite rhombs producing a poikilotopic texture.
6. This poikilotopic texture has certain characteristics generally associated with dedolomitization (in the sense of calcite concomitantly replacing dolomite). However, the petrographic evidence indicates that these textures formed through time separated processes of dolomite dissolution and passive calcite cementation.
7. Dedolomitization in the sense of dolomite being concomitantly replaced by calcite can be recognized by the presence of (1) rhombic calcite crystals, (2) palimpsest textures where the outlines of the original dolomite rhombs are highlighted by iron oxides and other substances, and (3) polycrystalline rhombic pseudomorphs of calcite after dolomite, or polycrystalline calcite cores surrounded by a rim of dolomite.
8. Vadose cements may be rooted on either the dolomite or spar calcite, and are thus a later diagenetic phase than both of these. Dissolution of spar crystals was also an important process in vadose diagenesis.
9. Micrite has replaced both dolomite and calcite spar, by the processes of dedolomitization and sparmicritization respectively.
10. The sequence from dolomite cementation, dolomite dissolution, calcite cementation, calcite dissolution to the formation of micrite by dedolomitization and sparmicritization, records increasingly fresh porewaters and probably the net uplift of the Island. This is probably the initial and only emergence of Grand Cayman since deposition of the Bluff Formation.

VI. DOLOMITIZATION

A. ARAGONITE DISSOLUTION

Originally aragonitic skeletal components have been preferentially dissolved in the Bluff Formation, while those originally consisting of HMC have mostly been dolomitized retaining their microtextures. Where originally aragonitic corals have been preserved as dolomite, their texture has been largely obliterated. Sibley (1980) documented similar differential preservation of originally aragonitic and HMC constituents of the Seroe Domi Formation of Bonaire. Corals, pelecypods and gastropods are generally leached, but rarely may be pseudomorphically replaced by dolomite. Red algae exhibit near perfect preservation ranging to mere ghost structures, while foraminifera may be perfectly preserved or dissolved. Ward and Halley (1985) also documented the preferential dissolution of coarser aragonitic components (as opposed to the mud matrix) in the Yucatan dolostones, which occurred prior to and/or during dolomitization.

Many of the fossil molds in the Bluff Formation are lined with limpid dolomite cement. The boundary between the cement and the dolomitic groundmass is distinct, and on the cathodoluminescence microscope, the dolomite cements have dull luminescent cores and brighter outer zones, whereas the groundmass has a brighter orange luminescence (Plate 18F, G). Thus, aragonite dissolved prior to at least the dolomite cementation.

In some cavities there is no distinct boundary between the euhedral rhomb faces protruding into moldic cavities and the anhedral - subhedral crystals of the groundmass. In addition, there is no difference in the luminescent properties of these crystals. This suggests that the crystals protruding into the cavities and the crystals of the groundmass have the same origin. The dolomite which replaced the $CaCO_3$ precursor formed a xenotopic to hypidiotopic texture, with the crystals being riddled with inclusions that were presumably derived from the lime mud. However, when the dolomite crystals that replaced the original carbonate reached a cavity, they were able to grow more freely and develop euhedral faces into the void. Aragonite dissolution therefore, must have occurred prior to or during replacive dolomitization.

Aragonite dissolution can occur in any region where solutions are undersaturated with respect to aragonite. It had commonly been thought that HMC, being more soluble, should preferentially dissolve before aragonite. However, Walter (1983) discovered that HMC with 12 mole % Mg was equal in solubility to aragonite, a much higher mole % Mg than previously thought. If a solution is undersaturated with respect to both calcite and aragonite, then aragonite may preferentially dissolve (Walter, 1985). The main controls on dissolution in this undersaturated solution are microstructural, in that aragonite grains with the greater surface area per unit mass, may dissolve more readily than even the thermodynamically less stable HMC (Walter and Morse, 1984).

Meteoric diagenetic systems may have the capacity to dissolve aragonite preferentially, but only periodically. Walter (1985) emphasized the dynamic nature of such systems, in which the dissolution of one mineral phase may cause the porewaters to evolve in composition towards a different saturation state. In the Yucatan Peninsula, aragonite is preferentially dissolving where Pleistocene limestone crops out in the mixing zone of fresh and marine phreatic waters (Hanshaw and Back, 1980; Ward and Halley, 1985; Back *et al.*, 1986). Porewaters from marine environments may also be periodically undersaturated with respect to both aragonite and calcite; thus, the potential for preferential aragonite dissolution may arise e.g. in buried shallow marine sediments, aerobic respiration increases pCO_2 values enough to force the porewaters below saturation (Walter, 1985).

None of the fossil molds are filled by internal marine sediment. This lack of sediment within molds implies that aragonite dissolution occurred away from the rock/sediment - water interface (Saller, 1986). Aragonite dissolution must have occurred after at least some lithification of the substrate (to preserve the shapes of the molds) and after the rock had been isolated from the direct influence of the rock/sediment - water interface. Since aragonite dissolution occurred prior to or during dolomitization, dolomitization must also have occurred away from this interface. Dolomitization was therefore not contemporaneous with deposition.

B. PERVASIVE DOLOMITIZATION

The following points are important to any discussion concerning the origin of the pervasive dolomitization that affected the Bluff Formation:-

(1) The dolomite originated through the replacement of a presumed calcium carbonate precursor. This is amply demonstrated by the fact that many skeletal fragments have been replaced by dolomite.

(2) Many of the skeletal components were metastable during dolomitization. This is confirmed by the supposition that formerly aragonitic skeletal components preferentially dissolved prior to and/or penecontemporaneous with dolomitization, while components originally of HMC are preferentially dolomitized. Such differential preservation indicates that the limestones were not stabilized to LMC prior to dolomitization.

(3) The microtextures of the HMC skeletal components were retained despite dolomitization (e.g. Plate 9A, F; 10A). The dolomitization of HMC results in the retention of the grain's fabric and crystallographic orientation,,whereas the dolomitization of LMC and aragonite usually results in grain destruction (Land and Epstein, 1970; Sibley, 1980). Had all the allochems been stabilized to LMC, then such differential preservation would not be observed.

(4) Sibley (1980, 1982) argued that only metastable carbonates are likely to be pervasively dolomitized, as LMC can essentially resist such pervasive dolomitization.

(5) Sibley (1982) suggested that inclusion free dolomite crystals result from replacement of aragonite, whereas cloudy, inclusion riddled dolomite crystals result from

calcite replacement. Some of the corals which were originally aragonite have been replaced by dolomite that is clearer and coarser than that in the groundmass. The inclusion riddled finer dolomite may be due to a larger proportion of calcite in the original carbonate mud, or due to the fine grained nature of the original lime mud. This differential style of dolomitization therefore also indicates that metastable carbonates were replaced.

(6) If, as has been suggested, the ^{13}C values of the dolomites retain the carbon isotopic signature of the precursor carbonate (Dunham and Olson, 1980; Land, 1985), rather than establishing a new one from the dolomitizing solutions, then the typically marine values observed in the Bluff Formation dolomites reflect a precursor marine carbonate phase. Most stabilization of limestones to LMC occurs in the meteoric environment, where carbon values are readily exchanged to equilibrate with the diagenetic fluids. Had the precursor carbonate been LMC, then isotopically light carbon values would be expected in response to the freshwaters that stabilized the limestone. This point however must be considered equivocal due to the controversy surrounding dolomite isotope interpretations.

The above considerations strongly suggest that dolomitization replaced a metastable precursor carbonate. This suggests that dolomitization must have been early in the diagenetic history of the rocks before the carbonates had a chance to stabilize to LMC. Indeed, it has been postulated that finely crystalline dolostone (less than 10 to 20 μm) may be indicative of early dolomitization that has replaced a metastable precursor (Morrow, 1982b). Nichols and Silberling (1980) also argued that fine grained dolomite with good texture preservation was indicative of early dolomitization (in that dolomitization occurred within the influence of the depositional environment). It is apparent however, that dolomitization in the Bluff Formation was not contemporaneous with deposition, as the fossil molds formed prior to and/or during dolomitization contain no marine sediment.

Although probably early, dolomitization occurred away from the direct influence of the depositional environment, and away from the sediment - water interface.

The types of preservation found in these dolomites provide a clue to the nature of this dolomitization process. The type of replacement is comparable to that which occurred in the dolostones of the Bahamas described by Supko (1977) and Kaldi and Gidman (1982), and Bonaire described by Sibley (1980; 1982). The retention of microtextures (e.g. in the red algae), the preservation of crystallographic orientations (e.g. in the foraminifera), and the replacement of echinoid spines and plates, each by a single dolomite crystal, all indicate that the replacement process was on a volume-for-volume, rather than a mole-for-mole basis.

C. NUMBER OF DOLOMITIZATION EVENTS

The number of dolomitization events in the Bluff Formation is important in establishing the timing of dolomitization.

Isotopically, the dolostones fall into one region (Fig. 10), especially with a fairly narrow range of oxygen values. This implies that dolomitization was either (1) a single, although not instantaneous, pervasive event, or (2) a number of repeated events produced by similar processes. The general covariant trend of the dolomite isotope values (Fig. 10) indicates a progressive change in the groundwaters during dolomitization, implying dolomitization may have been a slow process. Machel and Mountjoy (1986, p. 203) suggested that "If there is good fabric preservation and/or little or no intercrystalline porosity, dolomitization must have been a piecemeal dissolution - reprecipitation process. This indicates relatively *slow* replacement, which can be due to any combination of low flow rates and low degrees of supersaturation." Both good fabric preservation and little intercrystalline porosity are observed in the dolostones of the Bluff Formation.

The dolomite cements were obviously precipitated after pervasive dolomitization as these cements are rooted on a replacive dolomite substrate. Whether these cements are a later phase diagenetic event or part of the progression of increasingly ^{16}O and ^{12}C enriched groundwaters is unknown as they could not be physically separated for isotopic analysis. If these cements represent the final stages of a single dolomitization event, then the process must have been a long one involving the evolution of dolomitizing porewaters.

Cathodoluminescence shows that the dolomite of the groundmass is dully orange luminescent, while the dolomite cements are non-luminescent, except where bright orange narrow luminescent bands are visible towards the edges of the crystals (Plate 18F, G). This consistency of the luminescent properties of the dolomitic groundmass supports the contention that dolomitization was achieved by a single process. The consistency of the zoning in the dolomite cements suggests that the cements were precipitated from similar solutions at about the same time.

Due to the narrow range of dolomite values which encompass both the replacive dolomite and the dolomite cements, it is most probable that these were either (1) part of the same dolomitization process, albeit an evolving one, or (2) that the cements precipitated in a later event, from similar solutions to those precipitating the replacive dolomites. The latter possibility means that dolomite cement was not precipitated in the first dolomitization event, even though solutions were conducive to their precipitation. The former possibility seems the most likely. In view of the assumptions made in the previous chapter i.e. that the dolomite cements precipitated soon after the initial emergence of the island, it is most probable that the cements formed shortly after or immediately following pervasive dolomitization. Thus, dolomitization was probably a single pervasive event, culminating in the precipitation of dolomite cements in porewaters less saline than the first replacive dolomite crystals.

D. DOLOMITE NONSTOICHIOMETRY

Dolomite nonstoichiometry (expressed as mole % $CaCO_3$ in the dolomite crystal lattice) may be indicative of the type of dolomitization process that was operative (Lumsden and Chimahusky, 1980). Calcium is incorporated into the growing dolomite lattice with relative ease, and thus the proportions of anions in the dolomitizing solution may affect the type of dolomite produced.

The method of determining dolomite nonstoichiometry (or variations from ideal composition) of bulk dolomite samples (Goldsmith and Graf, 1958) using the $d_{(104)}$ values of 2.886 Å for 50 % $CaCO_3$ and 2.901Å for 55 % $CaCO_3$ is consistent with the values used by Füchtbauer and Goldschmidt (1965), Richter (1974) and Lumsden and Chimahusky (1980). Land (1985) believed that this method of determining dolomite nonstoichiometry is only "grossly useful", as it is very difficult to distinguish the effects of compositional changes from those of ordering in the crystal lattice. With this caution in mind, the Mg/Ca data of the dolomite crystals can be presented (Table 10).

The Bluff Formation dolomites range in composition from stoichiometric dolomite, with 50 mole % $CaCO_3$, to more calcian dolomite, with 55 mole % $CaCO_3$ (Table 10). It has been suggested that the degree of dolomite stoichiometry is controlled at least in part by the chemistry of the dolomitizing fluid (Katz, 1971; Lumsden and Chimahusky, 1980).

Folk and Land (1975) speculated that the calcium excess in dolomite was controlled by salinity. A decrease in salinity accompanied by a decrease in the Mg/Ca ratio of the solution should result in a more stoichiometric dolomite than dolomite produced in higher salinities with higher Mg/Ca ratios. This may be due to the slower precipitation in dilute solutions, which would overshadow the influence of the solution Mg/Ca ratio. It is more popularly believed, however, that there is, conversely, a positive dependence of the Mg/Ca molar ratio in the dolomite crystal upon that of the solution (Füchtbauer and Goldschmidt, 1965; Katz, 1971; Richter, 1974; Sass and Katz, 1982; Sperber *et al.*, 1984). At higher salinities and higher Mg/Ca ratios, the resulting dolomite is more stoichiometric than dolomite formed at lower Mg/Ca ratios, as the higher abundance of mg ions compensates for the relative difficulty that dolomite has to precipitate from solution (Morrow, 1982b).

In sabkha environments of the Persian Gulf, Patterson and Kinsman (1974) observed a decrease in the Mg/Ca ratio of the porewater with increasing depth in the sediment. This was accompanied by a decrease in the magnesium content and an increase in the calcium content of the dolomite crystals. This shows the positive dependence of dolomite composition on that of the solution as a result of the increasing degree of isolation of the system. Sass and Katz (1982) argued that an increase in the isolation of the system during dolomitization could produce a decrease in the Mg/Ca ratio of the porewaters. This results from the removal of magnesium from these waters, which then cannot easily be replenished owing to the restrictiveness of the system. This degree of isolation is reflected in the Mg/Ca molar ratio of the dolomite: the more isolated the system, the more Ca-rich the dolomites will be.

The stoichiometric to slightly Ca-rich nature of the Bluff Formation dolomites indicates that dolomitization occurred in an open to slightly isolated system. Dolomitization probably occurred away from the sediment-water interface, and thus the rocks were highly permeable and/or close to this interface during dolomitization. Finely crystalline modern and ancient dolomite not associated with evaporites range from about 51 to 62% $CaCO_3$, while those associated with evaporites range from about 50-51 to 56-57% $CaCO_3$ (Morrow, 1982b, Fig. 8, p. 102). The range of values in the Caymanian dolomites falls most closely into the latter type.

E. TIMING OF DOLOMITIZATION

Several lines of evidence serve to constrain the timing of dolomitization, the most obvious of which is the difference in composition between the Ironshore and Bluff formations. A distinct contact exists at the unconformity between these two formations; the former being almost 100% dolomitized whereas the latter contains no dolomite (Jones *et al.*, 1984). This indicates that the Bluff Formation was dolomitized prior to the deposition of the late Pleistocene Ironshore Formation.

