University of Alberta

Controls on Modern Reef Development Around Grand Cayman

By

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Nullius addictus jurare in verba magistri

(I am not bound to swear allegiance to the words of any master)

UNIVERSITY OF ALBERTA

FACULTY OF GRADUATE STUDIES AND RESEARCH

The undersigned certify that they have read, and recommended to the Faculty of Graduate Studies and Research for acceptance, a thesis entitled CONTROLS ON MODERN REEF DEVELOPMENT AROUND GRAND CAYMAN submitted by PAUL A. BLANCHON in partial fulfillment of the requirements for the degree of DOCTOR OF PHILOSOPHY.

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August 22nd, 1995

To Ian, Margaret, Angela, and Natalie
for the years lost and the birthdays missed

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Abstract

Like many islands in the Caribbean-Atlantic reef province, Grand Cayman is surrounded by two major reef types: a shallow, wave-dominated fringing reef parallels the windward margin, and a deeper submerged reef rims the edge of the shelf around the entire island. Although the fringing reef is covered by a framework of surf-adapted corals, its core consists of rubble layers capped by crusts of coralline algae. These layers result from cyclic alternation between hurricane-induced reef destruction and rapid re-generation of coral cover during intervals between hurricanes. The shelf-edge reef is also influenced by hurricanes. On margins exposed to the full force of hurricane waves, the reef develops as a closely-spaced series of domal-coral-covered buttresses that are separated by narrow, sediment-floored canyons. On margins protected from hurricane waves, this buttress-canyon architecture does not develop and the shelf-edge reef consists of an unbroken ridge of branching corals. These differences in architecture and coral form are consistent with variations in the degree of coral fragmentation and sand removal induced by hurricane-generated currents and show that shelf-edge reef architecture is controlled by hurricanes.

The presence of flat marine-planation terraces beneath fringing and shelf-edge reefs demonstrates that they are independent structures with a configuration controlled by intrinsic environmental processes, not by antecedent topography as previously thought. Specifically, reef configuration is controlled by the interplay between sea-level rise and coastal gradient. Sea-level rise following the last deglaciation was punctuated by catastrophic rise events at 14.2, 11.5, and 7.5 ka ago that drowned entire shelf systems. As a result, reef and terrace development backstepped to new upslope positions following each rise event. Fringing reefs developed first on low-gradient coasts where planation terraces were wide and coral rubble produced during hurricanes could accumulate. On high-gradient coasts with narrow terraces, rubble was thrown ashore thereby preventing accumulation and reef development. Shelf-edge reefs developed on top of the drowned sea cliffs at the shelf edge which, being relatively isolated from the deleterious influence of sediment accumulation, provided ideal locations for coral growth. Reef development and configuration therefore resulted from the interaction between hurricane-induced processes, episodic steps in sea-level rise, and variations in coastal gradient.

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Natural Resources Unit Staff: from left to right, Phil Bush (and Salty the saline canine), Mike Grundy, Dave Vousden, Scott Slaybaugh, and Ian Young.

1

Introduction

Reefs: Science, Economy, and Environment

Surprisingly, coral reefs have played an important role in the development of the industrialized world during the 20th century. Not only are they primary resources for the tourism and water-sport industries, but their ancient counterparts host prolific quantities of base metals and hydrocarbons. It is no great surprise, therefore, to learn that the primary stimulus for the scientific study of reefs has been economic.

Early accounts of reefs began soon after the discovery of the New World. True scientific exploration only started, however, when the monarchies of Europe realized that, in an increasingly industrialized world, the key to national development was the colonization of new lands with new resources (Bowler, 1993). One of these early economic ventures was the ill-fated voyage of H.M.S. *Bounty* under Captain Bligh in 1788. As Bligh stated:

> The object of all former voyages to the South Seas, undertaken by command of his present majesty, has been the advancement of science and the increase of knowledge. This voyage may be reckoned the first, the intention of which has been to derive benefits from these distant discoveries. (in Bowler, 1993, p.150)

It was, perhaps, inevitable that such voyages would lead to important scientific discoveries and stimulate research. One of the most famous of these was the voyage of H.M.S. *Beagle* in 1831. The observations and discoveries made by the ship's naturalist, Charles Robert Darwin, were to revolutionize the natural sciences and provide a major starting point for the study of coral reefs.

Although such economic ventures stimulated much scientific debate, the new knowledge made little economic impact. Exploration for petroleum (to make kerosene) began shortly after Darwin published on coral reefs, but this work seemingly made little impression on the oil finders themselves. In short "drillers didn't read, and readers didn't drill" (Owen, 1975, p. iv). In fact, application of the reef concept to exploration for hydrocarbons was not made until nearly 50 years after Darwin's death in 1882. Even when immense amounts of oil and gas were discovered in the Ordovician reef-bearing limestones of Ohio and Indiana two years later in 1884, little connection was made between oil and reefs (Owen, 1975). Finally, in 1923, the link was made when european geologists, who had applied Darwin's reef concept to their own Triassic reefs in the Alps (von Richthofen, 1860; Mojsisovics, 1879), were employed by the oil companies in the Middle East. Many of these companies were multinationals, and it was not long before the 'European reef influence' was felt in the United States. At the 1925 and 1926 annual meetings of the American Association of Petroleum Geologists, one of these Europeans, van-der Gracht, applied the reef concept to explain exposures in the Guadalupe Mountains of West Texas and New Mexico. It was not until the 1929 meeting, however, that reefs were put firmly into the oil business: W. Grant Blanchard and Morgan J. Davis demonstrated that Guadalupian-type reefs rimmed the Delaware Basin and continued into the subsurface where they formed the reservoirs of the giant Getty pool in New Mexico and the Henderson pool in West Texas (quoted in Owen, 1975, p 903). Needless to say, Blanchard had been directly exposed to van der Gracht's ideas.

'Reef fever' subsequently gripped the industry and culminated after the 2nd World War with the discovery of oil in Devonian reefs of western Canada. This discovery above all polarized both industry and academia and prompted a major research initiative to study modern reefs. Many companies, notably Shell, immediately set up research laboratories and staffed them with the best academics they could find. Subsequent studies of the Bahama Banks, Jamaica, British Honduras (Belize) and the Persian Gulf, produced rapid advancement and led to the proposal of new rock classifications (Ham, 1962) as well as many of the concepts that are still central to our modern approach.

The synergy between economy and reef science has, in part, led to the massive exploitation of fossil fuels during our 'Hydrocarbon Age.' It must therefore be held responsible, in part, for the problems of environmental pollution and degradation that have fueled fears of rapid climate change. From this perspective it is ironic that, while ancient reefs have provided the oil that helped cause the problem, modern reefs are now providing information and incentive to help solve it. For the first time then, the impetus for the study of reefs is beginning to move away from traditional economic incentives to environmental ones—although it is sad to say that this new impetus seems to be driven more by a fear of economic penalty than of new-found environmental ethics.

Thesis Rationale

Modern reefs are beginning to provide an extremely detailed record of past environmental change—a knowledge that is becoming increasingly important for predicting the consequences of global climate change. Unfortunately, much of this work is bein, .iducted without a basic understanding of reefs or their response to climate change (Fairbanks, 1989; Edwards et al , 1993; Eisenhauer et al., 1993; Peltier, 1994). As a consequence, limitations of these records are poorly understood or simply ignored and important implications are being overlooked.

This thesis is an attempt, therefore, to establish the controls on modern reef development so that records of environmental change contained in reef sequences can be understood more fully. It investigates local and regional environmental processes that presently control, and have influenced, reef morphology, structure, and distribution around Grand Cayman—a small Caribbean island renowned for its reef development. In so doing, it aims to provide a process sedimentologic framework within which the reef development can be viewed. Only from such a perspective can the limitations and implications of new-found coral records be recognized.

Before doing this, however, the scientific development of reef study is briefly reviewed so that the overall contribution of this thesis can be assessed.

Historical Development of Reef Study

Given that there are several exhaustive historical reviews of coral reef development (Davis, 1928, Hopley, 1982; Guilcher 1988), only those theories that introduced important concepts will be reviewed here.

Charting and explaining atolls

Although early Portuguese and Spanish navigators were well acquainted with reefs, these structures were not studied in any great detail until the British and French began to chart the Pacific in the latter half of the eighteenth century (Guilcher, 1988). Even so, early navigators were all too familiar with the three main types: fringing reefs were attached to or 'fringed' the shore closely, barrier reefs were separated by wide lagoons that prevented easy access to shore, and atoll reefs encircled deep lagoons with little or no land areas to speak of. Early navigational charts showed reef development was restricted largely to oceans and seas in tropical latitudes (Dana, 1879). They also showed that atolls and small volcanic islands with barrier reefs were virtually the only land forms present over vast areas of the Pacific and Indian oceans. These exotic and unfamiliar 'lagoon islands' captured the public's imagination but were a puzzlement to early scientists and philosophers. What processes were responsible for creating these 'curious rings of coral' dotted over such vast areas?

On the 27th December 1831, H.M.S. Beagle, a ten gun brig under the command of Captain Robert Fitz Roy RN, set out to chart coastal waters of Chile, Peru, and several coral islands in the Pacific. The return voyage, through the Pacific to New Zealand and Australia and across the Indian ocean to the Cape of Good Hope, was to provide the ship's 22-year-old naturalist, Charles Robert Darwin, with an unparalleled opportunity to study the natural history and geomorphology of the coral seas—an experience that would enable him to formulate revolutionary ideas about evolution and the development of coral reefs.

Having studied under the eminent Professor of Botany, John Stevens Henslow, at Cambridge, Darwin was primarily a biologist by training. But Henslow encouraged him to broaden his knowledge of other subjects, particularly geology, and arranged for one of the most eminent geologists of the time, Adam Sedgwick, to take young Darwin on a field excursion to north Wales. This was Darwin's only first-hand experience with geology before he embarked on the Beagle voyage, and by his own account it made little impact on his thinking (Darwin, 1887). Much more important in this respect was a book that Henslow forwarded to Darwin shortly after the Beagle set sail—Sir Charles Lyell's *Principles of*

Geology (Lyell, 1832). Probably the most important concept outlined in Lyell's book was the theory of "uniformism" and, during the course of the voyage, Darwin was to find repeated evidence to support it (Darwin, 1839).

Subsiding Reef Foundations

Among other things, Lyell's book also summarized what was then known about reefs. It was well established that the depths at which corals could construct reefs was limited to 20-30 fathoms (35-55 m). This being so, Lyell (1832) argued that the unusual form of atolls must represent coral growth on submerged volcanic craters. Darwin (1839) found this to be an unsatisfactory explanation due to the diversity of atoll forms on navigation charts. He knew that these low-lying atolls must at one time have had a foundation within the range of coral growth. He considered it unlikely that the foundation was built to this level by sediment accumulation because of the remoteness from sources of terrigenous sediment. The lack of variation in the elevation of atolls also suggested that uplift was not the cause. Similar problems arose in explaining the morphology of barrier reefs. Darwin (1839) knew that the only difference between barriers and atolls was the central island—morphologically the encircling reefs were identical. But why did barriers, with their deep lagoons, grow up at such a distance from shore? If the sea had eroded a shallow platform into the island shores before they were protected by reefs it would have created steep coastal cliffs—something that Darwin noted was rare on the islands he had visited (Darwin, 1842). Outbuilding of sediments derived from erosion of the island was also improbable because reefless islands invariably had narrow shelves that plunged abruptly into an unfathomable ocean. Darwin became convinced at an early stage in his voyage that the only explanation for these morphologies was that islands had slowly subsided while the reefs grew up and maintained their position at sea level. Initial subsidence of an island fringed by reefs would therefore produce a genetic succession from fringing to barrier to atoll reefs as the island progressively sank and disappeared below the water (Fig. 1.1A). Atolls therefore stood like monuments over old sunken islands, whereas fringing reefs marked young or stable islands.

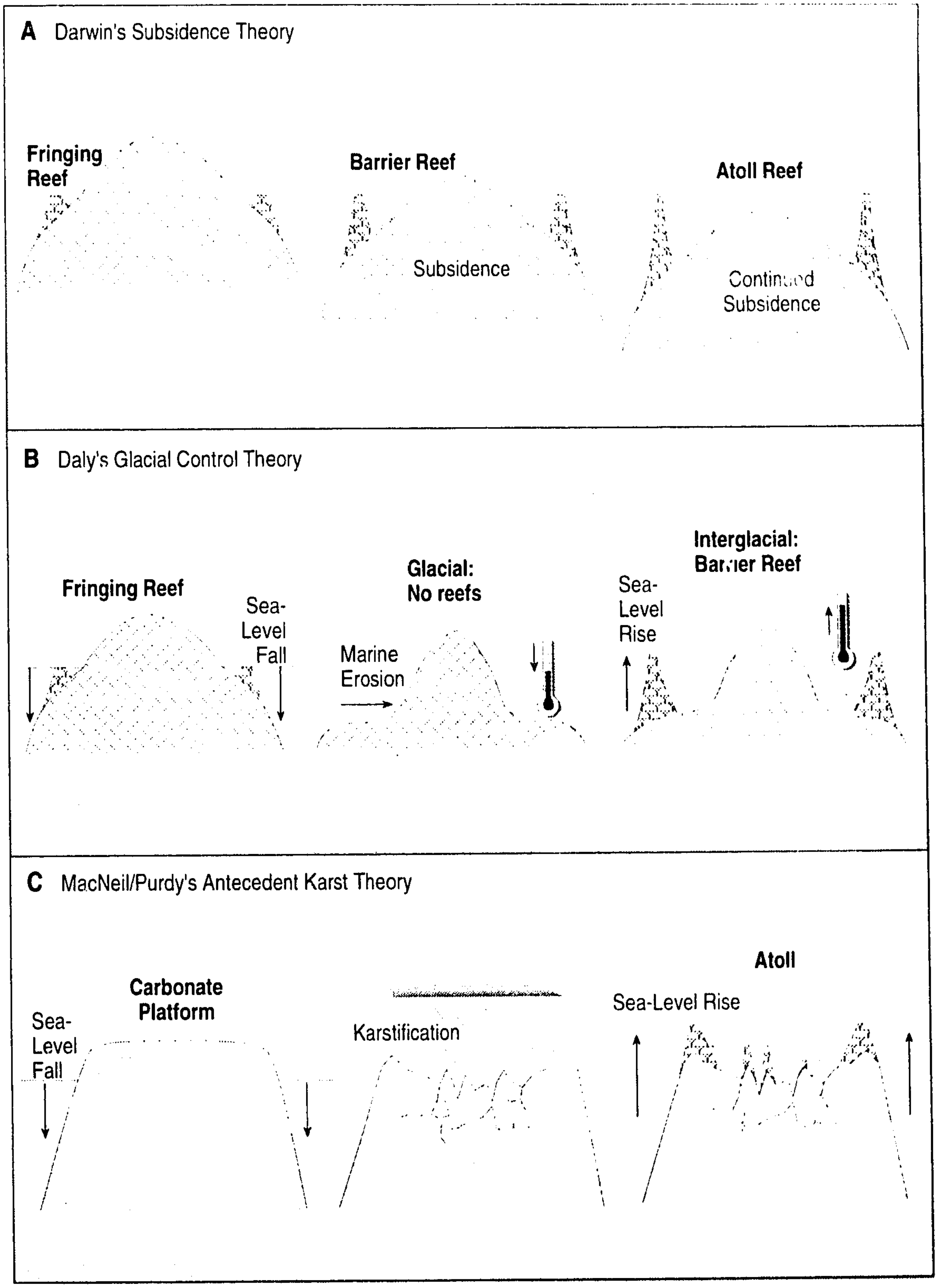

Figure 1.1. The three major theories on the origin of fringing reefs, barrier reefs, and atolls. (A) Darwin's (1842) subsidence theory, (B) Daly's (1915) glacial control theory and, (C) MacNeil's (1954) and Purdy's (1974) antecedent karst theory.

Darwin (1839) initially developed this subsidence theory to account for coral islands in the Pacific. His subsequent visits to the Australian Great Barrier Reef and islands in the Indian Ocean, however, prompted him to extend the subsidence theory to all reefs in all oceans a step that was to lead to much controversy in what later became known as the 'coral reef problem' (Davis, 1928).

The *Beagle* returned to England in October 1836, and immediately Darwin set to work publishing his journal (Darwin, 1839). In May the following year, he presented the new coral reef theory to the Royal Society and, although delayed by illness, went on to publish a book on the subject 5 years later (Darwin, 1842). So convincing was his theory that it remained unchallenged for more than 30 years following its publication. Early support for Darwin's ideas came from James D. Dana (1879), who had also spent several years studying Pacific reefs aboard an American survey ship. Not only did Dana enthusiastically back the theory with his own observations on reefs, but he also provided important evidence for subsidence that Darwin had overlooked: islands that were undergoing active subsidence, Dana argued, would generally lack cliffed shorelines and be dominated by drowned valleys and other embayments (Dana, 1879). Later Davis (1928) would add further morphological evidence for subsidence on coral ringed islands, including a lack of terrigenous detritus in the lagoons from denudation of high islands and the onlapping relationship of the reef deposits with their foundations. Support also came from geological studies that interpreted thick dolostone sequences in the Italian Alps as the accumulation of ancient reef deposits on subsiding foundations (Richthofen, 1860; Mojsisovics, 1879).

The main weakness of Darwin's theory was that evidence for subsidence was circumstantial and, without subsurface data documenting reef build-up over volcanic foundations, any number of theories could account for the different reef types. Shortly before his death, in a letter to one of his principle detractors, Alexander Agassiz, Darwin stated:

> I wish some doubly rich millionaire would take it into his head to have borings made in some of the Pacific and Indian Ocean atolls and bring home cores for slicing from a depth of 500 or 600 feet.

Unfortunately it was not until after Darwin's death (1882) that such borings were made. In the years from 1896-1898 the Royal Society sponsored a drilling project on the Funafuti Atoll to test the subsidence hypothesis. The well bottomed in shallow-water dolostones at 343 m, a fact which strongly suggested that subsidence had produced an anomalously thick shallow-water sequence. But many were unconvinced and wanted to see volcanic basement before they accepted island subsidence.

Although Darwin's first postulate, subsidence of atolls, had been partially confirmed, his second, that reefs developed from fringing to barrier to atoll reefs, had not. And, as Vaughan (1919 p.325) stated: "Although the theoretical possibility of the conversion of a fringing reef into a barrier and a barrier reef into an atoll may not be denied, no instance of such a conversion has yet been discovered." Other complaints suggested that appealing to widespread subsidence in continental and other oceanic areas, such as the West Indies, was unjustified. The small thickness of elevated reefs also showed that postulating great thicknesses for barrier reefs was not supported by the evidence (Guppy, 1888). More seriously, evidence that fringing reefs transformed laterally into barrier reefs could not be explained by simple subsidence alone (Semper, 1881). As more survey charts became available it also became increasingly apparent that the depths of marine shelves and lagoon floors were too uniform and might be expected to show more variation if tectonics was involved.

Effects of glaciations and sea-level changes

Reginald A. Daly (1915) was impressed by the apparent uniformity of lagoon depths and reasoned that it was inconsistent with the subsidence origin of reefs which would produce lagoons with different depths. Instead he postulated that marine abrasion of unprotected volcanic substrates during episodes of glacially lowered sea level prepared the platforms for reef growth during interglacials. Marine abrasion was an old but persistent idea formerly outlined by Wharton (1890) and later by Agassiz (1896) and Gardiner (1904). The incorporation of sea-level changes due to glaciation, however, was new, although it had been briefly noted earlier by Belt (1874), Upham (1878), and Penck (1894). To link marine abrasion and eustatic sea level changes. Daly (1915) suggested that colder ocean temperatures during glacials inhibited reef growth and left island slopes prone to wave attack causing erosion. Rising temperature and sea level during interglacials enabled reefs to initiate upward growth on the outer edges of the eroded platforms, where environmental conditions were optimal (Fig. 1.1B). Thus, Daly's argument was that the different reef types were not genetically related, but instead indicated the extent to which the substrate had been eroded; partial erosion produced barrier reefs whereas complete truncation produced atolls. Daly therefore believed that barrier reefs and atolls were solely a product of eustatic sea-level change and that only fringing reefs existed in times when sea level was stable (Purdy 1974).

Many initially saw this Glacial Control theory as a powerful alternative to Darwin's subsidence idea. Vaughan (1919), who studied Caribbean reefs, was probably the most notable of these. He showed that Caribbean reefs were distinctly different from their Pacific cousins. They generally grew in shallow water and lay some distance from the edge of their underlying platforms, unlike narrow Pacific reefs which were flanked by deep water. Vaughan (1919) thought these platforms were pre-existing erosional features that had become submerged when the glacial ice-sheets had melted. Like Daly, he suggested that the width of the platforms was only an indication of the variable rate of local erosion.

The main weakness with Daly's theory was that it required a number of unsubstantiated postulates to work. To remove reef protection during glacial lowstands it needed ocean cooling. Only then could widespread erosion prepare platforms for reef growth during subsequent interglacial sea-level rises. Davis (1928) argued that erosion could not have been widespread in the Pacific because of the absence of cliffs on barrier-reef islands. But the presence of islands with cliffs in more marginal seas did indicate that some reefs had been affected by ocean cooling and abrasion processes. Interestingly, Davis (1928) considered the Caribbean to be one of these 'marginal' areas because many of the islands had well developed cliff lines. Also, although reefs did re-establish during the subsequent warmings, they were according to Davis (1928, p.118) "...narrow and timid Postglacial novices...by no means comparable with the stalwart veteran barriers of the Pacific coral seas."

Although it was undoubtedly true that subsidence played an important long term role in the development of Pacific atolls, especially the atoll structure itself, the glacio-eustatic sea-level changes introduced by Daly were of far more significance for the development of coral reefs in the short-term. But if Daly's ideas about widespread marine erosion were inapplicable to most areas, what process controlled the configuration of reefs?

Karst Residuals

In the 1940's and 50's several Japanese scientists suggested that the emerged remnants of older reefs and karst surfaces would provide a foundation for modern atoll reefs when sea level rose during deglaciation (Yabe, 1942; Asano, 1942; Tayama, 1952). This hypothesis was explored further by MacNeil (1954), who pointed out that during sea-level lowstands the interiors of carbonate platforms would be lowered by karst weathering, whereas the periphery would be affected by case hardening (cementation) producing a more resistant raised rim. Upon the subsequent flooding of the platform during deglacial sea level rise, these topographic residuals would provide the foundations for reef growth: reefs would form over the peripheral rims and lagoonal patch reefs would reflect the karst residuals of the interior. In essence, karst residuals would provide a topographic blueprint for subsequent reef growth (Fig. 1.1C).