Many breccias occur in the Bluff Formation, commonly with dolostone clasts as the main component, which also help to constrain the timing of dolomitization. The most common types are caymanite and terra rossa breccias, but flowstone, oolitic, and skeletal limestone breccias are also present. The latter type have been described by Jones and Kahle (1985), who refer to them as karst breccias. These occur on the east end of the island, EEQ and BR, where they fill solution pipes. The caymanite and terra rossa breccias occur in all localities; the flowstone breccia has been found in EEQ and BH and the oolitic breccia is present in EEQ and BR.

The terra rossa, flowstone, oolitic and skeletal limestone breccias all have a $CaCO_3$ matrix. The presence of angular clasts of dolostone in these matrices indicates that dolomitization occurred prior to the breccia formation. Angular clasts of dolostone also occur in a fine caymanite matrix, commonly at the base of any one sequence. Even though both clasts and matrix are composed of dolomite, the boundaries between these are distinct. This suggests that recrystallization has not occurred since brecciation, and that the clasts of host Bluff Formation rock were dolomitized prior to incorporation into the caymanite breccia. This is further substantiated by the fact that caymanite commonly fills cavities in the Bluff Formation that had already been lined with dolomite cement (Lockhart, 1986). Cavities formed within the caymanite may be filled with a sequence of cements inferred to be of early diagenetic origin (Plate 13C). Hence, at least some caymanite was deposited early in the diagenetic history of the rocks, and dolomitization occurred before at least some caymanite was deposited.

The hypothesis that dolomitization occurred before the precursor carbonates were fully stabilized also indicates that dolomitization may have been early in the diagenetic history of the rocks. Shortly after deposition of the Bluff Formation, Grand Cayman became emergent. This would have taken the rocks relatively through the mixed water phreatic, the freshwater phreatic and the vadose diagenetic zones. By the end of this regression, and probably sometime before, the rocks would probably have been stabilized to LMC. This therefore constrains the timing of the dolomitization event, which would have had to have occurred prior to the end of this regression. Sea level reached a minimum in the mid-late Miocene (Hallam, 1984). In addition, isotopic evidence also indicates that dolomitization occurred under the influence of marine waters. Thus, dolomitization must have occurred prior to, or early in this regression.

F. SYNOPSIS

Any model of dolomitization for the Bluff Formation must take the following points into account:

1. Aragonite dissolution occurred prior to and/or penecontemporaneous with dolomitization,
2. Replacive dolomitization was pervasive, whereas dolomite cementation was localised,
3. The dolomite replaced a largely metastable lime mud,
4. Dolomitization occurred early in the diagenetic history of the rocks,
5. Dolomitization resulted in a cloudy dolomitic groundmass, with some of the euhedral rhombs having a rim of clear dolomite and core of cloudy dolomite, and some being more complexly zoned,
6. The dolomite is near-stoichiometric,
7. Replacement was on a volume-for-volume basis,

8. Dolomitization was probably a single process, which proceeded through evolving groundwaters,
9. The depth of dolomitization is unknown,
10. Dolostones of the Bluff Formation also occur on both Cayman Brac and Little Cayman,
11. There is no evidence of evaporite minerals or textures,
12. The cement stratigraphy in the pores indicates progressively freshening porewaters, probably associated with the initial emergence of the island,
13. The stable isotope data indicates precipitation in normal seawater.

G. MODELS OF DOLOMITIZATION

"Dolomite is thermodynamically stable in most natural solutions at earth surface conditions and a thermodynamic drive exists for the conversion of calcite and aragonite to dolomite" (Morrow, 1982a, p. 11). The process is hindered, however, by numerous kinetic obstacles at low temperatures. The inability to precipitate dolomite from solutions less than 100°C has inhibited direct interpretation of the physical and chemical conditions operative during precipitation of most dolomites. Dolomitization is thermodynamically and kinetically favoured at lower Ca/Mg ratios, lower Ca/CO_3 ratios and at higher temperatures (Morrow, 1982a; Machel and Mountjoy, 1986). Massive dolomitization requires a large source of magnesium and a pump mechanism to circulate the ions through the rock being dolomitized (Land, 1985). The dolostones of the Bluff Formation have not undergone any burial, and therefore, the only source of such a vast quantity of magnesium is seawater (whether modified or unmodified). Machel and Mountjoy (1986) pointed out that dolomitization may be kinetically favoured in any salinity. Hence, dolomitization of the Bluff Formation may have occurred in three types of seawater: hypersaline, normal marine, or diluted waters.

A Hypersaline origin?

Environments of increased salinity thermodynamically favour dolomitization due to an increase in supersaturation (Machel and Mountjoy, 1986). The sabkha model, based on the modern analogue in the Persian Gulf, has been one of the most popular models for large-scale dolomitization. It is based on the idea that as seawater increases its concentration by intense evaporation, aragonite and gypsum will precipitate, thereby causing an increase in the Mg/Ca ratio of the remaining brine (Kinsman, 1966). The precipitation of gypsum in such an environment also removes sulphate, which is thought to be an inhibitor of dolomitization (Baker and Kastner, 1981). Other kinetic obstacles are also removed, since the Mg ion is less hydrated at higher salinities. The resulting Mg rich brine may then have the potential to dolomitize either by reflux (Adams and Rhodes, 1960) or by evaporative pumping (Hsü and Siegenthaler, 1969).

Dolomitization in sabkha environments results only in small quantities of dolomite, the process being small in scale and penetrating only to shallow depths. Due to the unstable nature of sabkhas, dolomite may not form at all (Machel and Mountjoy, 1986).

Dolomites precipitated from hypersaline waters are generally near stoichiometric (about 50 to 56 mole % $CaCO_3$), fine grained and associated with evaporites (Morrow, 1982b). Although the first two criteria are satisfied in the dolomites of the Bluff Formation, the latter is absent. Despite extensive study, no evidence has been found to suggest evaporites were ever present in the Bluff Formation. In addition, no evaporites or evaporitic textures have been found in the Ironshore Formation. In the sea spray zones around the coast of Grand Cayman Island small insignificant pockets of salt have been found in notches in the rocks. Hence, sabkha dolomitization cannot be invoked to be the cause of dolomitization of the Bluff Formation. However, the finely crystalline, cloudy nature of the dolomite, and its stoichiometric to calcium rich (50 to 55 mole % $CaCO_3$) composition, is consistent with the dolomite's formation from slightly hypersaline waters.

Normal seawater ?

It is widely recognised that dolomite is the thermodynamically stable carbonate phase in seawater (e.g. Bathurst, 1975; Baker and Kastner, 1981). Seawater is also an infinite reservoir for the abundant ions required for dolomitization. In spite of the fact that seawater is about 20 times supersaturated with respect to dolomite (Bathurst, 1975), very little dolomite has been found on the seafloor. Kinetic factors such as the low activity of the carbonate ion compared to that of the Ca and Mg ion (Morrow, 1982a), or the inhibiting effect of the sulphate ion in seawater (Baker and Kastner, 1981) may be responsible. Additionally, there is the idea that simple immersion in seawater for several millions of years is insufficient to cause massive dolomitization (Folk and Land, 1975). This may be due to the inefficiency of static fluids to dolomitize (Land, 1985). Hardie (1987, p. 178) suggested that given enough time "...normal seawater at normal oceanic temperatures might accomplish the kind of massive, near-surface replacement dolomitization seen in the Pleistocene-Upper Tertiary carbonates of the Bahamian-Caribbean buildups...". He also emphasized the importance of mass transfer processes required for dolomitization, in particular the need for a stable hydrologic system. Therefore, a hydrological system that had the capacity to pump seawater through the rocks and sustain this circulation for thousands of years must be found. Obviously, such properties as the permeability of the carbonates will determine how effective the pump system is, and this may explain the lack of massive dolomites in deep sea sediments that accumulated slowly in seawater.

Sass and Katz (1982) reported an example of large scale penecontemporaneous dolomitization by normal saline to hypersaline waters, which affected the Cretaceous Soreq Formation of Israel. This has been termed the 'diagenetic model' or the 'shallow-subtidal' model by Machel and Mountjoy (1986) who pointed out that all models of dolomitization are diagenetic. These Israelian dolostones consist of (1) laminar dolomites devoid of skeletal grains, (2) heterogeneous dolomites with moldic porosity, and (3) homogeneous, sterile dolomites. Crystal size ranges from 1 to 100 µm, with the coarsest crystals associated with the heterogeneous dolomites. The differing depositional textures of the Soreq and Bluff formation dolostones make it difficult to compare the resulting diagenetic textures. Both dolostones are commonly over 95% dolomitized.

Two driving mechanisms have been invoked to account for dolomitization by essentially unmodified seawater: (1) downward or plume reflux, and (2) upward (or thermal or Kohout) convection.

Downward plunging plumes of slightly hypersaline (dense) bank/lagoonal water overlying lighter porewaters can dolomitize to depths of hundreds of metres provided there are no obstructive aquicludes (Simms, 1984). Evaporites need not be formed by this process (Simms, 1984).

The 'thermal convection' (Machel and Mountjoy, 1986), 'upward convection' (Aharon *et al.*, 1987) or 'Kohout convection' (Simms, 1984) model of dolomitization operates in the reverse direction to the reflux model. Simms (1984) expanded the work of Kohout (1965) and Kohout *et al.* (1977) based on convection patterns in the Floridan Aquifer. A density gradient, developed between warm platform interior waters and cold surrounding seawater, causes seawater to enter the platform at depth. This flows laterally towards the interior, displacing the warm waters upwards. These warm waters may dolomitize and then exit to the sea via geothermal springs (Fanning *et al.*, 1981). Saller (1984) proposed a similar small scale mechanism, whereby cold seawater enters an atoll at depth, is warmed within the atoll and rises, dolomitizing the limestone *en route*. Aharon *et al.* (1987) proposed the same mechanism for the dolomitization of the atoll at Niue in the south Pacific. The geochemical evidence indicated dolomitization by normal seawater, and the upward convection model was chosen rather than the reflux model owing to the distribution of non-alkaline elements in the dolomites. Fe, Cu, Mn and Zn distributions derived from the volcanic basement, showed that seawater must have been moving upwards. Aharon *et al.* (1987) suggested that many of the Pacific atolls were probably dolomitized in the same way.

A more extreme view involves hydrothermal circulation near Mid Oceanic Ridges. This can result in mass transfer of ions (Mottl and Holland, 1975) as well as a release of warm seawater. What influence such hydrothermal circulation, associated with the oceanic ridge just south of Grand Cayman, has had on the Bluff Formation cannot be determined, but at best it may have reduced some of the kinetic obstacles by heating the water.

The stoichiometric to Ca rich nature of the dolomites in the Bluff Formation can be explained by a partially closed system, with solutions of normal marine to hypersaline waters. The stable isotope data indicates that dolomitization occurred in waters with strong marine affinities. The normal marine depositional components and the lack of evaporites preclude strongly hypersaline waters, but do not preclude early dolomitization in slightly hypersaline waters. Thus, dolomitization from essentially normal marine waters, away from the sediment-water interface in a partially closed system, seems plausible. The dolomites on Grand Cayman reach a depth of at least 20 m. The Bluff Formation extends to a depth of about 400 m, but it is not known how much of this is dolomite. Until the depth relations are studied, it is indeterminable which process (upward convection or plume reflux), if either of these, may have been operative. Temperature predictions from stable isotope data (Fig. 12) indicate that the oxygen values could be a result of precipitation from normal marine to slightly hypersaline waters (assuming SMOW H_2O 0 to +2.5‰) at temperatures ranging from 25° to 36°C (for the average dolomite value) or from 21° to 41°C (for the minimum and maximum dolomite values).

A mixed water origin ?

Hanshaw *et al.* (1971) launched the mixing zone dolomitization idea by documenting small dolomite crystals on the walls of caves in the "boulder zone" of the Floridan Aquifer. Badiozamani (1973) formulated the Dorag mixing model based on the assumption that dolomitization of calcite will occur in waters undersaturated with respect to calcite, but supersaturated with respect to dolomite. This is based on the idea that the mixing of two solutions saturated or supersaturated with respect to calcite but at differing CO_2 pressures may produce a solution that is undersaturated with respect to calcite but supersaturated with respect to dolomite (Bögli, 1964; Howard, 1966; Thrailkill, 1968; Badiozamani, 1973; Plummer, 1975; Wigley and Plummer, 1976). Equally, if two solutions undersaturated with respect to calcite are mixed, the resulting solution may be supersaturated with respect to calcite. This is due to the non-linearity of the solubility curve (Runnells, 1969). Matthews (1971) noted that the mixing of seawater with freshwater could result in a brackish solution undersaturated with respect to calcite.

Most mixing zones, however, are not supersaturated with respect to dolomite and/or undersaturated with respect to calcite (Machel and Mountjoy, 1986). This is a result of the dependence of the mineral saturation levels on temperature, pCO_2, pH and meteoric water composition, which will differ from one mixing zone to another (Plummer, 1975; Plummer *et al.*, 1976). In addition, the thermodynamics of the Dorag model were calculated for ordered dolomite rather than for Ca-rich, less ordered dolomite that is typically formed at surface temperatures (Hardie, 1987). This further constrains the application of the model.

The schizohaline mixing model described by Folk and Land (1975) is based on the idea that the mixture of seawater and its evaporated brines with freshwater, may result in the retention of a high Mg/Ca ratio and a decrease in salinity. This was based on the unsubstantiated assumption that a low salinity favoured dolomitization, even with Mg/Ca ratios as low as 1:1. Machel and Mountjoy (1986) however, argued that dolomitization is kinetically favoured at low as well as high salinities, and at higher Mg/Ca ratios.

The major stumbling block for the mixing models is the lack of a modern analogue. Most of the studied active coastal mixing zones in carbonate rocks exhibit no dolomitization, and where dolomite is found, these are small numerically insignificant cement crystals (Hanshaw *et al.*, 1971; Steinen and Matthews, 1973; Plummer *et al.*, 1976; Steinen *et al.*, 1978; Gebelein *et al*., 1980; Hanshaw and Back, 1980; Budd, 1984).

Waters of the mixing zone create porosity and permeability needed to provide conduits for the dolomitizing solutions, but "It would be a unique chemical environment in which the rate of dolomite growth equalled the rate of limestone dissolution." (Plummer, 1975, p. 234). Carpenter (1976, p. 254) further argued that "...the mixing of seawater with groundwater will generally act to inhibit dolomitization and that any dolomite which may form in the zone of mixing is produced despite the mixing rather than because of it." Thus, in no way can the minor dolomitization found in a few of the modern mixing zones be compared with the pervasive dolomitization present in the Bluff Formation.

Mixing zone dolomites in the Pleistocene of the Yucatan Peninsula have been inferred to have been precipitated from almost 100% seawater (Ward and Halley, 1985). The rare dolomite, which is of mainly primary origin, is localised to permeability conduits. This constitutes the most plausible example of ancient mixing zone dolomitization, and occurs in much higher salinity waters than those postulated by Folk and Land (1975).