The karst residual explanation of atolls was a radical departure from the view that reef morphology was controlled principally by growth processes during relative sea-level change. But like previous theories, the evidence was largely circumstantial and, although MacNeil suggested the theory could be applied to barrier reefs in other settings, he provided little corroborating evidence. The extension of this idea to other reefs, especially barriers, was skillfully made by Purdy (1974). Using seismic profiles and bore hole data, Purdy suggested that the morphology of the Belize Barrier Reef and its lagoonal patch reefs were determined by the positions of topographic residuals in the underlying bedrock. These residuals, he argued, were morphologically akin to tropical karst landforms that developed on a marginal carbonate plain, and supported the analogy by documenting drowned sink holes in the Belize lagoon. To show that karst processes could account for reef morphologies he refined acid-rain experiments devised previously by Hoffmeister and Ladd (1945) who had simulated the development of peripheral rims on limestone blocks. By varying the rate of simulated rainfall he produced several atoll-like surfaces. A balanced supply rate of acid rain gave a gullied rim with a flat interior, an undersupply gave a gullied rim and an interior with residual pinnacles, and an oversupply gave no rim but just lowered the block surface. As well as simulating atoll rims and lagoonal basins, Purdy suggested that solution breaches in the rim were analogous to reef passes and that the smaller scale gullies on the front of the rim were replicas of reefal spur-and-groove structures. Although reef growth accentuated these features, Purdy (1974) argued that reefs had little to do with the basic configuration, which was essentially karst induced.

Although elegant, Purdy's assessment of the importance of karst in the development and configuration of reefs was overstated and the evidence circumstantial. As Bloom (1974) noted, "...a reef complex is not a homogeneous mass of fine-grained limestone..." and young Pleistocene limestone substrates are extremely porous and permeable. Instead of run-off and overland flow, as simulated by the acid-rain experiments, the dominant mode of water movement on Pleistocene limestone terrains was infiltration and by-pass, especially in cavernous reefal facies (Hopley, 1982). For this reason, emergent Pleistocene reefs would not be degraded by karst processes, but would be topographically accentuated by them (Bloom, 1974). Thus, Purdy's karst residuals underlying modern reefs in the Belize Barrier Reef lagoon could just as easily be residuals of earlier Pleistocene reefs. In fact several studies have shown that some are just that (Halley et al., 1977, Shinn et al., 1977; Multer and Zankle, 1988). Only in older carbonate terrains could karst development produce the requisite land forms necessary to account for reef morphology. Yet the presence of barrier reefs around islands that lack any significant carbonate bedrock—and there are several (see Guilcher 1988)—clearly show that reef configurations are controlled by other non-karst processes. This has been suggested by Choi and Ginsburg (1982) who inferred siliciclastic deposits as the foundation for a significant part of the Belize Barrier. In short, what these observations demonstrate is that *some* shallow water reefs mainly those in lagoonal settings—are nucleated on topographic residuals that have a variety of origins.

New perspectives on modern reefs

The importance of topographic residuals in explaining reef con guration is still a contentious issue today. As alluded to, many workers insist that reefs owe their siting and architecture to topographic residuals (Purdy, 1974; Bloom, 1974; Braithwaite, 1982, 1987; Rosen, 1990; Tucker and Wright, 1990), whereas others suggest that reef growth alone is responsible (Adey 1978; Shinn, 1982; Hopley, 1982; Macintyre, 1988). In the Pacific, only data from the terrestrial parts of reef complexes such as coral cays and the reef crest are available, making an assessment of reef configuration difficult (see Hopley, 1982, for summary). In the Caribbean, however, a substantial amount of data on the anatomy and foundations of Caribbean reefs has been collected over the last 20 years from more seaward parts of the reef (see Macintyre, 1988, for summary). In light of the controversy over the controls on reef configuration, it is important therefore to briefly review these data.

Belize Barrier Reef: The largest reef complex in the Caribbean-Atlantic province, stretching for some 250 km and ranging from 10-32 km wide, is found along the edge of the Belize shelf (James and Ginsburg, 1979; Rutzler and Macintyre, 1982). This almost unbroken barrier reef begins as a fringing reef along the shores of Ambergris Cay and extends south to the Gulf of Honduras. Just beyond Ambergris Cay, the reef tract becomes a barrier reef as the lagoon width abruptly increases to ~20 km. At that point, however, the depth of the lagoon only averages a few metres and so, in a Darwinian sense, the reef is not a true barrier reef. By the time the reef reaches the Gulf of Honduras, however, the lagoon is almost twice as wide and up to 70 m deep—this is therefore a true barrier reef by any standard.

The reef complex itself, which consists of a lagoonal sediment apron and a seaward reef tract, rims the outer shelf and is fronted by a vertical escarpment that forms the upper part of the continental slope (James and Ginsburg, 1979). In the north and central sections, the reef complex is 3-10-km-wide, but as the lagoon deepens to the south the complex narrows to less than 100 m at its southern end. Much of this variation is accommodated by the lagoonal sediment apron as the reef tract seldom attains more than 1 km in width. Yet despite this narrowing, the reef complex maintains roughly the same volume along its full length (Stoddart, 1962; Steers and Stoddart, 1977).

The reef tract, which starts at the reef crest and extends to the shelf break, is a narrow belt of luxuriant coral growth that consists of two general zones of reef development (Fig. 1.2). The inner zone, which extends approximately 300 m seaward from the reef crest, consists predominantly of a shallow-water spur-and-groove structure that is terminated by a steep escarpment with 10-30 m of relief (Burke, 1982). Excavations into the escarpment and the spur-and-groove shows that they are constructional and composed of a deeper-water head and platy coral association that is up to 23 m thick (James and Ginsburg, 1979; Shinn et al., 1982). Although no foundation has yet been penetrated directly beneath the spur-and-groove zone, drilling in the back reef area reached a Pleistocene limestone ridge at ~17 m below msl. Dating shows that coral growth initiated on this ridge at ~7.6 calendar ka when rising seas first flooded the shelf, and an essentially modern zonation (although notably lacking in *Acropora palmata*) was established on the inner reef zone by 6-7 cal. ka (James and Ginsburg, 1979; Shinn et al., 1982).

The outer reef zone, usually separated from the inner by a gently sloping sand plain or rock pavement, consists of a coral-covered ridge having "a luxuriance that equals that of the shallow-water spurs and grooves" (James and Ginsburg, 1979, p. 63). Reconnaissance studies show that this ridge rises to within 15 m of sea level and, in places has up to 35 m of relief (Burke, 1982). A 12-m core into the crown of this reef showed it to be completely composed of a deep-water coral association (Macintyre et al., 1982; Fig. 1-2).

Reefs of St. Croix: The shallow shelf around St. Croix is ~1 to 12 km wide and consists of two flat, gently-sloping, bedrock terraces (0-10 m and 13-18+ m below msl) that provide

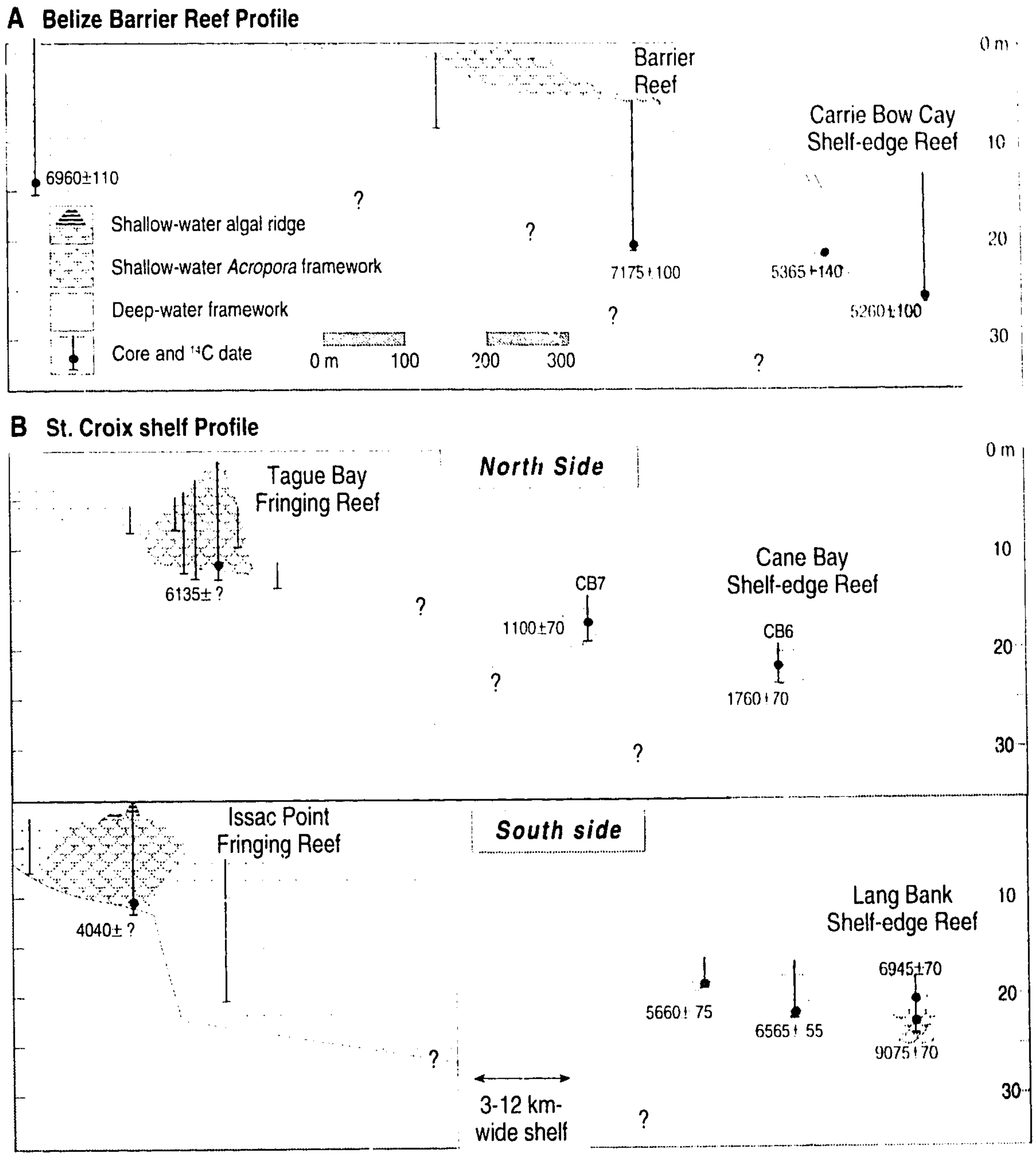

Figure 1.2. Summary of the findings of underwater drilling and excavations on modern reefs from the Caribbean. (A) Belize Barrier Reef at Carrie Bow Cay (James and Ginsburg, 1979; Macintyre et al., 1981; Shinn et al., 1982) (B) St. Croix reefs along the north side of the island at Tague and Cane Bays (Burke et al., 1989; Hubbard, 1989). St. Croix reefs along the south side of the island at Issac Point and Lang Bank (Adey, 1975; Adey et al., 1977).

a foundation for reef development (Fig. 1-2). In shallow waters of the inner shelf, a narrow emergent reef is nucleated on the upper terrace or close to the break-in-slope between the two terraces (Adey, 1975; Burke et al., 1989). Although this emergent inner-shelf reef is dominated by the surf-adapted coral *Acropora palmata* on the surface (Adey, 1975; Burke et al., 1989) its internal anatomy is not as easily determined because the narrow-diameter cores preclude a confident assessment of coral orientation. Drilling laterally in positions along the crest has demonstrated that reef growth initiated at different times in different areas but had started as early as ~7 cal. ka ago at a depth of ~10 m (Burke et al., 1989).