Certain features of the Caymanian dolostones are thus inconsistent with the mixed meteoric-marine dolomitization model:

1. The Bluff Formation is almost 100% dolomitized. This is inconsistent with the volume of dolomite observed in present day mixing zones.
2. The isotopic compositions indicate predominantly seawater affinities (normal to slightly hypersaline), rather than meteoric.
3. The contention that dolomitization was early in the diagenetic history of the rocks, prior to the stabilization of the precursor carbonate, suggests that dolomitization occurred prior to extensive meteoric influence.

4. The dolomite nonstoichiometry (50 to 55 mole % $CaCO_3$) also precludes dilute solutions with low Mg/Ca ratios, although the dependence of the dolomite composition on that of the solution is somewhat equivocal.

In contrast, the later diagenetic cement stratigraphy can be accounted for by mixed waters. The limpid dolomite cements in cavities in the Bluff Formation may be of a mixing zone origin. They only occur locally and are volumetrically insignificant. Folk and Land (1975) argued that limpid dolomite was a result of precipitation in dilute solutions. However, it has also been shown that this type of dolomite may form ~~from~~ quite saline waters in the mixing zone (Ward and Halley, 1985). Aulstead and Spencer (1985) even demonstrated that vein-lining, clear dolomite precipitated from concentrated brines. Thus, the type 1 and 2 dolomite cements of the Bluff Formation probably precipitated from the most saline waters of the mixing zone after the pervasive dolomitization event in more saline waters. The type 3 dolomite-calcite alternating cements probably precipitated in slightly more dilute waters of the mixing zone in chemically fluctuating conditions. A schizohaline environment could give rise to such fluctuations and result in the successive precipitation of the two minerals. As one mineral precipitates, the porewaters consequently evolve resulting in the precipitation of the other. The calcite cements were precipitated in mixed to dilute, freshwaters. This sequence of cements which records progressively freshening groundwaters is comparable to the sequence of cements described by Ward and Halley (1985) and interpreted to represent the passage of the rocks through the mixing zone.

The lack of any dolomite in the Pleistocene Ironshore Formation is problematic as these have passed through at least one mixing zone in their history. Possibly, environmental conditions have changed resulting in a mixing zone not conducive to any dolomite precipitation, or the passage through the mixing zone was too rapid and conditions too unstable at this time. The absence of dolomite supports the idea that mixing zones are not always conducive to the formation of dolomite.

II. A DOLOMITIZATION MODEL FOR THE BLUFF FORMATION ?

To make a conclusion concerning the origin of the Bluff Formation dolostones 'by default' is unavoidable. It is apparent that burial, sabkha hypersaline and mixing zone dolomitization can be ruled out for the pervasive dolomitization. The latter however, could account for the cement sequences found in the Bluff Formation which succeed replacive dolomitization. Thus, essentially unmodified seawater must have caused the pervasive dolomitization. This idea is supported by the following observations:

1. The fine grained (5-10 μm) and cloudy nature of the dolomite,
2. The enriched carbon and oxygen isotopic compositions,
3. The stoichiometric to Ca-rich nature of the dolomite,
4. The lack of evaporites,
5. The normal marine precursor carbonates,
6. The early timing of the dolomitization process, before the precursor carbonates were stabilized.

Plume reflux and thermal convection have been postulated to account for dolomitization by normal marine waters, most notably in Pacific atolls. Further work, when core is available, may elucidate which of these mechanisms is most plausible, or if another mechanism may be invoked. Any gradational changes in the petrography or the geochemistry of the dolomites may elucidate in which direction the dolomitizing solutions were moving. It may be pertinent to mention the similarities in the physiography between atolls, small carbonate islands perched on volcanic ridges and pinnacle reefs in the rock record. Fluid movements within these features may have been similar and on a similar scale.

Other factors such as plate tectonic influence (Mottl and Holland, 1975), changing seawater compositions (Mackenzie and Piggott, 1984), changing seawater temperatures or changing climate may all have played a role in dolomitization. It is possible that there is not one identifiable model to account for the massive dolomitization. This replacement process may have come about as a result of a number of factors. It is apparent that in terms of dolomitization, 'the present is not the key to the past'.

I. COMPARISON WITH DOLOSTONES OF OTHER CARIBBEAN AREAS

Sibley (1980) outlined the similarities of the dolostones of the Pliocene Seroe Domi Formation of Bonaire with those of the Pleistocene Hope Gate Formation of Jamaica, and the Pleistocene Bahamian dolostones of San Salvador. Jones *et al.* (1984) noted the same similarities, and included the Plio-Pleistocene dolostones of Great Abaco Island (based on the description of Kaldi and Gidman, 1982) and the Bluff Formation of Grand Cayman. Sibley (1980) suggested that the similarities indicated dolomitization events common to the entire Caribbean area, related to sea level and/or climatic controls. Despite the similarities in the diagenetic features of the dolomitized limestones of the Yucatan Peninsula with the other dolostones, Ward and Halley (1985) refuted the idea of a Caribbean-wide regional dolomitization event.

In many respects, the dolostones of the Bluff Formation resemble those of the other Caribbean areas documented, especially in superficial terms. Jones *et al.* (1984) drew special attention to the 'striking' resemblance between the Bluff Formation and the Hope Gate Formation described by Land (1973a). However, the Hope Gate Formation is not pervasively dolomitized (only 30 - 40%), whereas the Bluff Formation is:- a major flaw in any comparison of dolomitization events! Likewise, the Yucatan dolostones are at a maximum only 50% dolomitized, while bulk analysis indicates less than 20% dolomite (Ward and Halley, 1985). Most of this is represented by dolomite cements. No figures are given for either the Great Abaco dolostones (Kaldi and Gidman, 1982) or the Seroe Domi Formation dolostones (Sibley, 1980). The latter are composed of 80 µm dolomite crystals, much larger than the typical Bluff Formation dolomites. The dolostones of Great Abaco Island bear a remarkable resemblance to those of the Bluff Formation, in which preferential dissolution has removed the aragonitic corals to leave molds containing natural casts of *Entobia*. Similarly, the crystals have an average size of 30 µm and are very inclusion rich. The dolostones of San Salvador, Bahamas (Supko, 1977) are on the other hand, 100% dolomitized, and have similar petrographic features to the Bluff Formation dolomites. Those Caribbean dolostones analysed isotopically, all exhibit similar composition ranges, and this has been emphasized in the comparisons of the dolostones. However, these isotopic similarities should be used with caution, in view of their highly equivocal nature.

Thus, in terms of pervasive dolomitization, only those dolostones of San Salvador and possibly Great Abaco Island can be confidently compared to those of the Bluff Formation, and probably formed from similar processes. However, even these three dolomitization events cannot be synchronous. The Bluff Formation dolomitization is interpreted to be very early in the diagenetic history of the rocks, before the first relative passage of the rocks through the mixing, freshwater and vadose zones. Grand Cayman Island emerged shortly after the middle Miocene (Perfit and Heezen, 1978), corresponding to a major sea level fall at this time (Vail and Hardonbol, 1979; Hallam, 1984). The other two dolostone formations are Pliocene to Pleistocene in age and therefore had not even been deposited at this time.

More convincing similarities can be drawn between the ensuing cement sequences of many of these formations. Certainly the regressive sequences documented by Land (1973a) for the Hope Gate Formation, by Kaldi and Gidman (1982) for the Great Abaco dolostones, and by Ward and Halley (1985) for the Yucatan dolostones are directly comparable to that of the Bluff Formation. These are interpreted to reflect the passage of the rocks from the mixing zone, through the freshwater phreatic zones. Again, these sequences are not synchronous but are a result of similar diagenetic processes due to a relative fall in sea level. The cement sequences of San Salvador dolostones and the Seroe Domi Formation are more complex, but this is probably due to variations in hydrological and climatic settings, original carbonate properties etc. Even these latter two are interpreted to represent mostly mixing zone diagenesis.

Dolomitization events were therefore not synchronous in the Caribbean region, and some may not have even formed from the same processes. The cement sequences record progressively freshening porewaters as a result of the passage of the rocks relatively through the mixing zone and freshwater phreatic zone. Most sequences also end up in the freshwater vadose zone. These cement sequences, although having formed from similar processes, are not correlative.

VII. CONCLUSION

1. The sediments of the Bluff Formation of Grand Cayman Island were deposited in normal marine waters in a backreef to lagoonal setting, during Oligocene to middle Miocene times.
2. Casts of borings made by endolithic organisms of the original environment occur in the skeletal molds of the Bluff Formation, and are responsible for an average of 38% of the coral molds.
3. Entobians dominate the boring assemblage, and constitute about 75% of the borings. *Trypanites*, *Gastrochaenolites*, *Maeandropolydora*, *Talpina*, *Caulostrepsis*, new ichnogenera A and B, and microborings are also present.
4. Bioerosion by middle Tertiary times was comparable in terms of the organisms involved and the scale of damage to bioerosion in modern reefs.
5. Grand Cayman Island became emergent following the middle Miocene, resulting in the passage of the rocks from the marine through the mixing and freshwater phreatic to the vadose zones (Fig. 14). This is reflected in the diagenetic textures present in the dolostones.
6. Dolomitization occurred away from the sediment-water interface, but under the influence of normal to slightly hypersaline waters, before the island became emergent, and before the carbonates were stabilized to LMC. The dolomites have an average $\delta^{18}O$ value of +2.4‰ and $\delta^{13}C$ value of +2.0‰, relative to PDB, indicating that the dolomites are in equilibrium with carbonates formed in normal seawater.
7. Aragonite dissolution occurred prior to or penecontemporaneous with dolomitization.
8. With the influx of meteoric waters as the island emerged, a sequence of zoned dolomite and calcite cements were precipitated, as a response to increasingly fresh groundwaters. The precipitation of zoned dolomite with cloudy cores and clear rims in the mixing zone was followed by preferential dissolution of the cloudy cores to produce hollow

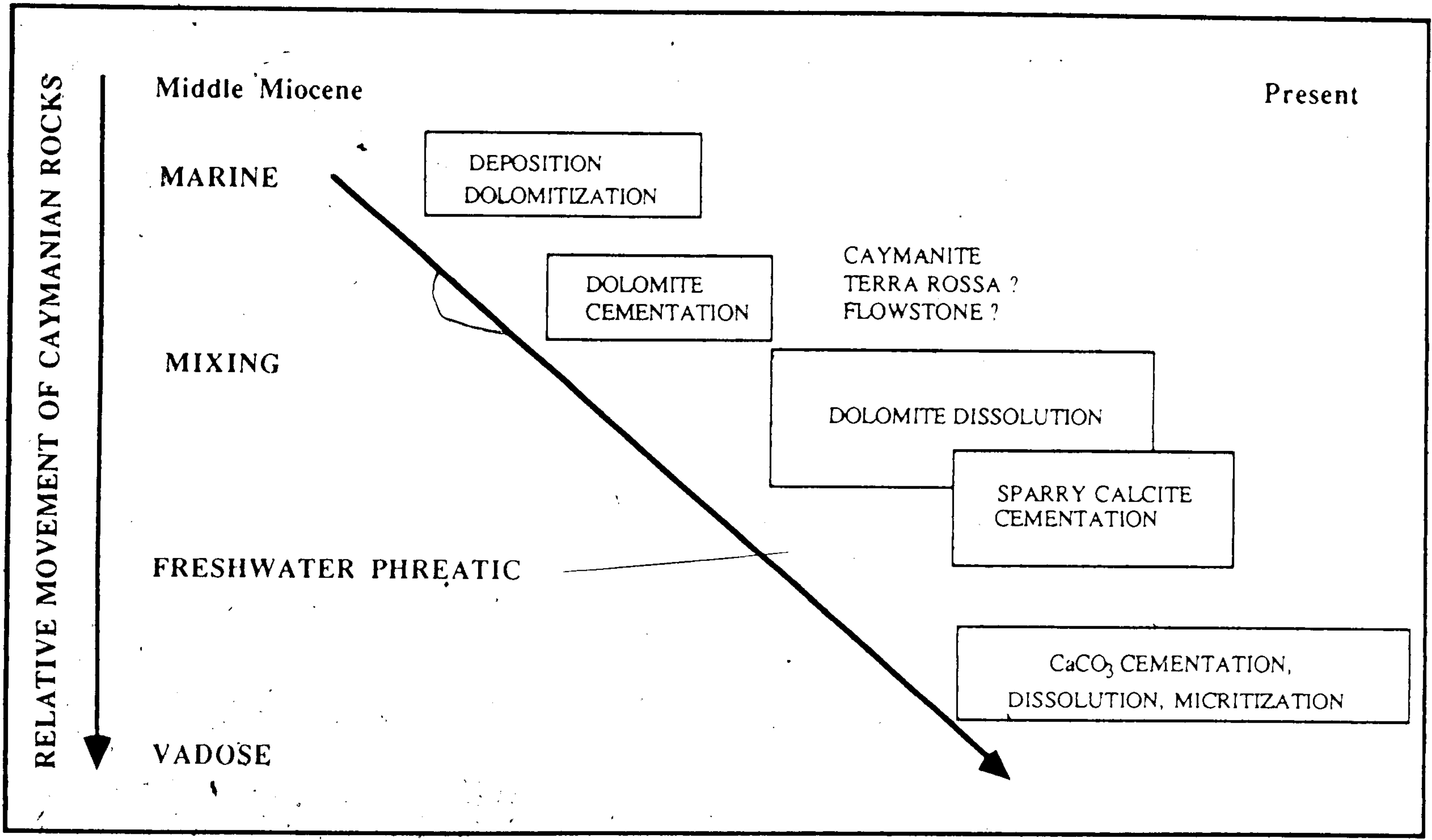


Figure 14. Schematic diagram to show the paragenetic sequence of the Bluff Formation.

rhombs. Sparry calcite cementation occurred in fresher waters of the mixing zone and freshwater phreatic zone. A covariant trend towards increasingly depleted $\delta^{18}O$ and $\delta^{13}C$ values of the calcites indicates progressively fresh porewaters as they precipitated. This calcite commonly filled and encased the hollow dolomite rhombs, producing a poikilotopic texture. This can be considered a process of dedolomitization by distinct time separated stages of (1) dolomite precipitation, (2) dolomite dissolution and (3) calcite cementation.

9. Vadose diagenesis, which followed dolomite precipitation and some sparry calcite cementation, is represented by (1) dissolution features: spiky calcite, knobbly textures, (2) $CaCO_3$ cements: columnar crystals, calcified filaments, acicular crystals, microstalactitic cement, and (3) replacement products: dedolomitization (by concomitant dolomite dissolution and calcite precipitation) and sparmicritization textures.
10. The regressive sequence of diagenetic features is probably a result of the initial emergence of Grand Cayman Island following the middle Miocene. Since this time, the rocks studied herein have remained predominantly in the vadose zone.
11. Superimposed on this regressive sequence is the deposition of caymanite, terra rossa and speleothem deposits, which may have begun forming since the emergence of the island. Caymanite deposition is inferred to have started very soon after emergence began, as some cavities in caymanite contain the regressive cement sequence.
12. Despite the similarities of Tertiary and Quaternary dolostones of other Caribbean areas, dolomitization events were neither synchronous nor necessarily a result of the same processes. The later cement sequences in the dolostones resulted from progressively freshening porewaters, and thus formed from similar processes as a response to fluctuating sea level and/or tectonic movements. The development of these regressive cement sequences was not, however, a synchronous event in the Caribbean.

VIII. PHOTOGRAPHIC PLATES

Plates 1 to 18

Plates 1 to 8

Plates captions contain reference to locality by abbreviation (refer to Figure 1B) for field and hand specimen photographs. Hand specimens have University of Alberta Type Collection numbers, and specimens are housed in the University of Alberta, Paleontological Collections, Edmonton. Field photographs have locality reference only.

Plates 9 to 18

Plate captions contain reference to locality by abbreviation (refer to Figure 1B), followed by catalogue number of the samples.

PLATE 1

Hand specimen and field photographs of *Entobia* in coral molds.

(A) *E.* new ichnospecies A. Note style of branching and density of galleries. No. UA 7793, from locality CK. Scale bar = 2 cm.

(B) *E.* new ichnospecies A and *Trypanites* (?). Note vaguely camerate galleries and buds, which may be beginnings of new branch. No. UA 7794, from locality HRQ. Scale bar = 2 cm.

(C) *E.* new ichnospecies A. No. UA 7795, from locality HRQ. Scale bar = 2 cm.

(D) *E. geometrica* Bromley and D'Alessandro, in *Agaricia* mold. Locality HRQ. Scale bar = 2 cm.

(E) *E. geometrica* Bromley and D'Alessandro, in *Agaricia* (?) mold. Locality HRQ. Scale bar = 2 cm.

(F) *E. cateniformis* Bromley and D'Alessandro, in *Stylophora* mold. No. UA 7796, from locality HRQ. Scale bar = 1 cm. Photograph courtesy of Brian Jones.

(G) *E. cateniformis* Bromley and D'Alessandro, in *Stylophora* mold. Note, part of boring (bottom right hand side) is xenomorphic after *Trypanites*. No. UA 7797, from locality HRQ. Scale bar = 1 cm.


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PLATE 2

Hand specimen and field photographs of *Entobia*.

(A) *E. laquea* Bromley and D'Alessandro, in *Agaricia* mold. Note the development of triangular shaped chambers. Locality HRQ. Scale bar = 1 cm.

(B) *E. mammillata* Bromley and D'Alessandro (m), *E. paradoxa* (Fischer) (p), *E. ovula* Bromley and D'Alessandro (o) and *E. laquea* Bromley and D'Alessandro (l) in stylophorid mold. At least one other *Entobia* sp. in this mold, but not readily visible in this photograph. No. UA 7798, from locality HRQ. Scale bar = 2 cm.

(C) *E. mammillata* Bromley and D'Alessandro (ma) to the right, non camerate *E. megastoma* (Fischer) in rest of photograph. No. UA 7799, from locality HRQ. Scale bar = 2 cm.

(D) Crowded *E. laquea* Bromley and D'Alessandro and *Trypanites weisei*. Locality EEQ. Scale bar = 2 cm. Photograph courtesy of Brian Jones.

(E) Ribbon-like *E. mammillata* Bromley and D'Alessandro, with flattened branches and numerous apertural canals connecting to the wall of stylophorid mold. NO. UA 7800, from locality HRQ. Scale bar = 1 cm.

(F) *E. ovula* Bromley and D'Alessandro, growth phase D, characterised by boxwork of spherical chambers. No. UA 7801, from locality HRQ. Scale bar = 2 cm.

(G) *E. ovula* Bromley and D'Alessandro, which is xenomorphic after *Trypanites* sp. *Trypanites weisei* and *E. laquea* (top left) also present. Locality EEQ. Scale bar = 2 cm.


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PLATE 3

Scanning electron micrographs and hand specimen photographs of *Entobia*.

(A) Scanning electron micrograph showing xenoglyphic sculpture (x) from *Stylophora* sp. on cast of *E. ovula* Bromley and D'Alessandro. Scale bar = 1 mm.

(B) Scanning electron micrograph showing scalloped surface of *E. ovula* Bromley and D'Alessandro. The scallops are a negative impression of the pitted microsculpture produced during boring as a result of the removal of chips of substrate. Scale bar = 200 µm.

(C) *E. ovula* Bromley and D'Alessandro, growth phase C, characterized by open branching network of spherical to sub-cylindrical chambers. Chambers are separated generally by a weak constriction, in contrast to the tight constrictions or intercameral canals of growth phase D. No. UA 7802, from locality HRQ. Scale bar = 2 cm.

(D) *E. ovula* Bromley and D'Alessandro, growth phase C, with strings of chambers generally running parallel to the long axis of the stylophorid mold. No. UA 7803, from locality HRQ. Scale bar = 1 cm.

(E) *E. paradoxa* (fischer) (p) with large irregularly shaped chambers. Also *E. ovula* (o) and *E. laquea* (l). No. UA 7804, from locality CK. Scale bar = 1 cm.

(F) *E. volzi* Bromley and D'Alessandro, showing dense boxwork of small spherical chambers and wide subcylindrical apertural canal (top right). No. UA 7805, from locality HRQ. Scale bar = 1 cm.

(G) *E. paradoxa* (Fischer). No. UA 7806, from locality CK. Scale bar = 1 cm.

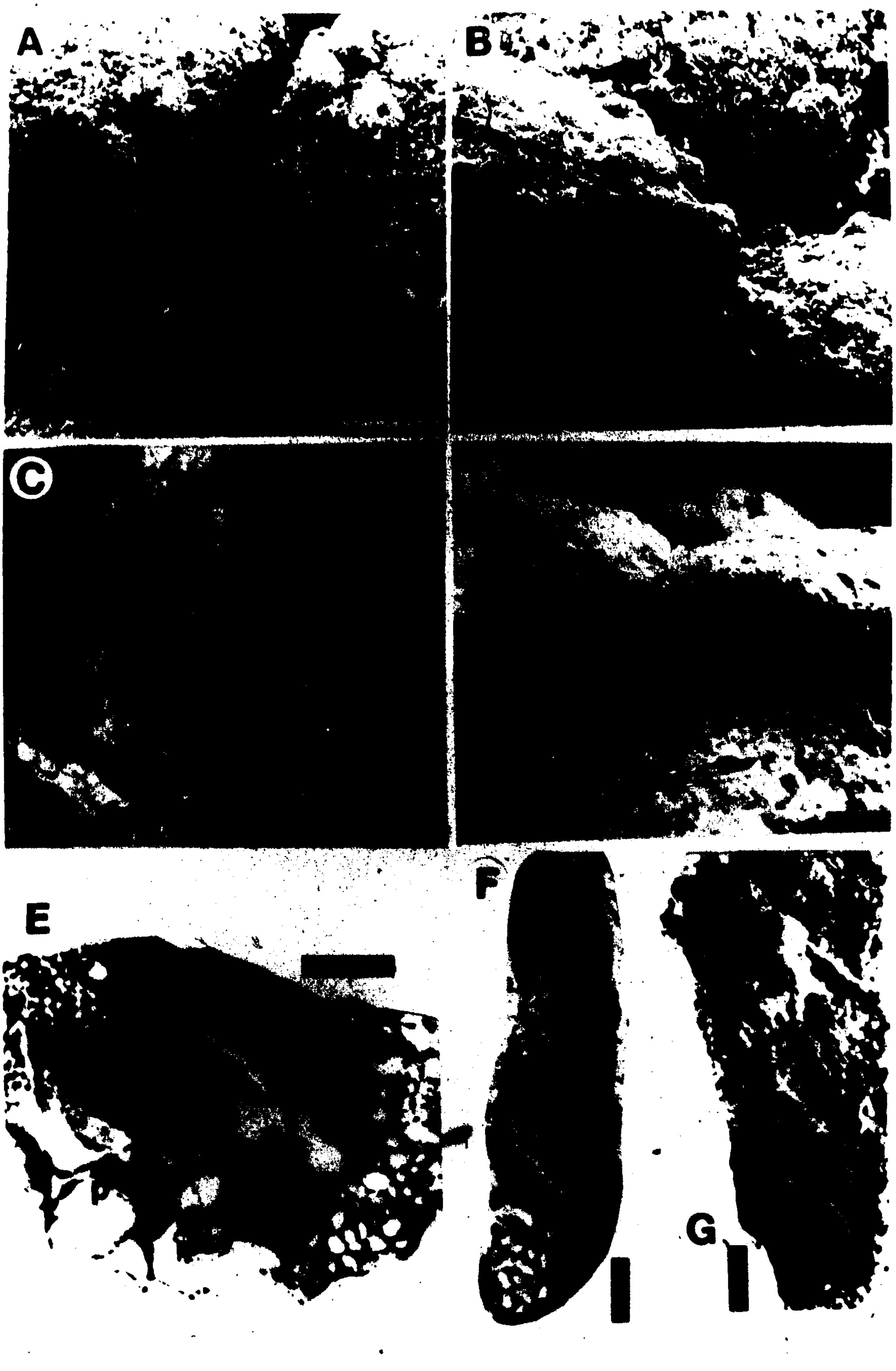
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PLATE 4

Hand specimen and field photographs of new ichnogenus A.

(A) Large new ichnogenus A in massive coral mold, which has been completely leached. No. UA 7807, from locality HRQ. Scale bar = 2 cm. Photograph courtesy of Brian Jones.

(B) Slab impregnated with fibreglassing resin to show section through new ichnogenus A. No. UA 7808, from locality PCQ. Scale bar = 1 cm.

(C) Clusters of new ichnogenus A in massive coral mold. Locality HRQ. Scale bar = 2 cm.

(D) Enlargement of one new ichnogenus A from 14C. Note long apertural canals. Locality HRQ. Scale bar = 1 cm.

(E) New ichnogenus A removed from mold. No. UA 7809, from locality HRQ. Scale bar = 1 cm.

(F) Small new ichnogenus A, with numerous apertural canals, in branching coral mold. Associated with entobians. Locality EEQ. Scale bar = 1 cm. Photograph courtesy of Duncan Smith.

(G) Small new ichnogenus A in stylophorid mold associated with *E. cateniformis* (c) and stenomorphic *E. laquea* (in Background). No. UA 7810, from locality HRQ. Scale bar = 1 cm.

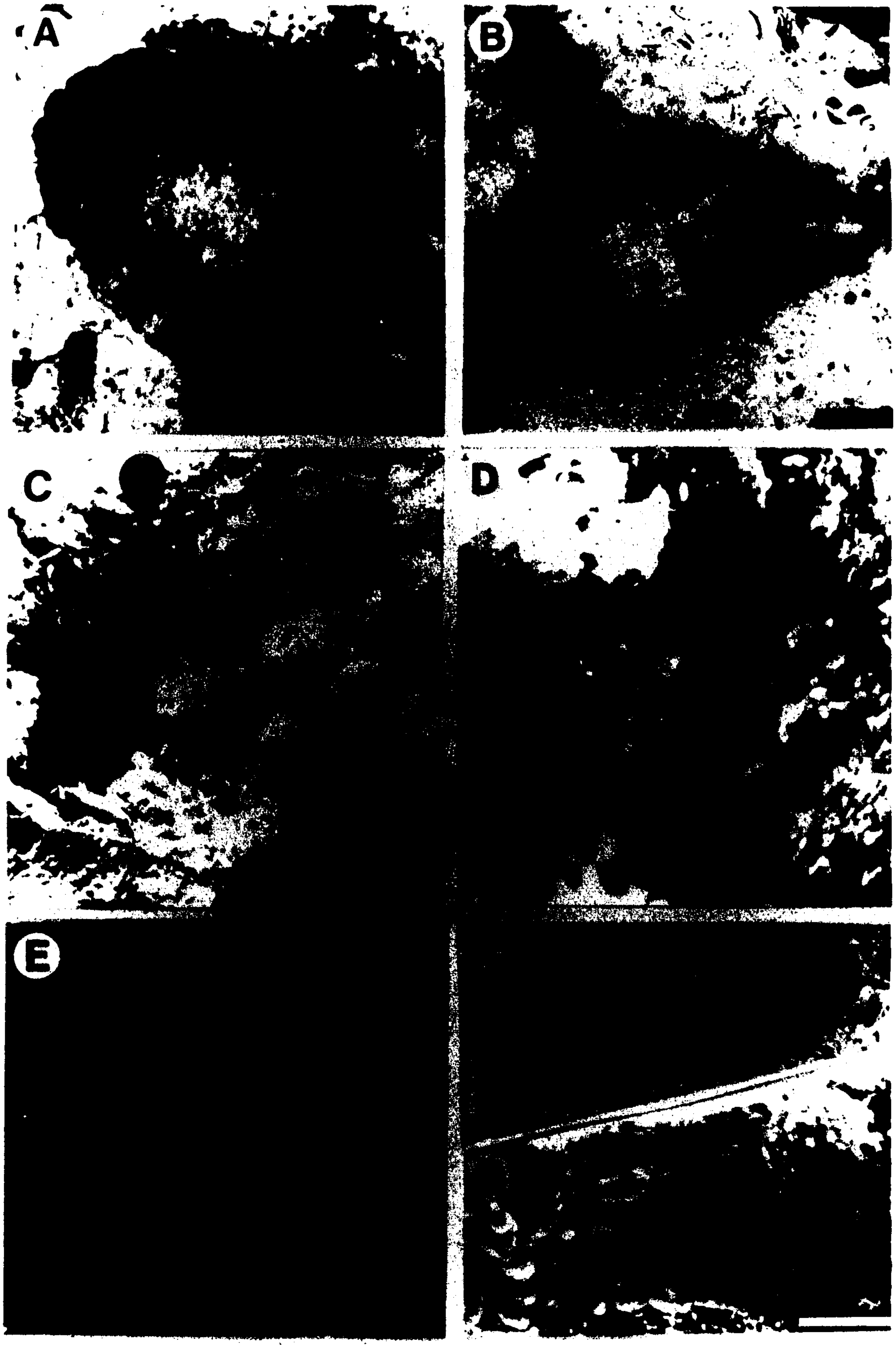
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PLATE 5

Hand specimen and field photographs of *Gastrochaenolites* and new ichnogenus B.

(A) *G. turbinatus* Kelly and Bromley (g) in branching coral mold. Also *Trypanites solitarius*. No. UA 7811, from locality HRQ. Scale bar = 5mm.

(B) *G. torpedo* Kelly and Bromley, removed from massive coral mold. No. UA 7812, from locality HRQ. Scale bar = 1 cm.

(C) *G. torpedo* Kelly and Bromley, removed from massive coral mold. Note constriction just before aperture. No. UA 7813, from locality HRQ. Scale bar = 1 cm.

(D) *G. torpedo* Kelly and Bromley (g) in massive coral mold. Also tunnel-like boring (p), where branching is at right angles to the main tunnel. Locality HRQ. Scale bar = 2 cm.

(E) New ichnogenus B in massive coral mold that has been completely leached. The two borings at the front are *G. torpedo*. Locality HRQ. Scale bar = 5 cm.

(F) New ichnogenus B in massive coral mold. No. UA 7814 - 5, from locality HRQ. Scale bar = 4 cm.

(G) New ichnogenus B with slightly ovate aperture. No. UA 7816, from locality HRQ. Scale bar = 2 cm.

(H) New ichnogenus B with rounded base and circular aperture. Locality HRQ. Scale bar = 1 cm.