Much less drilling has been done on the other type of shelf reef, which is developed on the shelf edge, close to or directly superimposed over the shelf break itself. This shelf-edge reef is developed around the entire perimeter of the island shelf with little interruption (Adey et al., 1977), but is completely submerged and has been described as "an actively accreting entity that exhibits all the criteria of card-carrying reefs" (Hubbard, 1989). It consists of an array of buttresses that rise from depths of ~50 to 20 m. Although reef morphology has not been studied in any detail, individual buttresses generally have gently sloping upper surfaces and near vertical or in some cases overhanging fronts (Hubbard, 1990). Limited drilling into these buttresses shows that they are composed of a framework of deep-water corals that have accreted a minimum of 7 m vertically and about the same horizontally in a seaward direction (Adey et al., 1977; Hubbard et al., 1986). In 2 out of the 3 areas investigated, this deep-water framework is underlain by a drowned *A. palmata* reef that started growing sometime before ~10.5 cal. ka but had drowned by ~7.5 cal. ka (Adey et al., 1977)—apparently only 500 years before the initiation of inner-shelf reefs.

These studies of modern reefs from Belize and St. Croix provide clear evidence that reef architecture and anatomy is growth induced and not underlain by topographic residuals. In the case of Belize, no foundation was encountered seaward of the reef crest and major reef features were constructional. Similarly, the reefs around St. Croix were also constructional, but those along inner parts of the shelf were founded on flat terraces of unknown origin.

Based on the evidence, therefore, it seems that topographic residuals do not control reef configuration and development in these two areas.

Growth versus karst

Some modern reefs are clearly capable of producing their own configurations. It seems, therefore, that an important component of Purdy's (1974) karst theory has been falsified: namely that underlying karst residuals control reef configuration and architecture. The role of karst in reef development clearly has to be redefined or abandoned.

Recently, Purdy and Bertram (1993) presented a new hypothesis for the development of atolls which accepts that present day reef configurations are growth induced but maintains that initiation and overall configuration is still controlled by karst residuals. Specifically, they suggest that prior to Plio-Pleistocene glaciations the reef configurations were growth induced and quickly accreted to sea level. Slowly changing sea levels and low subsidence rates favored overproduction of sediment creating sedimentary systems dominated by lateral progradation. This produced flat-topped carbonate platforms that tended to merge, eventually producing carbonate megabanks like the present Bahama platform (e.g., Eberli and Ginsburg, 1989). With the onset of Plio-Pleistocene glaciations these flat-topped platforms were repeatedly exposed and flooded as ice sheets waxed and waned. During eustatic sea level falls, platforms were affected by karst processes and developed the familiar peripheral rims and interior solution basins (Purdy, 1974). Upon subsequent flooding the rims were preferentially colonized by reef growth. During the initial rapid stages of the rise, reef accretion was vertical and therefore accentuated the relief of the underlying karst rim, but later became dominated by lateral accretion as the rise slowed and growth processes modified the original configuration of the rim. Repeated episodes of rise and fall perpetuated the karst rim configuration producing a stacked sequence of karst-modified reef units over an ancestral karst rim. Purdy and Bertram (1993) concluded, therefore, that atoll and barrier reefs are ephemeral Pleistocene features that are the result of arrested infilling and karsting during sea-level fall and vertical accretion during the following rise a conclusion also made by Daly (1915) for different reasons.

In this context, Purdy and Bertram (1993) claimed that the initiation of modern reefs on older reefs is not at odds with their hypothesis. Neither is the fact that modern reef architecture is controlled by growth processes, as this would develop naturally as reef systems switched from vertical to lateral progradation during late stages of the rise. The only fact that could negate the karst hypothesis, they claim, is the absence of an erosional core beneath the reef position. They suggest that the only alternative to the karst residual hypothesis is one of Pleistocene reef growth modified by karsting but this "...leaves unanswered the question of why reefs assume this pattern in the first place." (Purdy and Bertram, 1993, p. 47). They conclude that "The reef argument is an old one based solely on faith, and it provides no insight as to cause and effect." (Purdy and Bertram, 1993, p. 47).

Coral reef development—a 150 year problem

Purdy and Bertram (1993) have to a certain extent redefined the karst residual idea and in doing so have thrown down a gauntlet for reef workers to take up. Yet, as far as the credibility of karst in producing reef configurations is concerned, the evidence is still as circumstantial as ever. Furthermore, there are inconsistencies in Purdy and Bertram's (1993) argument that suggest there is more to the problem than just karst. For instance, they implied that reefs initiate on the edge of shelves and platforms for no other reason than the existence of karst rims produced during episodes of sea-level fall. Are we to believe that all shelf-edge reefs in the geological record owe their siting and configuration to residual karst rims? If reefs initiated on shelf margins during times when sea level was stable, then the role of karst in determining reef configuration might be seriously compromised. Yet, with few credible arguments from the proponents of reef growth, the karst residual theory remains the only attempt to explain reef configuration—something that is no doubt responsible for its wide and uncritical acceptance (Longman, 1981; Tucker and Wright, 1990; Jones and Desrochers, 1992).

Clearly, the coral reef problem is alive and kicking and outstanding problems regarding the development and configuration of modern coral reefs still abound. Reef workers have had good reasons for doubting the role of karst in reef configuration but have largely proceeded in the classic empirical tradition, collecting much needed field data but largely ignoring theoretical discussion and making few attempts to address the basic problems. Much of the data is equivocal because it has been collected from the inner parts of the reef tract (reef crest and flat); outer parts are inaccessible because of rough open seas and have come to be known as the *mare incognitum* in light of these difficulties (e.g., Smith and Harrison, 1977). Only in the Caribbean, where open water conditions are less severe, have reef tracts been studied in their entirety. Even then data has been restricted to upper-reef sections that only represent events during the last 8 ka or so. Results have also been conflicting. Some reefs, particularly those in lagoons, show initiation on topographic residuals, whereas others, particularly those developed on open shelves, show initiation on flat terraces of unknown origin. To add to the confusion, local sea-level curves (e.g., Neumann, 1971) have been used to extrapolate reef accretion patterns (Adey, 1978). And then, in a classic case of circular reasoning, these extrapolations have been used to identify late-stage responses to sea-level rise (Neumann and Macintyre, 1985; Davis and Montaggioni, 1985). Not surprisingly, the responses identified have been varied, even within the same reef (see Macintyre et al., 1981; Montaggioni, 1988).

This 'extrapolation' approach to identifying a reef's response to sea-level rise has, as Purdy and Bertram (1993) have so rightly stated, provided little insight into cause and effect. Processes controlling reef initiation and demise, as well as the basic controls on reef architecture and configuration are little understood. Consequently, several key questions regarding reef development remain unanswered:

1) Why are some modern reefs preferentially located along the edge of shelves and platforms whereas others develop in inner-shelf positions? In other words, what processes and agents control reef configuration?
2) Do modern reef configurations initially reflect a substrate control or are they growth induced? Were they influenced by the presence of karst residuals or was some other, as yet undefined, ecological or physiological process involved?

3) What was the exact response of reefs to Pleistocene sea level rise? Did reefs accrete vertically during early rapid stages of the rise and then accrete laterally during later stages, as Purdy and Bertram (1993) suggested? Or, did the locus of reef growth change as sea level rose?

4) What were the effects of Pleistocene sea level falls? Were the exposed reefs significantly modified by karst solution as Purdy (1974) suggested or did marine erosion during the subsequent rise obliterate these? If so, what modifications did these processes impart?

5) Are modern reef configurations a product of Pleistocene sea level changes as Purdy and Bertram (1993) and Daly (1915) have suggested? And if so, can they be used as analogues for ancient reefs?

Despite 150 years of study, we still have a very inadequate understanding of the controls on reef configurations and architecture and several long-standing problems remain. Yet, if we are to use reef records for addressing more pressing concerns of climate change, these problems must be addressed. Needless to say, some change of approach seems warranted. New work must concentrate on establishing cause and effect through the application of a process sedimentological approach. For carbonate workers, traditionally trained in biological backgrounds, this is unfamiliar ground, and yet clastic sedimentologists have embraced such principles for many years. Perhaps the time has come for carbonate workers to do the same.

Thesis Objectives

In light of the outstanding problems regarding the development of Holocene and Pleistocene reefs, the aim of this thesis is to establish the controls on modern reef configuration and development around the island of Grand Cayman. The thesis is written in a paper format and the objectives of individual chapters are outlined below:

Chapter 2 describes the general setting of Grand Cayman.

Chapter 3 establishes the response of reefs to rising sea level following the last deglaciation and uses this to elucidate the interaction between sea-level, ice-sheet, and climate change.

Chapter 4 determines the nature, origin, and genesis of the foundation beneath coral reefs around Grand Cayman.

Chapter 5 illustrates the architecture, anatomy, and configuration of the fringing reef and determines the processes and agents responsible for reef development.

Chapter 6 illustrates the architecture, anatomy, and configuration of the shelf-edge reef and determines the processes and agents responsible for reef development.

Chapter 7 summarizes the controls on modern reef architecture and configuration around Grand Cayman and discusses their importance for determining the overall development of modern reefs.

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2

Grand Cayman: Cultural and Physical Setting

Introduction

Grand Cayman is part of a small British dependency (including the islands of Little Cayman and Cayman Brac) that is located in the northwest Caribbean Sea between Jamaica and Cuba (Fig. 2.1). Covering a mere 197 km^2, the island is small, relatively flat and low-lying, with a maximum elevation of only 17 m above msl. Its low surface is mainly covered by dense vegetation, and just over half the island is subtropical wetland composed of tidal mangrove swamps; the remainder, on higher ground, is dry evergreen thicket and bushland (Proctor, 1984). These rather unremarkable terrestrial environs are surrounded by shallow marine waters that are renowned for their reef development. Although averaging less than a km in width, the insular shelf hosts two diverse reef systems. A fringing reef, which affords protection for several large shallow-water lagoons, parallels the south, east, and north shores. And on the edge of the shelf around the entire perimeter of the island is a little known submerged reef that rivals the shallow reef in both size and diversity.

These impressive reefs are the topic of this thesis, and the background information presented below is therefore largely slanted towards marine environments and conditions. Readers wanting more land-based details are referred to Proctor (1984) and Brunt and Davies (1994)

Cultural Aspects

The island possesses a unique and largely overlooked maritime heritage. Until 1946, when sporadic air services started mass tourism, Grand Cayman was principally recognized for its abundance of sea turtles---a resource that not only helped open the New World, but also gave rise to a unique culture of seafarers whose nautical skills became world renowned (Smith, 1981).