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PLATE 6

Hand specimen and field photographs of *Trypanites*, *Maeandropolydora* and *Talpina* .

(A) *Trypanites solitarius* (von Hagenow). No. UA 7817, from locality HRQ. Scale bar = 1 cm.

(B) *Trypanites* sp. Torpedo shaped boring comparable to recent borings of sipunculids. No. UA 7818, from locality HRQ. Scale bar = 5 mm.

(C) *T. weisei* Mägdefrau and *T. solitarius* (von Hagenow) in small coral mold. Locality HRQ. Scale bar = 2 cm.

(D) *Maeandropolydora* cf. *sulcans* Voigt. No. UA 7819, from locality HRQ. Scale bar = 1 cm.

(E) Coral mold lined with *T. weisei* Mägdefrau and *T. solitarius* (von Hagenow). Locality HRQ. Scale bar = 3 cm.

(F) *M.* cf. *sulcans* Voigt, *Maeandropolydora* sp. and *Trypanites* sp. Locality HRQ. Scale bar = 1 cm.

(G) *Talpina* (*eduliformis* ?) in *Stylophora* mold. Also broken *Trypanites*. No. UA 7820, from locality HRQ. Scale bar = 1 cm.

(H) *M.* cf. *sulcans* Voigt (m), *Trypanites* sp. (t) and *Entobia* sp. (e). No. UA 7821, from locality HRQ. Scale bar = 1 cm.


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PLATE 7

Hand specimen, field photographs and scanning electron micrographs of *Caulostrepsis*, problematica borings and microborings.

(A) *Caulostrepsis* cf. *cretacea* Voigt in large completely leached coral mold. Note flattened cross section that distinguishes this ichnogenus from *Trypanites*. Locality QR. Scale bar = 15 mm.

(B,D) Problematica. Large tunnel-like borings. No. UA 7822-3, from locality HRQ. Scale bar for both = 15 mm.

(C) *Caulostrepsis* cf. *cretacea* Voigt. Locality HRQ. Scale bar = 15 mm.

(E) Problematica tunnel-like borings (p) and new ichnogenus B (ib) in large coral mold. New ichnogenus B No. UA 7816. Locality HRQ. Scale bar = 1 cm.

(F) Scanning electron micrograph showing microboring encrusted with small subhedral to euhedral dolomite crystals. Microboring on septal mold of *Diploria*. Scale bar = 4 µm.

(G) Scanning electron micrograph showing hollow microboring. Scale bar = 50 µm.

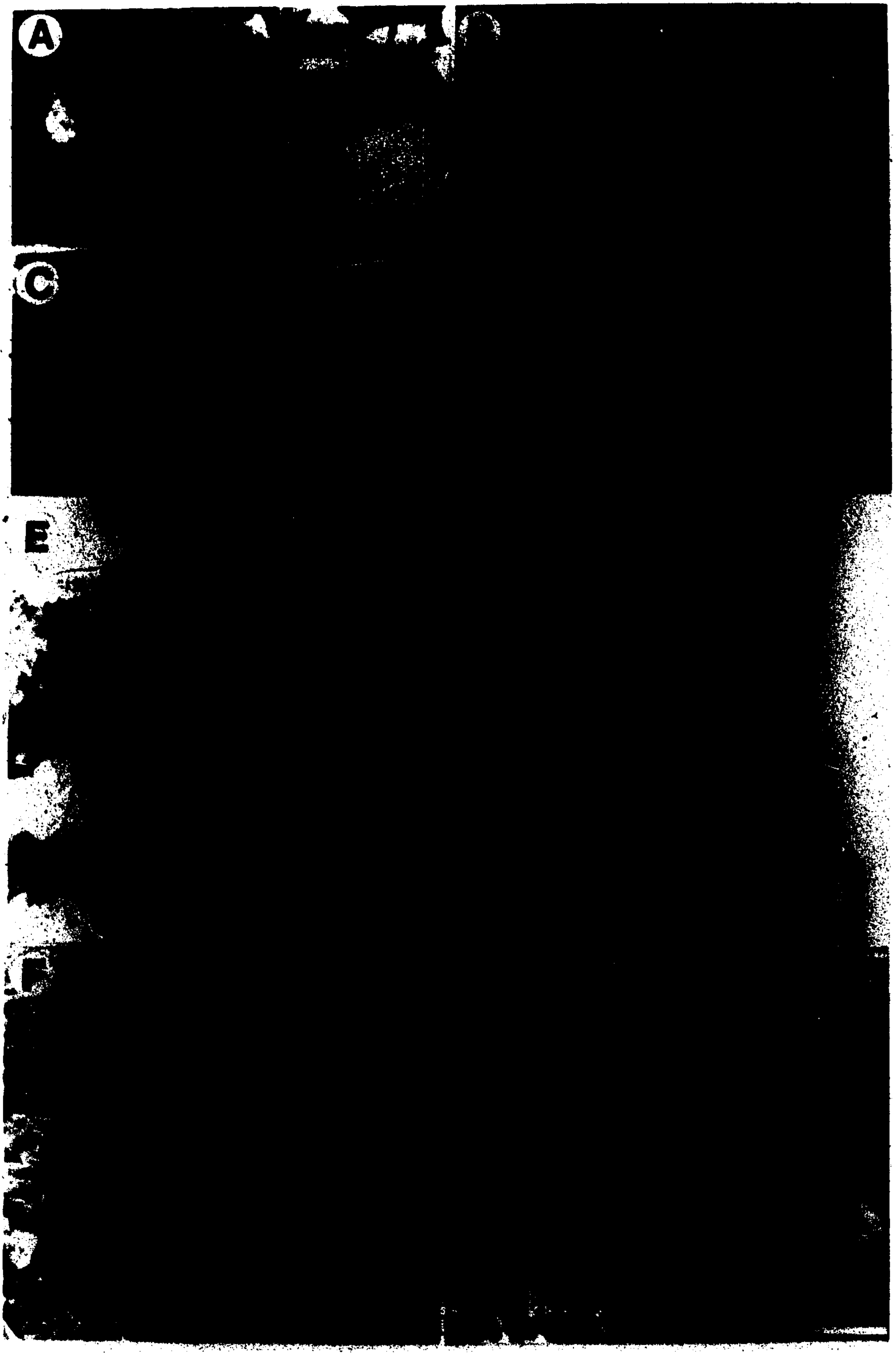
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PLATE 8

Entobian borings in molluscan and coral substrates.

(A) *E. ovula* phase C in bivalve mold. Locality HRQ. Scale bar = 2 cm.

(B) *E. ovula* in gastropod mold. Locality HRQ. Scale bar = 1 cm.

(C) *Entobia* sp. with apertural canals penetrating between septal molds. Scale bar = 1 mm.

(D) *E. ovula* (arrow) following the meandering morphology of *Diploria* . Note the inverse preservation of the coral. Scale bar = 1 cm. Photograph courtesy of Brian Jones.

(E) Entobian (e) encircling corallite mold (c). Scale bar = 1 mm.

(F) Entobian branching around and avoiding small corallites (c) of *Stylophora*. Scale bar = 500 µm.


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PLATE 9

(A) Photomicrograph of well preserved dolomitized red algae. QR 1184. Scale bar = 100 µm.

(B) Photomicrograph of dolomitized red algae with evidence of leaching (c = cavity). HRQ 52. Scale bar = 200 µm.

(C) Photomicrograph of cavity (c) in red algae which has been lined with limpid dolomite cement. Note that dolomite crystals have hollow cores. PBQ 1060. Scale bar = 200 µm.

(D) Scanning electron micrograph of dolomitized red algae. The skeleton has been replaced by small subhedral dolomite rhombs, while the conceptacles have been lined with larger, euhedral dolomite rhombs. HRQ 801. Scale bar = 10 µm.

(E) Photomicrograph of partially preserved, dolomitized foraminifera adjacent to leached foraminifera (c = cavity) in dolomitic groundmass. QR 1182. Scale bar = 200 µm.

(F) Photomicrograph of well preserved, dolomitized foraminifera in dolomitic groundmass. EEQ 1221. Scale bar = 100 µm.

(G) Photomicrograph of echinoid replaced by dolomite. A single dolomite crystal has replaced the spine, while smaller dolomite crystals have precipitated in the original void spaces of the spine. HRQ 52. Scale bar = 200 µm.

(H) Photograph of dolomitized *Montastrea*. The interior of the corallites and the void space between the individual corallites have been preserved as reverse molds ('inverse porosity' of Supko, 1977). HRQ. Scale bar = 17 mm.

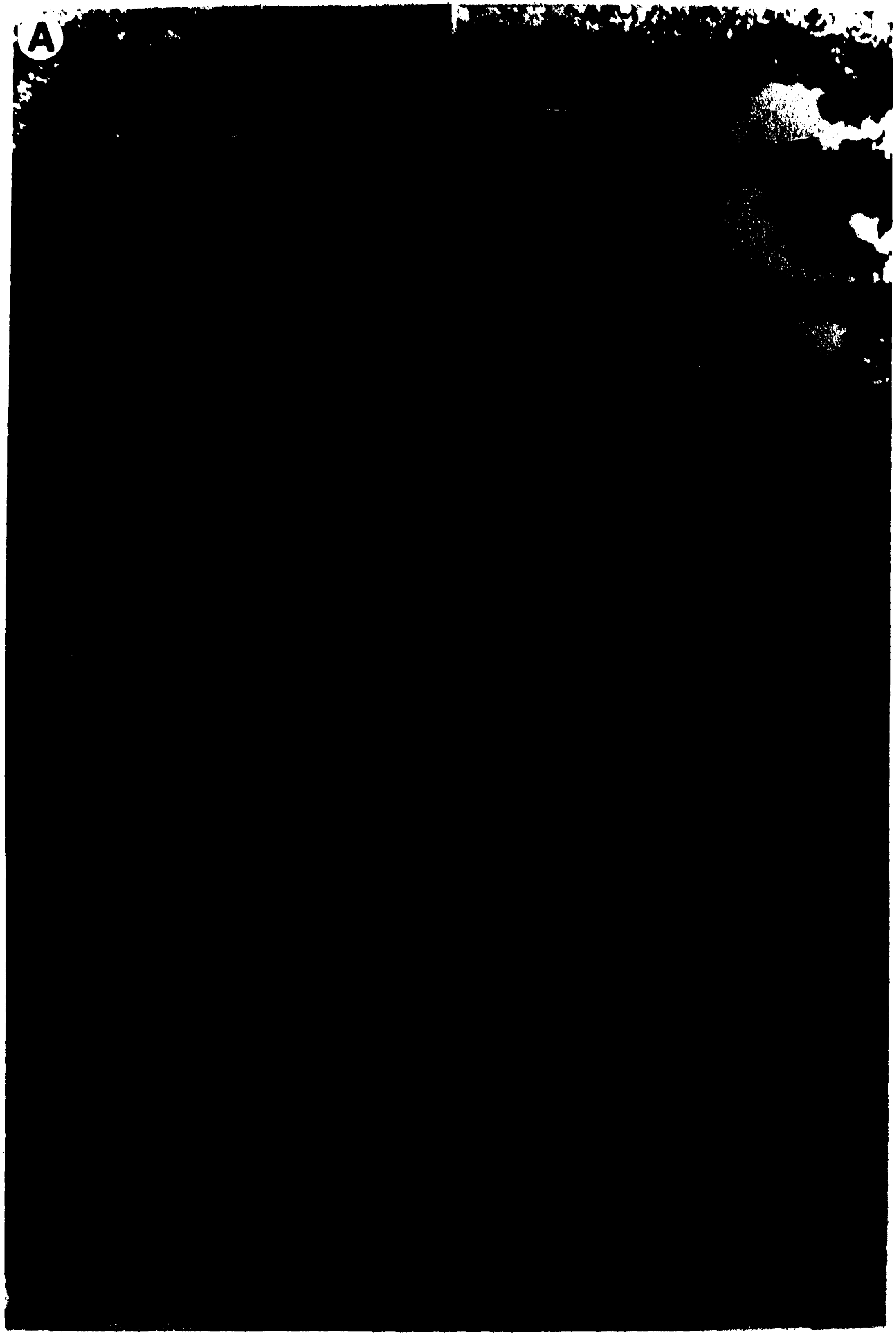
A

PLATE 10

(A) Photomicrograph of dolomitized Amphisteginid foraminifera which has retained the crystallographic orientation of its wall structure. Crossed nichols. PCQ 1243. Scale bar 100 µm.

(B) Photomicrograph showing peloids in coral cavity. Peloids overlie isopachous crystals that coat the coral structures. Both peloids and isopachous crystals are dolomitized. BH 1229. Scale bar = 200 µm.

(C) Photomicrograph showing coral structure filled by internal sediment (s) and coated by isopachous crystals. Whole field of view is composed of dolomite. Crossed nichols. BH 1229. Scale bar = 200 µm.

(D) Photomicrograph of coral septal structure partially filled by dolomitized, geopetal internal sediment (s). Dolomitized isopachous cement (ic) overlies this internal sediment, but does not occlude cavity (c). Dolomitized isopachous cement also coats and lines coral structure. A dark dolomicritic outline (see arrow) represents the edge of the original coral structure. The isopachous cement both within and outside the coral structure is in optical continuity, and overgrows the dolomitic outline. BH 1229. Scale bar = 200 µm.

(E) Photomicrograph of coral structure outlined by dark dolomicrite, filled by internal dolomitized sediment and coated with dolomitized isopachous cement. Crossed nichols. BH 1229. Scale bar = 100µm.

(F) Red algal drape (arrow) well developed on one side of leached stylophorid coral. HRQ 59. Scale bar = 3 cm.

(G) Photomicrograph showing limpid dolomite cement (d) lining cavity. The dolomite is filled and encased by poikilotopic calcite (c) which occludes the cavity. The sparry calcite in the centres of the dolomite rhombs is in optical continuity with the calcite surrounding the rhombs, but not with the dolomite. PCQ 1240. Scale bar = 35 µm.