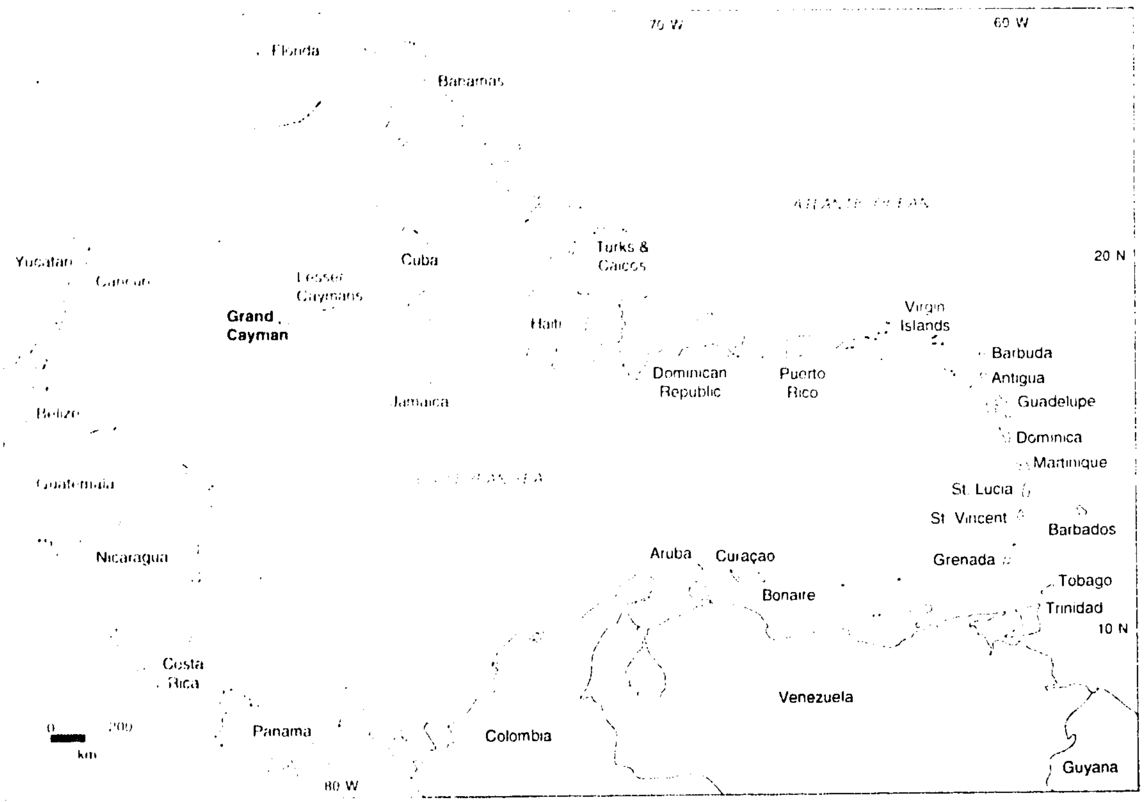

Figure 2.1. Grand Cayman, location and oceanographic setting.

The island was discovered by Christopher Columbus, on Wednesday, May 10th, 1503, and together with the lesser Caymans, initially provided way stations for passing mariners. The leeward west shore of the island, with its distinctive long beach of white coral sand, provided safe anchorage for these early visitors and an opportunity to restock provisions of food and water. It was also ideal nesting habitat for several species of sea turtle (including Green, Hawksbill, Ridley, Loggerhead, and Leatherback turtles). In fact these creatures were so numerous that Columbus gave the islands the Spanish name for turtles—Las Tortugas.

As word spread about these plentiful resources, the Cayman Islands became a strategic staging post for conflagrations between nations seeking territories in the New World. By the mid 1600's, French colonists were spending extended periods of time on the islands during turtling season, and unsubstantiated reports suggest that they, along with shipwrecked mariners, had established the first settlements by 1661 (Giglioli, 1976; Smith, 1981). Among these early colonists, however, it was the British who finally realized the strategic importance of the Cayman turtle fishery. Following their invasion of Spanish-held Jamaica in 1662, and the exhaustion of local food supply, they dispatched ships to the island in order to take charge of this resource. This action prompted immediate reprisals from the Spanish and subsequently led to unrestrained warfare. Hostilities were curtailed only by the ratification of the Treaty of Madrid in 1670, which proclaimed peace between the colonies of the two countries. In addition, and importantly for the Cayman Islands, the treaty also specified that the King of England was to "...have, hold, keep and enjoy for ever, with plenary right of sovereignty....all those lands, islands, colonies and places whatsoever situated in the West Indies, or in any part of America, which he and his subjects at present hold." Thus, Jamaica and the Cayman Islands became legal British territory (Smith, 1981).

In 1734 land grants were issued for permanent settlement of Grand Cayman, and by the later half of the century a distinct culture of "160 white men, women and children" had established themselves on the island (Smith, 1981). The primary occupation, of course, was turtling, and the islanders monopolized the trade by supplying turtle products directly to Jamaica and visiting mariners in exchange for manufactured goods. Unfortunately the use of nets to take turtles of all sizes led to a rapid depletion of stocks, and the islanders were forced to sea and distant shores to catch turtle. Here then was the start of the maritime traditions that were to be the mainstay of Caymanians for the next 170 years (Smith, 1981).

This was the basic situation until the tourism boom began in the late 1950's and early 60's (Giglioli, 1976). In a mere 20 years, this small underpopulated island of turtling mariners transformed into a sophisticated mecca for tourism and finance with a population of just under 32,000, and a GDP of ~US$670 million (CIA, 1994). Such unbridled growth has presented its problems. The rapid construction of new hotels and luxury villas has largely taken place within the return period of major hurricanes. In many areas this construction is located in areas that are severely impacted by hurricane charged seas. Needless to say, 'when the big one hits' old lessons will have to be relearned.

With this economic and population boom has also come the more familiar problems of environmental stress, degradation, and pollution. In some cases, such as the effects of giant cruise liners on the coral reef, this impact has been devastating and irreversible (see Smith, 1988). Although such problems have been sidelined by many other nations, they pose a serious threat to economies that are so heavily dependent on their fragile natural resources. For Grand Cayman, and her sister islands, it seems that the key challenge for the future will be to develop an economy that is sustainable. Maintaining the health of the reef resource base is critical to such an economy for, if its degradation continues, 'boom' could very well turn into 'bust'.

Local Climate

With its small size and low-lying nature, Grand Cayman enjoys a subhumid, tropical, and orographically unmodified, oceanic climate that is dominated by moisture-laden air masses of the North-East Trade Wind System. Like many other Caribbean islands subject to this system, wind and rainfall patterns are distinctly seasonal (Burton, 1994). During the wet season from May to November, the island is subject to hot temperatures (averaging ~29°C), frequent showers (averaging 4-8 mm/day), high humidity, and easterly or southeasterly winds (averaging 4-5 m/sec). From December to April, during the dry season, temperatures fall slightly (averaging 25°C), showers are less frequent (< 3mm/day), and winds move round to the east and north east (averaging 5-6 m/sec). Cyclonic disturbances, which provide a large proportion of the annual rainfall, are common during both seasons. Tropical storms and hurricanes track east to northeast during the wet season, and storms associated with continental cold fronts (Nor'westers) track west to northwest during the dry season. Although such disturbances inflict serious damage on tropical marine ecosystems (Woodley et al., 1981), their long-term affects on these systems are poorly known (but see Connell, 1978). When the historical frequency of severe storms affecting Grand Cayman is considered, however, it is clear that they are the primary physical agent impacting the island's marine environments (Fig. 2.2).

Storm-derived rainfall drains centripetally through the mangrove swamps towards the low western shores of the island (Giglioli, 1976). Run-off into the ocean is slight and mainly effected by percolation through beach ridges or solution-channel networks in the island's bedrock during periods of lower monthly tides (Giglioli, 1976; Ng, 1990). This process filters out suspended solids and only soluble compounds (mostly humic acids) are discharged into the surrounding ocean (Giglioli, 1976).

Marine Hydrology

The lack of surface run-off and suspended solids is largely responsible for the renowned clarity of the sea surrounding the island where horizontal visabilities up to 60 m have been reported (Giglioli, 1976). These clear surface waters are also pleasantly warm and closely reflect seasonal variations in air temperature, ranging from 26°C in the dry season to 29°C in the wet season. The limited amount of fresh water discharged into the ocean means that general salinity conditions in enclosed seawater bodies are normally marine (Kalbfleisch, 1995), although point-sourced seepages of freshwater have been observed in some lagoons (particularly North Sound) and in deeper water across the shelf (pers. obs.).

Tides

Grand Cayman's tidal regime is microtidal, consisting of mixed, dominantly semi-diurnal tides with an average amplitude of 26 cm and a mere 1 m range between 10 year maximum and minimum values (Burton, 1994). This limited amplitude means that tidal currents are generally weak and are easily modified by more dominant wind-generated currents. In spite of such weakness, however, they still play an important role in amplifying and suppressing other currents (Roberts et al., 1975; Darbyshire et al., 1976) particularly those generated by wave-overtopping in lagoons and westward drift over the open shelf.

Waves

Being flanked by the 7000-m Cayman Trench a short distance to the south and the 4500 m deep Yucatan Basin to the north, Grand Cayman is surrounded by exceptionally deep

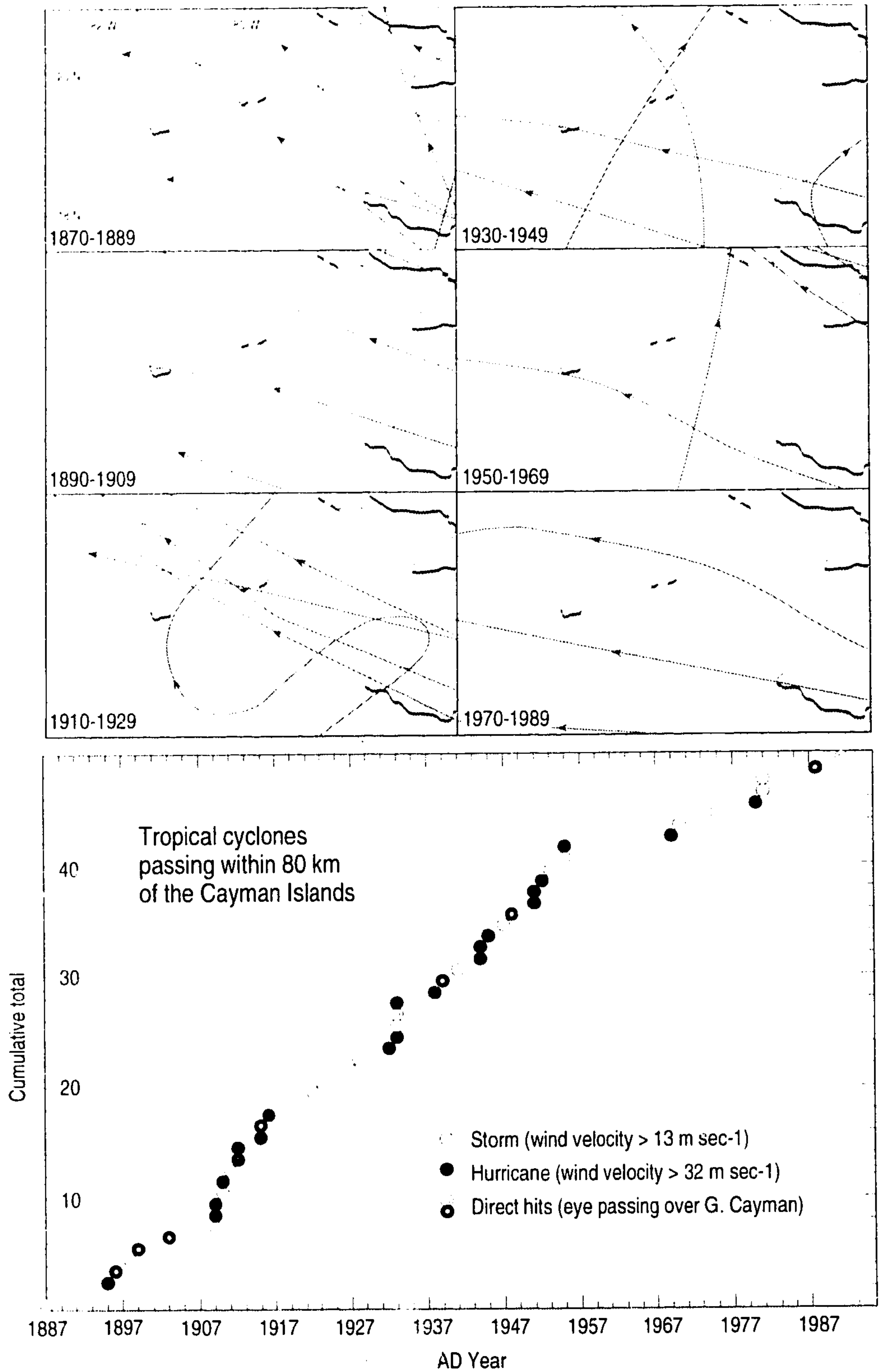

Figure 2.2. Historical frequency (lower) and tracks (upper) of tropical cyclones passing within 80 km of the Cayman Islands. (Frequency data from Clark, 1988; tracks from Woodley, 1992).

waters. This, together with narrow shelves and lack of coastal elevation makes the island vulnerable to any increase in wave activity. Fortunately, being sheltered from high-latitude storm swells by islands of the Greater Antilles chain, the wave field affecting Grand Cayman is not severe and is the combined product of the Northeast Trades and swells generated in the southern Caribbean. Annual mean wave-power values calculated for different sections of the coast (Roberts, 1974) show that east and southeast-facing margins receive the highest and most enduring wave energy (~3 x 10^9 ergs/sec), north and northeast-facing margins receive large to moderate energies (~0.7 x 10^9 ergs/sec), and west-facing margins receive the least energy (~0.04 x 10^9 ergs/sec).