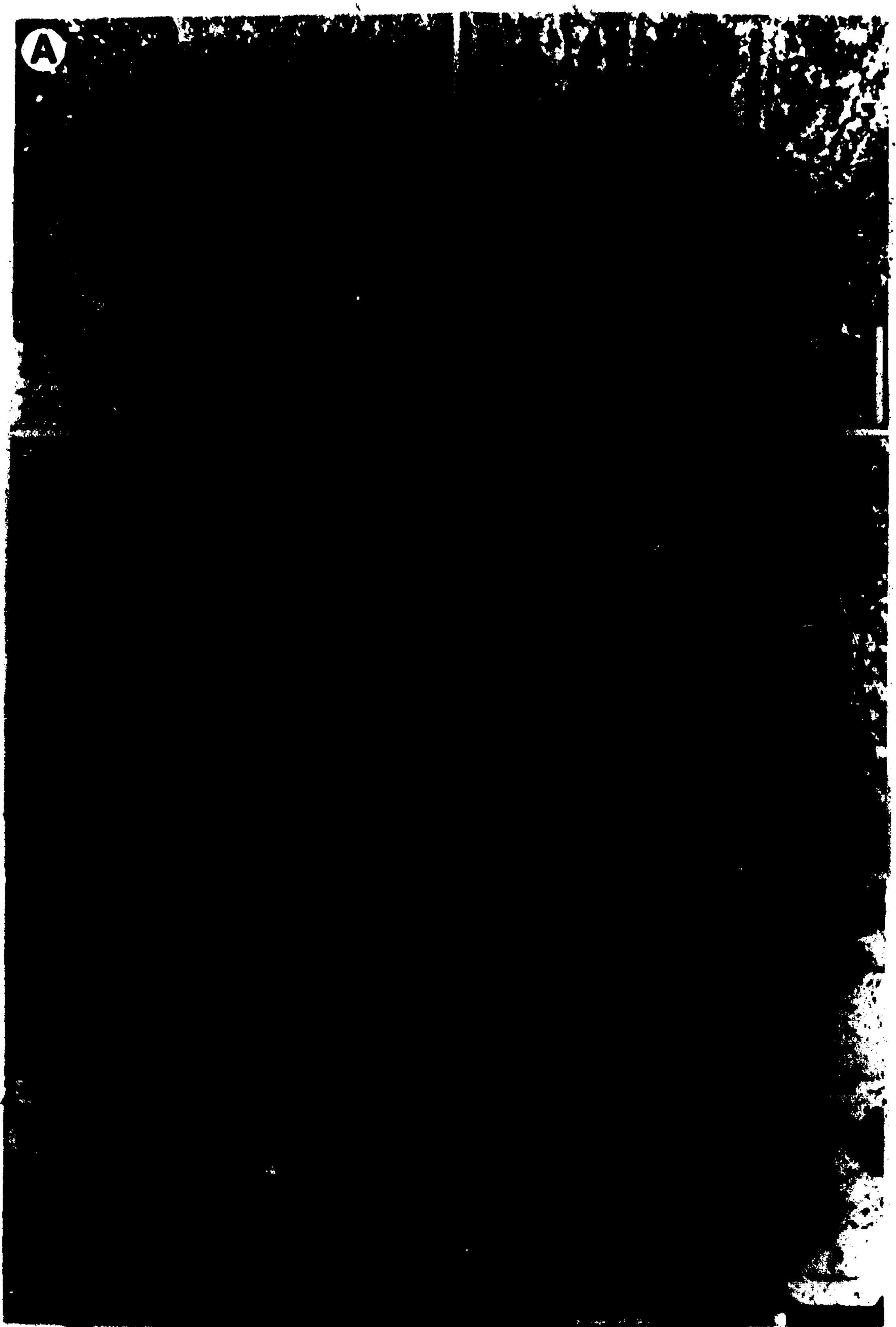
A

PLATE 11

Scanning electron micrographs of dolomite cements (all HRQ 801).

(A) General view of 1st generation limpid dolomite cement lining cavity wall. Leaching is apparent in many of the dolomite crystals. Dolomitic groundmass in the left of micrograph is finely crystalline. Scale bar = 20 µm.

(B) Detailed view of 1st generation dolomite cement showing the etched cores. Scale bar = 10 µm.

(C) Complex intergrowth of 1st generation dolomite crystals. Scale bar = 100 µm.

(D) General view of 2nd generation limpid dolomite cement. Euhedral rhombs are not intergrown with each other. Scale bar = 40 µm.

(E, F) Chasmolithic (?) algal or fungal filaments that have been encrusted with dolomite rhombs. Scale bar for both = 20 µm.

A

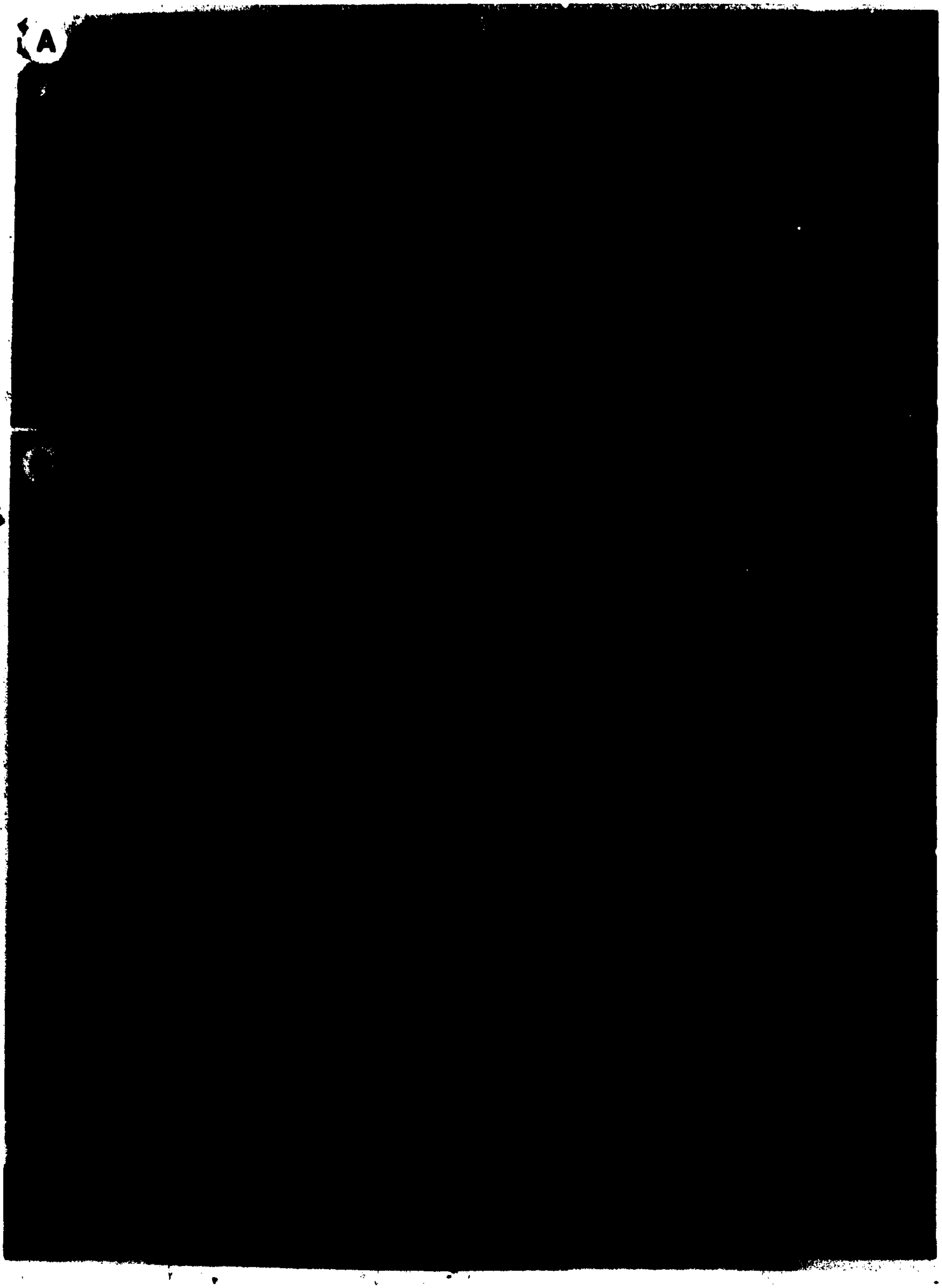

PLATE 12

Scanning electron micrographs of leached dolomite rhombs and poikilotopic calcite.

(A) Leached dolomite rhomb. HRQ 1208. Scale bar = 10 µm.

(B) Hollow dolomite rhomb. Note also recent fungal (?) filament. HRQ 842. Scale bar = 4 µm.

(C) Hollow dolomite rhomb surrounded by partially etched dolomite crystals. HRQ 1208. Scale bar = 20 µm.

(D) General view of cavity lined by large calcite (c) crystals that are encasing smaller dolomite crystals (arrow). Cavity occurs in finely crystalline dolomite (d). HRQ 1212. Scale bar = 100 µm.

(E) Enlarged view of euhedral dolomite (d) crystals encased by a large poikilotopic calcite (c)
crystal. Note that some of the rhombs show evidence of etching. HRQ 1212. Scale bar = 20 µm.

(F) Detailed view of etched dolomite (d) rhomb surrounded by calcite. HRQ 1212. Scale bar = 10 µm.

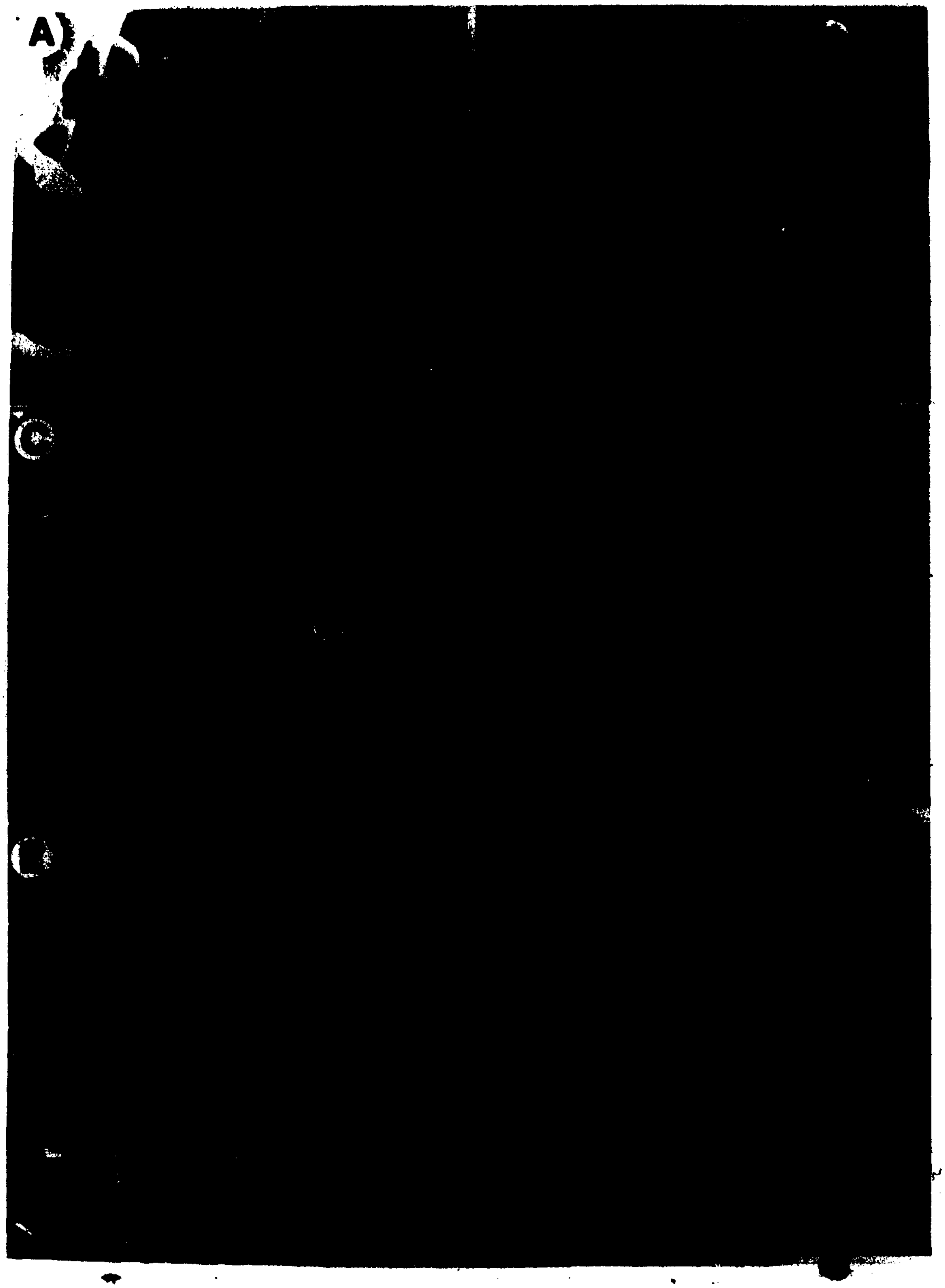
A)

PLATE 13

Colour photomicrographs of dolomite cement and poikilotopic calcite. The thin sections were impregnated with blue epoxy and stained with alizarin red solution. The red is therefore calcite, the unstained portions, dolomite, and the blue, porespace. Note, some white areas within the blue epoxy represent parts of larger cavities that were not totally filled with the epoxy.

(A) Cloudy-centred, clear-rimmed dolomite (arrow) and hollow dolomite rhombs lining cavity. Such a juxtaposition of the rhombs implies that the cloudy centres were less stable than the clear rims, and were therefore leached. PBQ 1060. Scale bar = 100 µm.

(B) Hollow dolomite rhombs lining cavity. PBQ 1060. Scale bar = 50 µm.

(C) Sequence of dolomite and calcite cements in cavity in dolomitic caymanite (d). Unzoned limpid dolomite (ld) lines the cavity. Much larger calcite spar crystals (cs) are rooted on the dolomite cement. The cavity is not occluded by these cements. QR 1184. Scale bar = 30 µm.

(D) Poikilotopic calcite encasing and filling dolomite rhombs. Note some of the dolomite rhombs appear to be floating in the calcite. HRQ 1195. Scale bar = 100 µm.

(E) Porous dolostone, with microvuggy porosity, adjacent to cemented dolostone. The dolomite (d) is poikilotopically enclosed by calcite spar (c) in the right of the photo. This calcite has not occluded the porosity (p) in the left of the photo, producing a soft, porous dolostone. HRQ 1212. Scale bar = 50 µm.

(F) Dolomite rhombs with clear rims, filled and encased by calcite spar. HRQ 1208. Scale bar = 15 µm.