Currents

Due to its small size, large-scale oceanic currents play an important role in the dynamics of water movement around Grand Cayman. General surface circulation in the Caribbean Sea is dominated by the western boundary current of the North Atlantic subtropical gyre. The southern sector of the gyre, composed of North Equatorial Current and the Guyana Current, flows west and enters the Caribbean Sea largely through passages between the islands of the Lesser Antilles (Fig. 2.1). In its early stages the current—from here known as the Caribbean Current—is relatively slow (<0.2 m s^{-1}) and diffuse but gradually intensifies (~0.5 m s^{-1}) as it moves westward through the southern Caribbean (Kinder, 1983). From there the current coalesces and turns northward over the Nicaraguan Rise producing intensified velocities (~2 m s^{-1}), especially across western parts of the rise (Triffleman, 1992). Downstream, the current initially slows, but as it is funneled towards the Yucatan Channel, it intensifies once more reaching peak velocities (~2 m s^{-1}) as it passes through the channel neck into the Gulf of Mexico (Wust, 1964).

Although Grand Cayman lies in a zone bordering this faster flowing upstream section of the Caribbean Current, the island is affected predominantly by a west flowing tributary of the current that enters the Caribbean through the Windward Passage between Cuba and

Hispaniola. Although less is known about the hydrodynamics of this current in its upstream section, by the time it reaches Grand Cayman it is a moderately strong unidirectional flow that averages 30 cm s^{-1} and can be detected down to ~300 m (Darbyshire et al., 1976). Detailed characteristics of this current and its interaction with the island have been studied using fixed, continuously-recording current meters deployed on the shallow shelf flanking the southwest coast (Roberts et al., 1975). These measurements show that the current is a tidally-enhanced drift coherent with the Trade winds. It shows a distinct periodicity corresponding to the tidal cycle, reaching 50 cm s^{-1} in midcycle and slackening thereafter, and shows a strong but lagged response (~36 hours) to increased wind stress. Tidally-induced current reversing is generally suppressed by the westerly drift component except during enhanced biweekly lunar tides. As the current passes around the island it is deflected up onto the shelf and suffers a 70 % reduction in velocity (< 8 cm s^{-1}) by the time it reaches inner parts of the reef dominated shelf. Using these data Roberts et al. (1975) divided the open shelf into two hydrographic zones; waters of the outer shelf are dominated by the tidally enhanced drift, whereas shallow inner-shelf water are dominated by wave-induced currents (Fig 2.3)

Storm Surges

Few major storm surges have been measured on Grand Cayman due to the practice of removing tide gauges just prior to storms to prevent their destruction. Hurricane Gretta (1978), tracking some distance to the south of the island, raised water levels by ~77 cm, whereas Hurricane Allen (1980) produced a surge of ~71 cm when it passed 90 km north of the island. More recently, Hurricane Gilbert (1988), with sustained wind speeds of 31 m s^{-1} and gusts of 60-70 m s^{-1}, produced a 150 cm surge and was associated with waves of up to ~3.5 m which breached raised shorelines flooding areas up to 3 km inland. Such large surges are "extremely rare" events because the island lacks wide shelves where storms can pile water (Clark, 1988). Although less reliable, eyewitness accounts of sea states during more

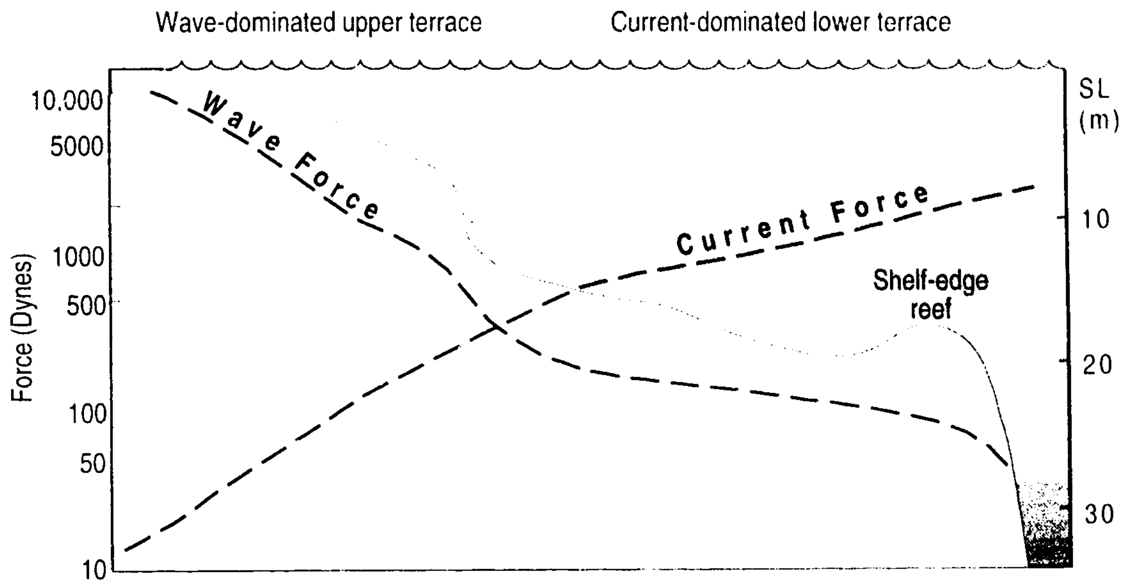

Figure 2.3. Distribution of wave and current force over the open shelf around Grand Cayman (modified after Roberts et al., 1975).

'memorable' hurricanes are useful. For example, the following is an account of a hurricane that passed 40 km to the north of the island in 1910:

> ...most severe damage occurred at East End and the South Side where the sea rose up to 15 feet washing away roads and depositing boulder ridges. Red Bay and Spotts roads were washed away and sea encroached 50 yards onshore bringing rocks and boulders and depositing them as a breakwater. At Newlands a school of Jack and Kingfish were washed inland ~500 m. (Hirst, 1910)

The reference to the sea rising 15 feet (~4.5 m) in this case was probably not describing surge since this would have inundated large areas of the island. The 15 feet in this context more likely refers to wave height. The effects of even larger magnitude waves on Grand Cayman's coastal areas has recently been documented by Jones and Hunter (1992). They described clusters of boulders along cliffed sections of the unprotected south coast. These boulders, some weighing up to 40 tonnes, were lifted 18 m from the base of the cliff and transported inland up to 60 m probably during hurricanes. Although their height was not estimated, it is clear that the waves responsible were of considerable magnitude (~18 m).

Marine Environments

The narrow shelf around Grand Cayman consists of two seaward sloping terraces whose age and origin are unknown (Figs. 2.3, 2.4). The upper terrace, which reportedly slopes from 3-10 m (Rigby and Roberts, 1976), is associated with two environments: lagoons and a fringing-reef complex. Both zones are easily accessible and have consequently received significant amounts of attention; various biological communities have been described in detail by Roberts (1971), Rigby and Roberts (1976), Swain and Hull (1976), Raymont et al., (1976), Logan (1981) and more recently by Hunter (1994) and Kalbfleisch (1995). The lagoons are characteristically shallow (2-3 m), narrow (most < 0.5 km), well flushed, and consist of a mosaic of marine grass beds, small patch reefs, bare sand, and rock. The two largest lagoons, North Sound and East Sound, have similar environments but are generally wider and deeper than the other lagoons (as much as 8 km wide and 8 m deep). Lagoonal sediments, composed of *Halimeda*/foraminifera grainstones to packstones, form only a thin veneer (< 1 m) over a flat, pitted bedrock surface. All lagoons are protected by a fringing-reef complex. This consists of a back-reef apron of coral/*Halimeda* grainstones and rudstones, and a reef crest and reef front dominated by the surf-resistant coral *Acropora palmata*. The thickness and anatomy of this unit has not been described, although Kalbfleisch (1995) found the sediment apron was at least 2 m thick in Frank Sound.

The lower terrace, which slopes from ~15-25 m (Rigby and Roberts, 1976), hosts the shelf-edge-reef. In contrast to the shallower environs, the shelf-edge-reef complex is less accessible and has received only limited attention. Most of this has been the identification of biotic communities (Hunter, 1994; Ghiold et al., 1994): details of reef architecture and anatomy have been overlooked. Generally, the shelf-edge-reef complex consists of an array of coral buttresses located along the outer half of the shelf. The buttresses are completely submerged, rising into waters no shallower than ~15 m below msl, and extend from the shelf edge back on to the shelf. They host a diverse biota including at least 30 coral species—significantly more diverse than any of the island's other reefs (Hunter, 1994).

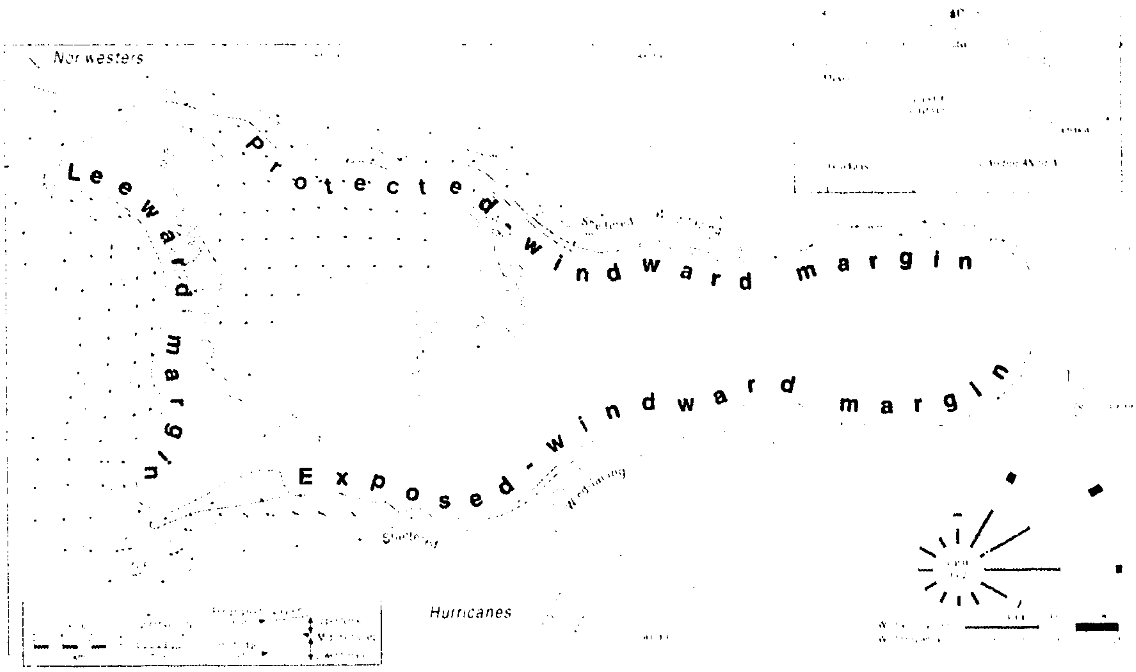

Figure 2.4. Configuration of reef types across Grand Cayman's insular shelf. Winds, currents, storm directions, and bathymetry also shown (modified from Blanchon and Jones, 1995).

Beyond the shelf-edge reef, is a 120-150 m precipice that forms the uppermost part of the island slope. This escarpment is least studied of all and has only been de[illegible] west side of the island (Messing and Platt, 1987).

Although the reefs of Grand Cayman are better developed than most, their arrangement and configuration over the shelf is similar to that of many other islands in the Caribbean-Atlantic reef province (cf. Goreau, 1959; Goreau and Goreau 1973; Rutzler and Macintyre, 1982).

Marine Geology

The island is situated in a complex tectonic zone that separates the North American Plate from the Caribbean Plate. This zone, which extends from Central America to the Virgin Islands, is a complex east-west trending Mesozoic orogenic belt, that is dissected by a major system of Cenozoic strike-slip transform faults (Lewis et al., 1990). These faults define an extensive rhomb-shaped graben known as the Cayman Trough, which has undergone ~1000 km of lateral extension over the last 50 Ma (Pindell and Barrett, 1990). Grand Cayman sits on the northern flank of this trough, along a narrow, east-west trending horst complex called the Cayman ridge.

Although the northern Caribbean area has undergone extensive tectonic activity, the Cayman Ridge has been stable since the middle Miocene when regional subsidence came to an end (Lewis et al., 1990). This cessation of activity coincided with the transfer of Caribbean-Plate movement to the northern transform fault in the Cayman Trough (Lewis et al., 1990). Activity along this fault continues today but uplift and subsidence are restricted to areas where fault bends or kinks produce transtensional or transpressional stresses (Mann et al., 1990). The proximity of Grand Cayman to planar faults has isolated the island from tectonic uplift or subsidence for the last 5 Ma, in spite of the fact that these faults have accommodated considerable lateral strike-slip movement (Ladd et al. 1990; Pindell and Barrett 1990).

Grand Cayman itself is the summit of a fault-block that rises from the Cayman Ridge. The ridge, which runs west from the Sierra Maestra of Cuba to Belize, consists of a series of small perched basins and uplifted blocks that rise up from the summit of the ridge at ~1500 m below msl. Although their stratigraphy is poorly resolved, dredging has shown the blocks to consist mainly of late Cretaceous to Pleistocene carbonate deposits (Holcombe et al., 1990). Two major textural types have been recovered; deep-water pelagic lime-mudstones, and boundstones containing shallow-water corals and other biota (Perfit and Heezen, 1978; Emery, 1980). The shallow boundstones are texturally and diagenetically akin to deposits that outcrop on Grand Cayman. The core of the island—to at least 400 m (Brunt et al., 1973) is composed of the Tertiary Bluff Group which consists of two formations: the Miocene Cayman Formation, a coralliferous dolostone, and the Pliocene Pedro Castle Formation, a mixed limestone-dolostone unit dominated by a fauna of free-living coral (Jones, 1994). A third formation, the Brac Formation, has also been identified from Cayman Brac, but has not yet been found on Grand Cayman (Jones, 1994). Onlapping this Tertiary core is the Pleistocene Ironshore Formation—a spectrum of reefal and lagoonal limestones distinguished by their essentially modern, shallow-water fauna (Matley, 1926; Brunt et al., 1973; Jones and Hunter, 1990). Areally, the Ironshore blankets much of the western half of the Island but is restricted to peripheral coastal areas in the east. Offshore drilling, however, has shown it to crop out across much of the inner shelf to a depth of at least 18.5 m below msl (Blanchon and Jones, 1995). Although the Formation was previously thought to represent deposition during the last interglacial (Jones, 1994) recent drilling and U-series dating (Vézina, pers. com.) has proved it to be a succession of units that accumulated during successive Pleistocene interglacials: this supports previous assertions that the island was stable for much of the Pleistocene (Emery, 1981; Jones and Hunter, 1990; Blanchon and Jones, 1995).

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3

Reef Drowning During the Last Deglaciation: Evidence for Catastrophic Sea-Level Rise and Ice-Sheet Collapse

Introduction

New evidence from Greenland ice-cores (Alley et al., 1993; Dansgaard et al., 1993; Taylor et al., 1993) and deep North Atlantic sediment cores (Bond et al., 1992; Lehman and Keigwin, 1992) demonstrates that the last glacial to interglacial transition involved sudden—and as yet unexplained—reorganization in ice-sheet, ocean, and atmosphere systems. The Laurentide Ice Sheet, for example, experienced two sudden collapse events during deglaciation, releasing huge volumes of ice into the North Atlantic and blanketing a broad swath of the sea bed with ice-rafted sediment (Bond et al., 1992, 1993). Atmospheric circulation also changed abruptly, switching between glacial and interglacial conditions in less than a decade (Alley et al., 1993; Taylor et al., 1993). These events were accompanied by equally dramatic (~40 yr) changes in North Atlantic circulation as the strongly stratified glacial ocean was disrupted by initiation of thermohaline circulation (Lehman and Keigwin, 1992).

Besides investigating such dramatic changes, there is a need to reassess the cause of the Younger Dryas episode—a brief and possibly global return to glacial-type conditions from 12.9 to 11.7 calendar ka (Taylor et al., 1993). Although this episode was attributed to changes in the North Atlantic salt budget (Broecker et al., 1990), the evidence from Greenland shows that it started and ended far too rapidly and was too widespread for ocean forcing to have been the sole cause (Denton and Hendy, 1994). Any theory explaining late-glacial climate must account for the abruptness of these changes. This rules out mechanisms with slow response times such as insolation, atmospheric CO_2 content and whole-ocean salt budgets, and points to a triggering mechanism.

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To address these problems and identify the deglacial triggering mechanism, we constrain the rate, magnitude, and timing of glacio-eustatic sea-level change from elevations and ages of drowned *Acropora palmata* reefs (hereafter referred to as *Acropora* reefs) in the Caribbean-Atlantic province. When these data are integrated with a coral-based sea-level curve, they show three catastrophic, metre-scale sea-level rises during deglaciation. By converting radiocarbon-dated marine and ice-sheet events to a sidereal chronology (Bard et al., 1993), we show that the timing of these catastrophic rises is coincident with ice-sheet collapse, ocean-atmosphere reorganization, and large-scale releases of meltwater.

Reef-Drowning Events

Suitability for radiometric dating (Edwards et al., 1987) and tendency to maintain themselves at sea level by rapid vertical accretion (Buddemeier and Smith, 1988), make reefs ideal for studying glacio-eustatic sea-level changes during the Quaternary. Reefs composed of the common Caribbean reef-crest coral *Acropora palmata* (Lamarck) are well suited for the task because (1) this is the only coral to form monospecific reef-framework in waters less than 5 m deep and, (2) it has a depth-restricted habitat range to ~10 m (Goreau, 1959; Gladfelter and Monahan, 1977), although in rare instances it has been reported as deep as 17 m (Goreau and Wells, 1967). This limited depth range means that *Acropora* reefs can track rising sea level, providing the rate of sea-level rise does not exceed the maximum reef-accretion rate of 14 mm/yr (Buddemeier and Smith, 1988). Consequently, rises that are below this threshold rate can be accurately determined by dating the elevation of *A. palmata* reef framework (Lighty et al., 1982).

What has not been previously recognized, however, is that sea-level rates that exceed the accretion threshold can be quantitatively constrained from framework changes in an *Acropora* reef as it drowns. Sea level rising faster than 14 mm/yr will displace *A. palmata* from its (monospecific) framework range (0-5 m) into its remaining habitat range (5-10 m), where a mixed framework with other corals develops (Goreau, 1959). During the final stage

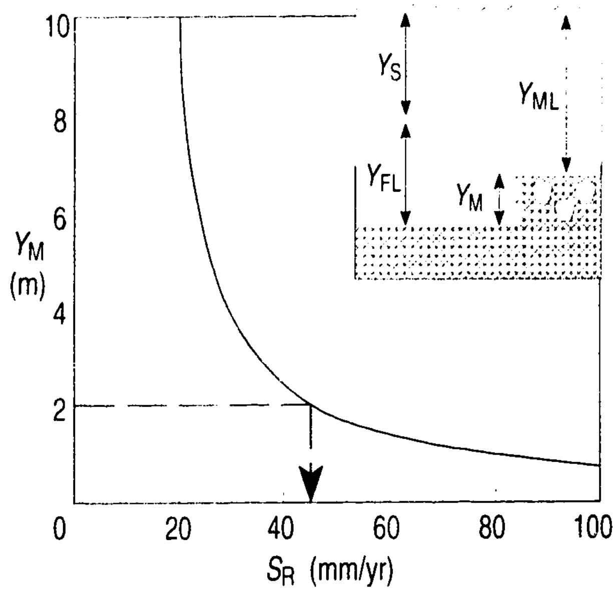

Figure 3.1. Relation between thickness of mixed-coral framework (Y_M) developed during reef drowning and rate of sea-level rise (S_R). This is given by $S_R = 1/t\ (Y_{ML} + Y_M - Y_{FL})$ derived from inset diagram, where t is time taken for accretion of Y_M during drowning, Y_{FL} is framework depth limit for *Acropora palmata*, Y_{ML} is depth limit for mixed *A. palmata*-other-coral framework, and Y_S is sea-level rise during accretion of Y_M. Note, t is obtained from Y_M assuming reef accretion rate of 13 mm/yr—the maximum accretion rate of *Acropora* reefs during last deglaciation (from Bard et al., 1990). Dashed arrow shows rate of sea-level rise required to form 2 m of mixed framework—a thickness that could be easily distinguished in core.

owning, the reef surface passes out of the *A. palmata* habitat range and the mixed nework is replaced by deeper water corals. Because the residence time of the reef surface in the habitat range is controlled by the rate of sea-level rise, there is an inverse relation between the thickness of mixed framework developed during drowning and the rate of sea-level rise (Fig. 3.1). Where mixed framework is >2 m thick, the rate of sea level rise must be >14 mm/yr—the maximum *Acropora* reef accretion rate—but <45 mm/yr . At higher rates, the residence time of the reef surface in the habitat range is insufficient for a significant thickness of mixed framework to develop. Hence, mixed frameworks <2 m thick indicate sea level rising at >45 mm/yr (Fig. 3.1).

To use this mixed-framework/rise-rate relation, the possibility of framework changes induced by autogenic processes—such as progradation—must be eliminated. Progradation can clearly be ruled out if reef-framework changes in the drowned *Acropora* reef are accompanied by reef back stepping—i.e., the establishment of *A. palmata* growth further upslope following drowning. Thus, we argue that abrupt framework changes accompanied by back stepping in *Acropora* reefs indicate a sea-level rise of >45 mm/yr.