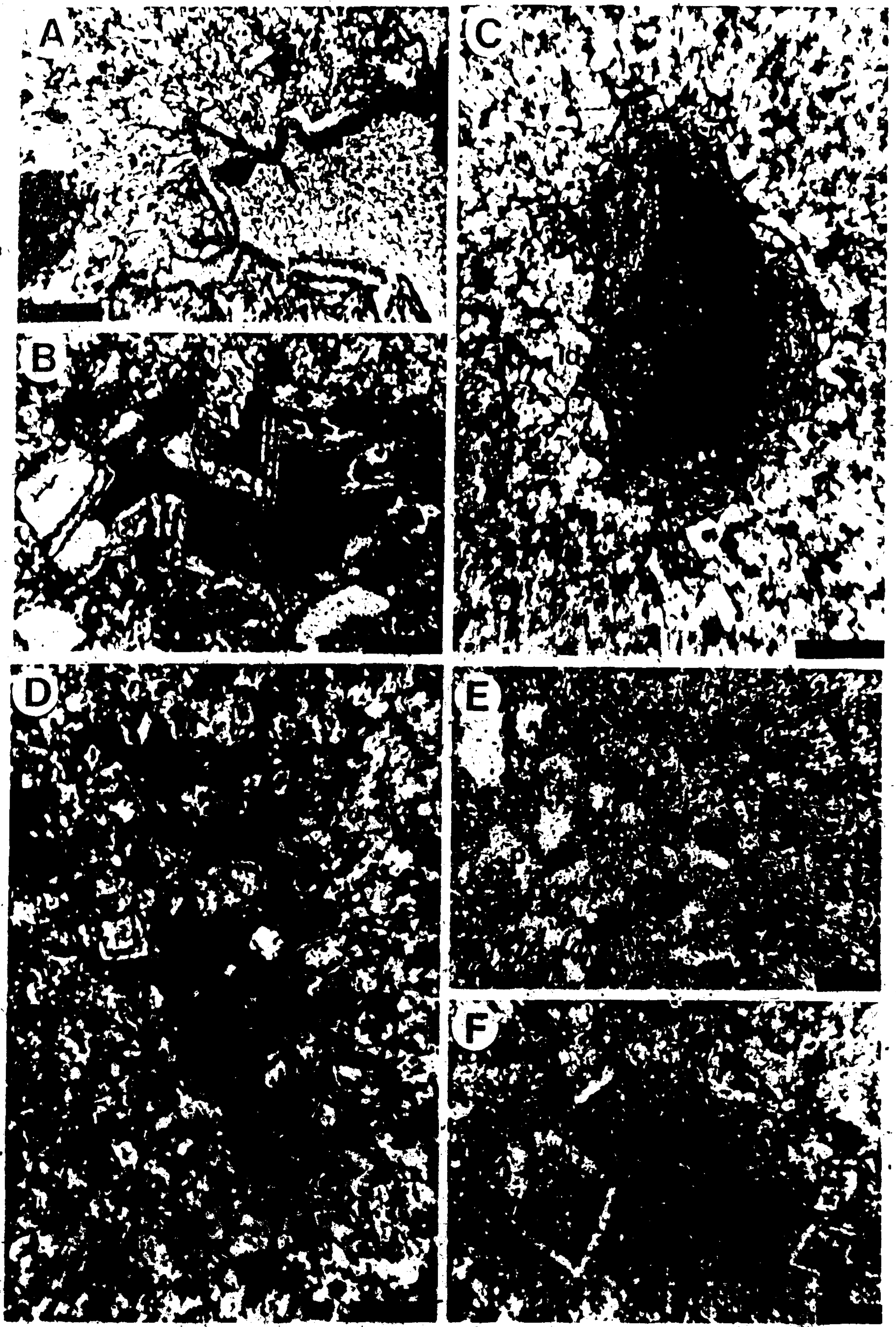
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PLATE 14

Scanning electron micrographs of dissolution features of calcite spar cement (all PBQ 1063).

(A) Well developed spikes in spar calcite crystal. Scale bar = 20 µm.

(B) Spikes developed along microcrack (cleavage plane?) of calcite spar crystal. Spikes are developed on both upper and lower surfaces of the microcrack, and may even join across this crack. Note that well developed spikes in the left of the micrograph grade into more stunted pillars on the right. The whole spar crystal is shown on bottom half of Plate 14C. Scale bar = 100 µm.

(C) General view of spiky calcite. Lower spar crystal shows well developed spikes in microcrack (enlarged in Plate 14B). Upper spar crystals is more extensively spiked. Scale bar = 400 µm.

(D) View of extensively spiked crystal (+marks position on Plate 14C). Original dislocation crack just visible across centre of crystal. Scale bar = 400 µm.

(E) Knobbly texture developed on two faces of spar crystal. The same spar crystal has spikes on other faces. Note the alignment of the textures, being in different directions on each face. Scale bar = 100 µm.

(F) Detailed view of distinct knobbles that grade from the knobbly texture of Plate 14E. Micrograph of same crystal face as left hand face on 14E (just off the micrograph to the left). Knobbles or 'monticuli', are aligned and have begun to form triangular terminations towards right of micrograph. Scale bar = 20 µm.

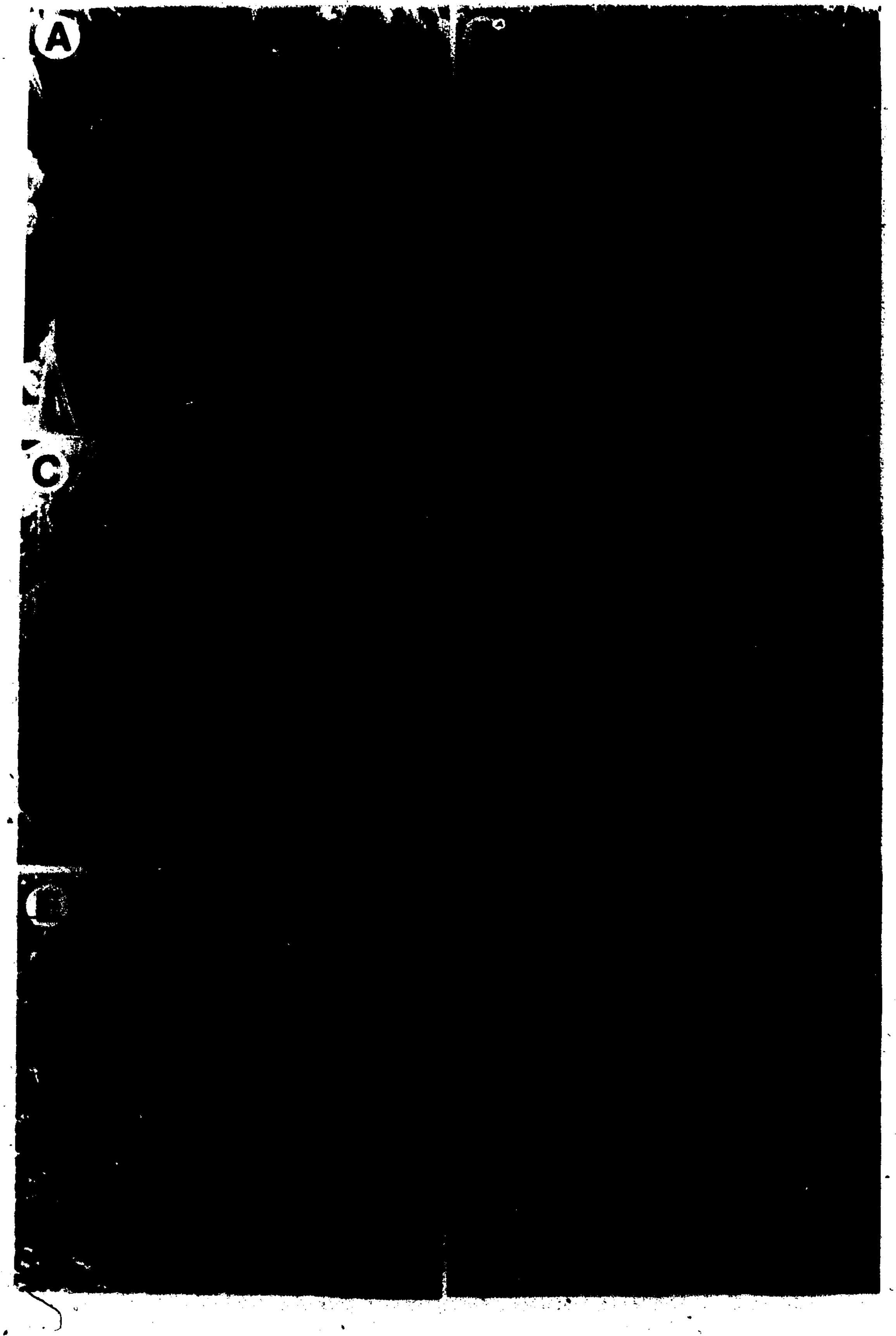

A
C

PLATE 15

Scanning electron micrographs of 'monticuli', columnar crystals and acicular crystals.

(A) Enlargement of elongate knobble or 'monticuli' from centre of Plate 14F. PBQ 1063. Scale bar = 2 µm.

[illegible] Columnar calcite crystals growing from calcite substrate into cavity. Note the distinct boundary between the substrate and crystals. Orientation indeterminable. PBQ 1060. Scale bar = 40 µm.

(C) Dense lining of columnar calcite crystals facing into cavity. Note the pointed terminations of the crystals, which give them a teeth like appearance. Arrow points to broken columnar crystal which exposes the internal structure of the crystal (see Plate 15D). PBQ 1060. Scale bar = 50 µm.

(D) Enlargement of broken calcite crystal from Plate 15C. Seven crystallites surround a central cavity to form the single columnar crystal. PBQ 1060. Scale bar = 10 µm.

(E,F) Calcified fungal (?) filaments with acicular crystals growing perpendicular to the substrate. PBQ 1060. Scale bar for both = 4 µm.

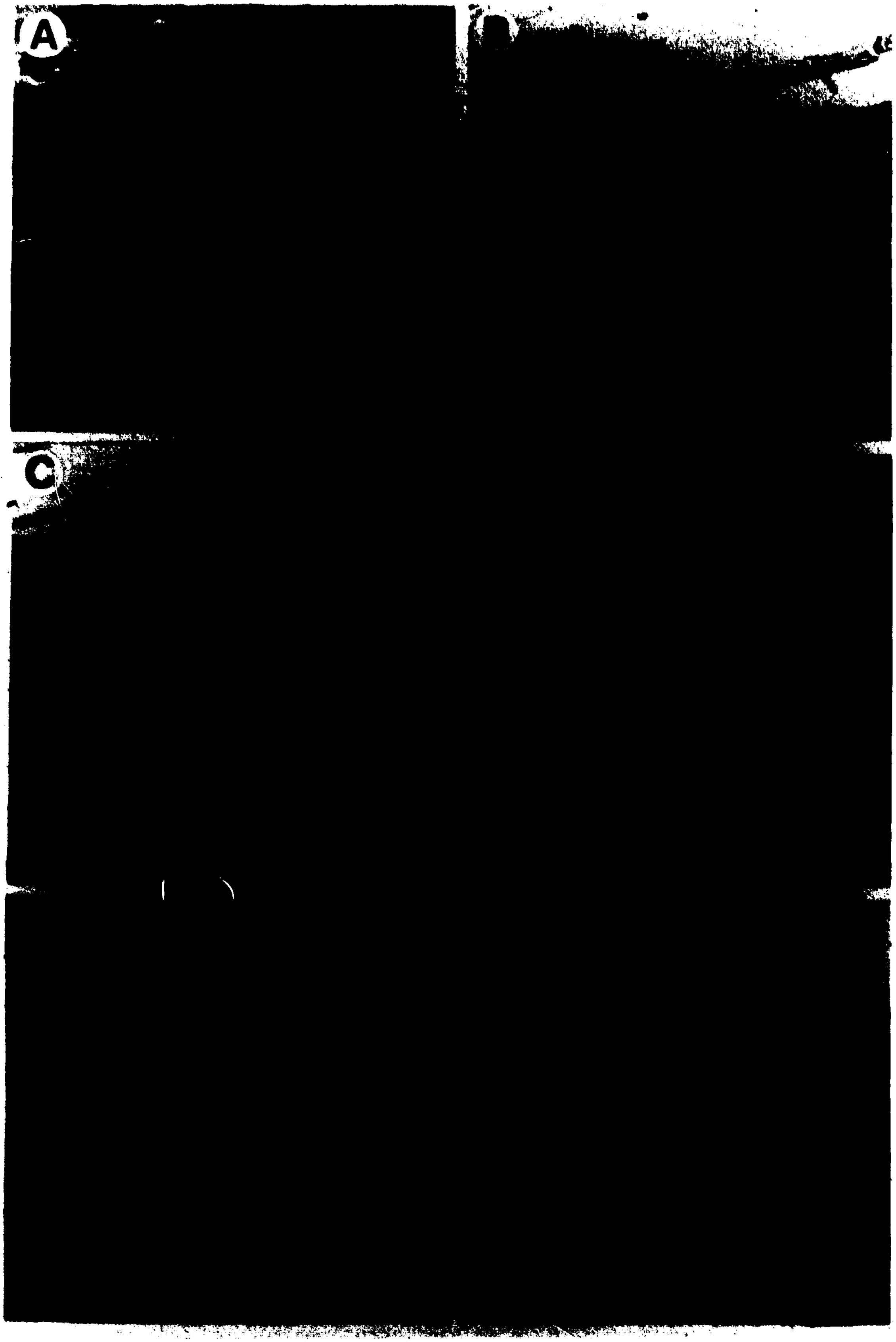

A
C

PLATE 16

Scanning electron micrographs of acicular crystals that occur in coatings on the boring casts and as linings in the coral molds.

(A) Lublinite-type needles irregularly oriented. Associated with Fe and Mn-rich coating on borings. Note the apparent *en echelon* stacking within each crystal. CK 1176. Scale bar = 5 µm.

(B) Detailed view of lublinite-type crystal, having a bladed morphology. CK 1176. Scale bar = 2 µm.

(C) Lublinite type crystals breaking down into rectangular shaped micrite particles. CK 1176. Scale bar = 4 µm.

(D) Needle-fibres associated with the lublinite-type crystals. Note the jagged terminations, bladed morphology and central axial depression. CK 1176. Scale bar = 2 µm.

(E) General view of irregularly oriented needle-fibres associated with the lublinite-type crystals. CK 1176. Scale bar = 10 µm.

(F) Dense needle-fibre bundles as coating on boring cast. HRQ 1213. Scale bar = 20 µm.

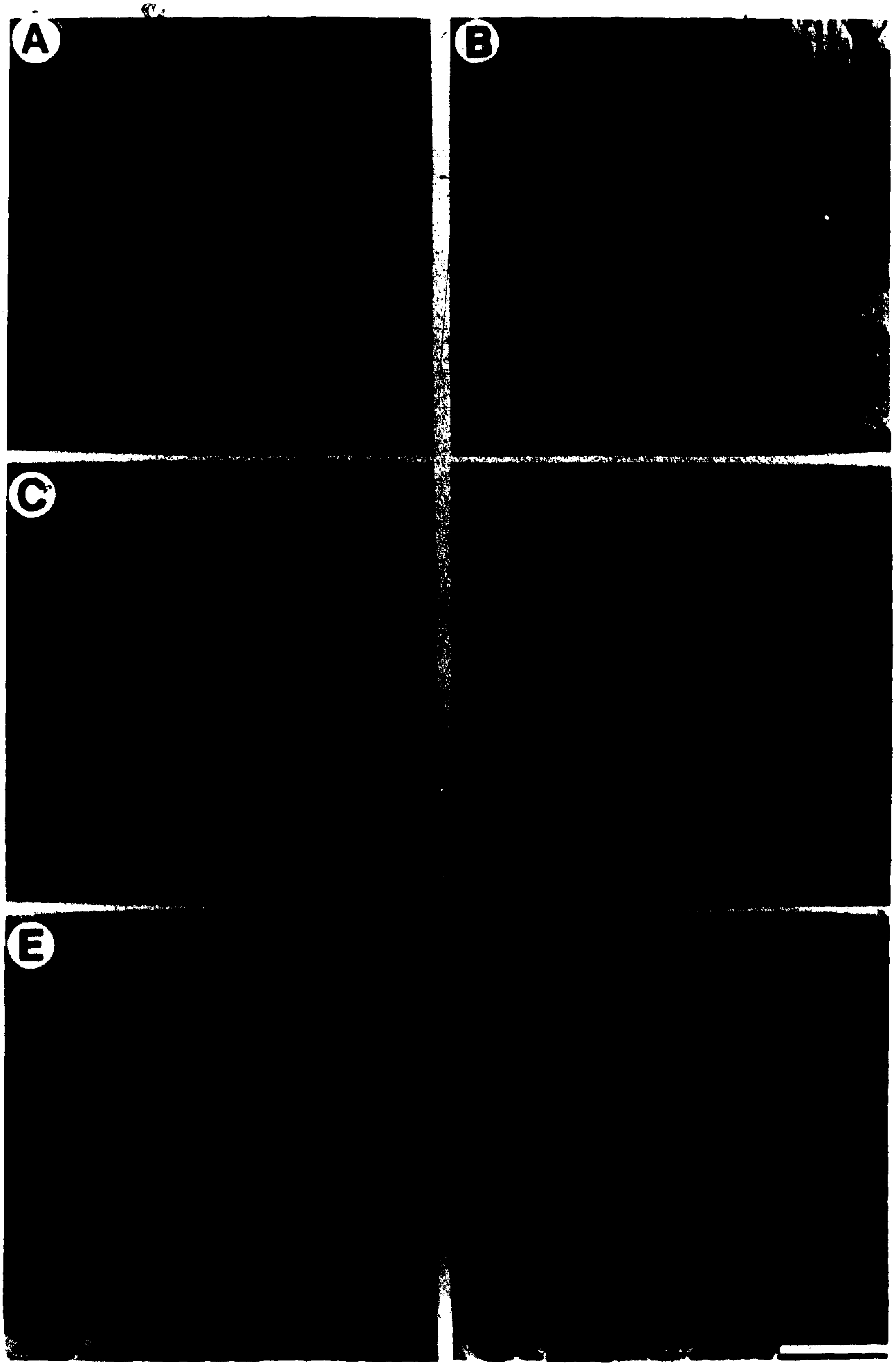

A
B
C
E

PLATE 17

(A) Scanning electron micrograph of needle-fibre bundles enlarged from Plate 16F. Fine micrite binds needle fibre bundles. Needle-fibre crystals composed of a number of fibres (thus needle fibre bundles). In cross section these are equidimensional, in contrast to the bladed needle-fibres illustrated in Plates 16D and 16E. HRQ 1213. Scale bar = 5 µm.

(B) Scanning electron micrograph showing enlarged needle-fibre bundle bound and coated by fine micrite. HRQ 1213. Scale bar = 1 µm.

(C) Scanning electron micrograph of acicular crystals associated with black Fe and Mn rich coating on boring. Crystals oriented perpendicular to substrate, irregularly shaped and much more stout than needle-fibre crystals. CK 1174. Scale bar = 10 µm.

(D) Photomicrograph of microstalactitic cement rooted on a fine dolomite (d) substrate. A few larger limpid dolomite rhombs also act as a substrate. Microstalactitic cement is composed of calcite (c) and is developed hanging from the roof of cavity. The laminae follow the contours of the cavity wall, with protuberances in the wall reflected in the bulbous morphology of the microstalactitic cement. CK 1171. Scale bar = 100 µm.

(E) Photomicrograph showing microstalactitic cement under crossed nichols. The cement exhibits a characteristic sweeping extinction. CK 1171. Scale bar = 100 µm.

(F) Rock slab showing laminated micritic body, aswell as irregular patches of dark micrite, in the host dolostone. HRQ 1208. Scale bar = 2 cm.

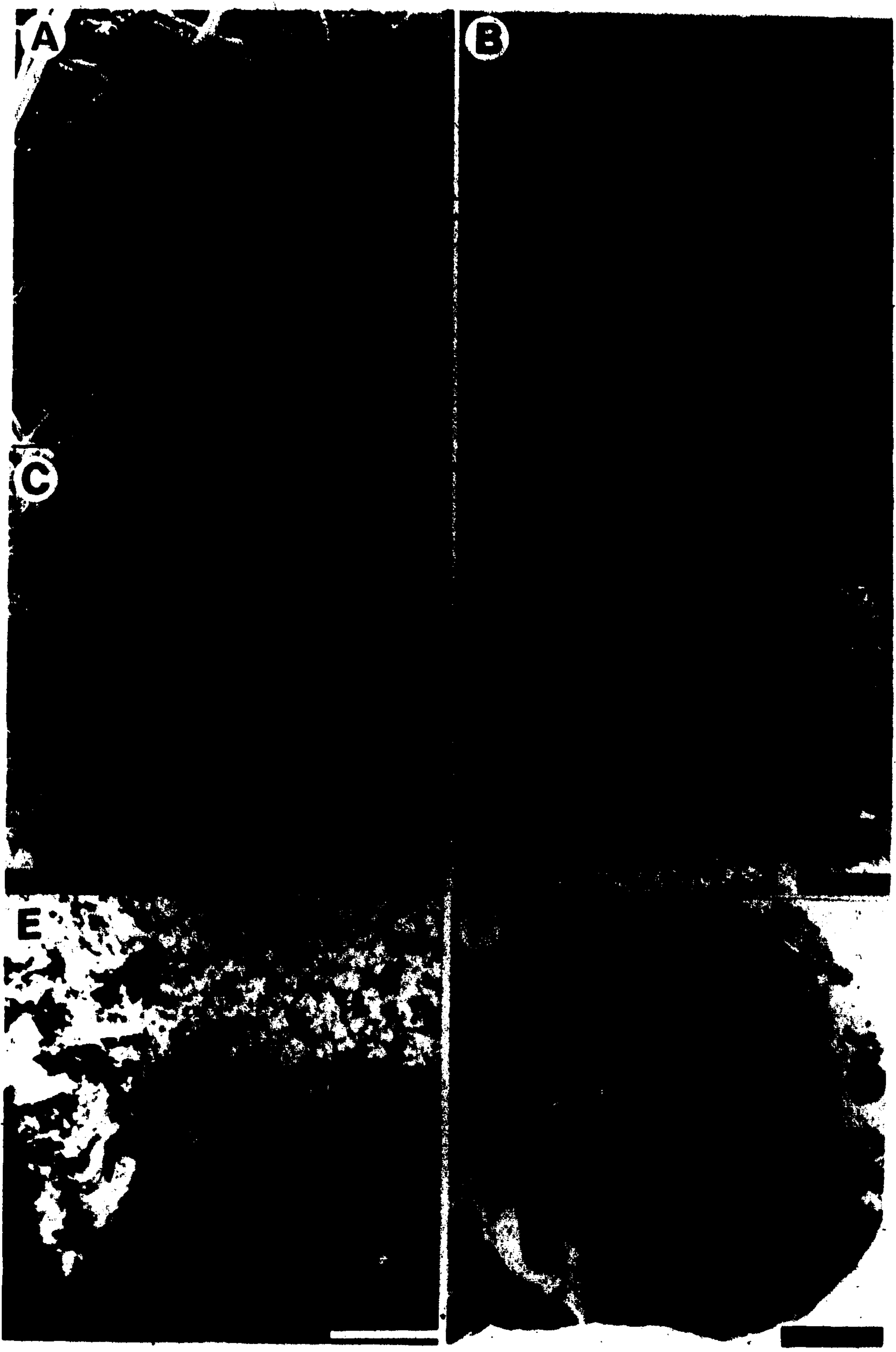

A
B
C
E

PLATE 18

Colour photomicrographs of vadose features and cathodoluminescence photomicrographs of the dolomite. For explanation of colours see Plate 13.

(A) Needle-fibres (nf) lining cavity in the dolostone. Needle-fibres are rooted on dark micrite (m) and dolomite (d). BH 1227. Scale bar = 25 μm.

(B) Microstalactitic cement (mc) hanging from cavity roof. The microstalactitic cement is rooted on dark micrite (m), calcite spar (cs) and type 3 dolomite cement (d). BH 1227. Scale bar = 100μm.

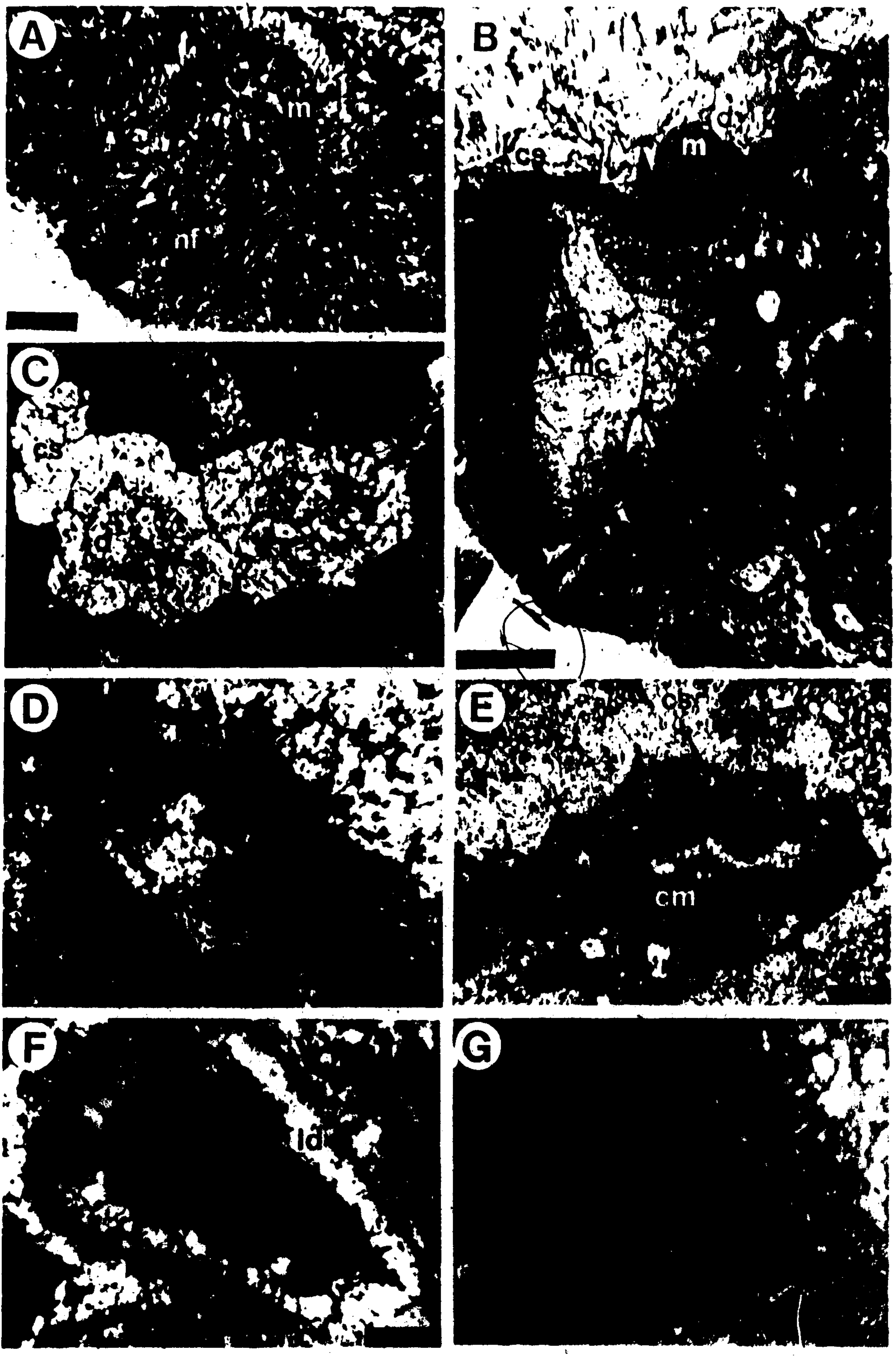
(C) Dark micrite (m) replacing calcite spar (cs). The sparry calcite has poikilotopically enclosed and filled dolomite rhombs (d). HRQ 1208. Scale bar = 50 μm.

(D) Dark micrite in centres, and embaying into edges, of dolomite rhombs. HRQ 1194. Scale bar = 30 μm.

(E) Clotted micrite (cm) apparently replacing calcite spar (cs) that lined a cavity in the dolomitic groundmass (d). HRQ 1208. Scale bar = 100 μm.

(F) Limpid dolomite (ld) lining a cavity in dolomitic groundmass (d). Photomicrograph taken under plane light and to be compared to Plate 18G. HRQ 287. Scale bar = 100 μm.

(G) Same as Plate 18F, but taken under cathodoluminescence. Note, the orange luminescence of the groundmass and the zoning apparent in the limpid dolomite cements. The cements have a dull, non-luminescent core, but bright orange luminescent rims. The boundary between the cements and groundmass is distinct. HRQ 287. Scale bar = 100 μm.


A
m
nf
B
cs
d
m
mc
C
cs
d
D
E
cs
cm
F
ld
G

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X. APPENDICES

Appendix I

Glossary of ichnological terms used in this study (for more extensive glossary see Ekdale *et al.*, 1984):

Bioerosion : this term was first introduced by Neumann (1966) as an abbreviation of 'biological erosion'. The process of penetrating and eroding hard substrates by biological activity. This activity includes borings, raspings, gnawings, scrapings and bites.

Biogenic: biologically produced.

Boring: penetration of a hard substrate (including rock, shell, wood), by mechanical and/or chemical means. No purpose of boring is implied.

Burrowing: penetration of a soft, unconsolidated substrate. For discussion of the terms boring and burrowing see Warme (1975). Obviously, penetration of substrates that are transitionary e.g. firmgrounds, raise problems with such a distinction (Ekdale *et al.*, 1984). Distinction is generally based on whether grains or cement have been transected or not.

Chasmolith: an organism that lives within a cavity in the substrate, also termed chasmoendolith by Golubic *et al.* (1981).

Endolith: an organism that lives within a hard substrate. This is used as a very general term (see Golubic *et al.*, 1981)

Epilith: an organism that lives on a hard substrate.

Euendolith: an endolithio organism that bores its own cavity (Golubic *et al.*, 1981).

Idiomorph: the uninhibited, full morphological development of a trace fossil.

Stenomorph: a trace fossil whose morpholgy is physically restricted or influenced by heterogeneities in the substrate.

Xenomorph: a trace fossil whose morphology is controlled by that of a pre-existing trace.

Xenoglyph: a sculpture produced on the wall of a trace fossil by the characteristic fabric of the host substrate, as opposed to a bioglyph, which is ornament produced on the trace wall by the activity of the organism itself. Xenoglyph is a non-biogenic sculpture or 'fremdskulptur' (Voigt, 1971). Bioglyph is a biogenic sculpture.

Appendix II

Table showing calculated temperatures of dolomite formation using, average, minimum and maximum Bluff Formation dolomite values.

‰ WATER (SMOW)	using average dolomite value 33.3 (SMOW) T °C	using minimum dolomite value 32.3 (SMOW) T °C	using maximum dolomite value 34.2 (SMOW) T °C	$10^3 \ln \alpha \, dol - H_2O = 3.2\,(10^6)\,T^{-2} - 3.3$
4	43	47.5	38.5	
3	38	42	34	
2	33.5	37.5	29.5	
1	29	33	25	
0	24.5	29	21	
-1	20.5	24.5	17	
-2	17	20.5	13.5	
-3	13	16.5	10	
-4	9.5	13	6.5	
-5	6	9	3	
-6	3	6	0	
-7	-0.5	2.5	-3	