Using this approach, we have constrained rapid rates of sea-level rise during deglaciation from drowned *Acropora* reefs (Fig. 3.2). Depths of the drowned reefs are grouped at ~80,

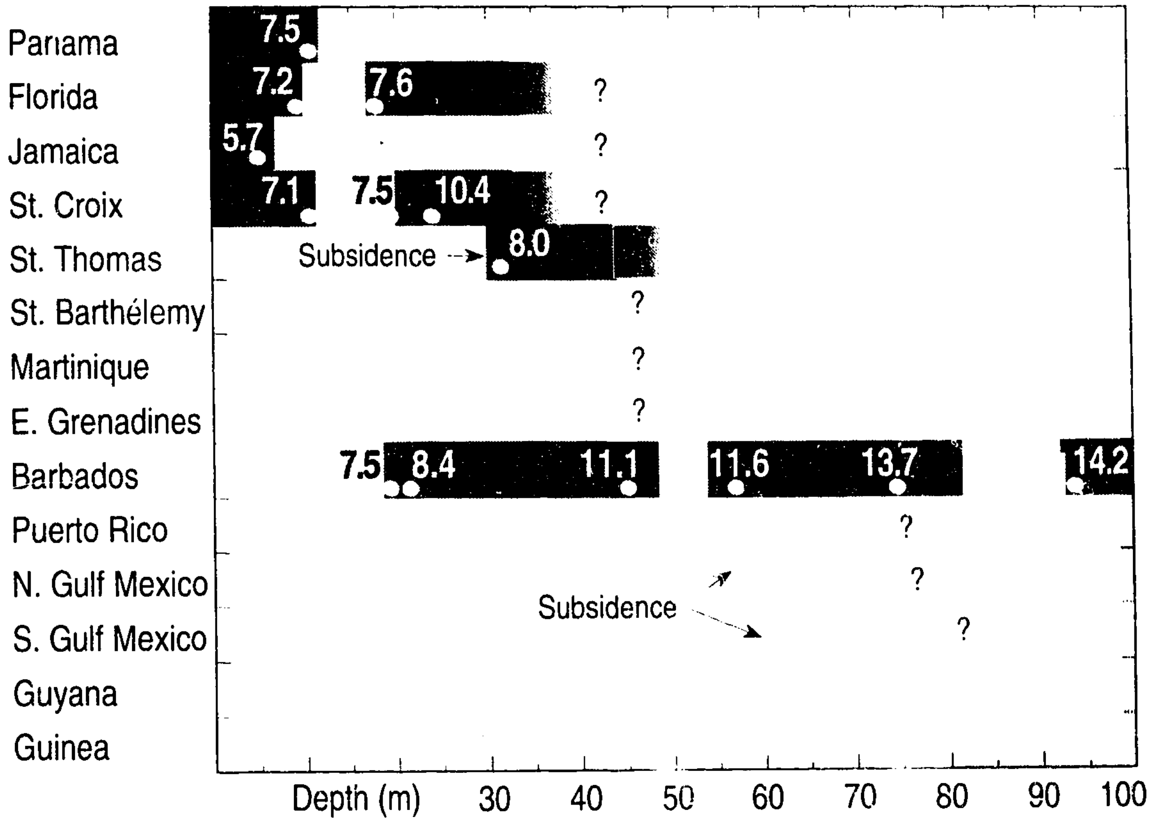

Figure 3.2. Depth below sea level, age (calendar ka), and framework character of drowned reefs in Caribbean-Atlantic province (modified after Blanchon and Jones, 1995). Dark shading indicates *Acropora palmata* framework; light shading indicates other-coral and unknown framework. Ages are corrected ^{14}C or U/Th dates. Barbados reef positions corrected for tectonic uplift of 0.34 m/ka (Fairbanks 1989) based on U/Th chronology (Bard et al., 1990). Note, reestablishment following drowning of 15 m group was not synchronous because (1) there was local lack of substrate (e.g. rise did not fully drown local sea cliffs) or, (2) lowest date does not reflect true date of reestablishment. (Date on youngest St. Croix reef from Burke et al. 1989).

50, and 15 m below present sea level. Although fewer data exist for the 80 and 50 m groups, drilling on the Barbados shelf (Fairbanks, 1989) shows them to be composed of thick, back-stepping sequences of *A. palmata* framework overlain abruptly by 10-15 m of deeper water coral framework. Depths of the deep reefs on Barbados correspond to deep reefs in other areas (Fig. 3.2), pointing to a common drowning history. This is especially clear for the 15 m *Acropora* reef group, which drowned at ~7.6 ka and back stepped at least 5 m instantaneously (e.g., compare the drowning date of Florida reefs with establishment of Panamanian reefs in Fig. 3.2).

Distinct breaks between the *A. palmata* reef-framework (Fig. 3.2) demonstrate that these reefs drowned and back stepped to upslope positions three times during deglaciation. Not only did the rate of sea-level rise exceed accretion rates during these drowning events, but the magnitude of the rise was also sufficient to remove *A. palmata* from its 5 m framework range, thereby preventing reef recovery. Furthermore, the abrupt transition from monospecific *A. palmata* framework to other-coral framework documented in cores from Barbados and St. Croix (Fig. 3.2) indicates that sea-level rise events also displaced *A. palmata* from its habitat range before significant accretion of a mixed framework could occur, implying the rise rate was >45 mm/yr (Fig. 1). Such drowning events must have been truly catastrophic, involving—to our knowledge—the fastest rates of glacio-eustatic sea-level rise yet reported.

When these catastrophic rise events (or CREs) are integrated with the corrected Caribbean sea-level curve (Fig. 3.3), the stepped nature of sea-level rise during deglaciation becomes clear. The first two *Acropora* reef-drowning events confirm and further constrain previously identified rapid rises (Fairbanks, 1989), and the third identifies a new rise event. Each step in the curve starts with a CRE with a rise-rate of >45 mm/yr, and concludes with a slower rise rate of <15 mm/yr. Using established coral dates (Bard et al., 1990; Fairbanks, 1989; Lighty et al., 1982) and gaps between *A. palmata* framework, the timing and magnitude of each CRE can be constrained (Fig. 3.3): CRE-1 started at 14.2 (±0.1) ka and had a magnitude of 13.5 (±2.5) m; CRE-2 started at 11.5 (±0.1) ka and had a magnitude of 7.5 (±2.5) m; and CRE-3 started at 7.6 (±0.1) ka and had a magnitude of 6.5 (±2.5) m. The exact duration of the CREs is unknown but, given that the minimum rate of sea-level rise was >45 mm/yr, the duration of the 14.2 ka event must have been <290 (±50) yrs, the 11.5 ka event was <160 (±50) yrs, and the 7.6 ka event was <140 (±50) yrs.

Sea Level—Ice Sheet Link

Catastrophic steps in sea level recorded by drowned *Acropora* reefs demonstrate that oceans were inundated by massive volumes of meltwater and/or icebergs at least three times during deglaciation. Smoother step-like rises in the deglacial sea-level record have been identified, but were attributed to large increases in seasonal meltwater discharge (Fairbanks,

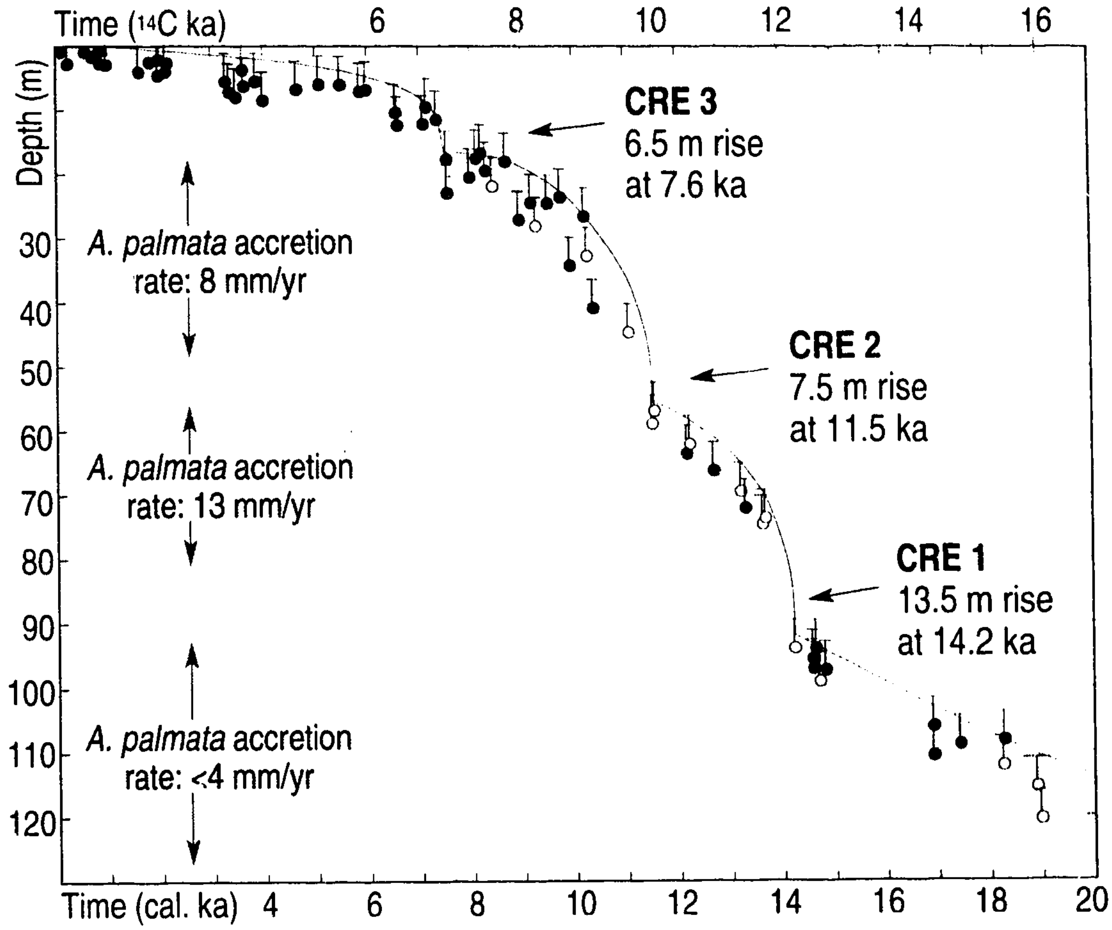

Figure 3.3. Caribbean deglacial sea-level curve showing positions of drowned *A. palmata* reef framework (light shading). Curve is extended after Lighty et al., (1982) and incorporates data from Bard et al., (1990), Fairbanks, (1990). Curve must lie on or above all data points because corals grow below sea level. Circles show positions of U/Th dated *A. palmata* (white) and corrected ^{14}C dated (black) *A. palmata*; error bars represent 5 m range of sea-level due to framework range of *A. palmata* and age error (1σ). Gaps between *A. palmata* framework enable magnitude of sea-level-rise events with rates >45 mm/yr to be quantified using 0.5 (2h+5) where h is height between successive frameworks and 5 is framework depth range for *A. palmata*.

1989). Although melting rates varied during deglaciation, it is unlikely that this alone could account for the magnitude of CREs. This assertion is supported by widely dispersed layers of ice-rafted detritus in cores of deep North Atlantic sediment (Heinrich, 1988; Bond et al., 1992). These Heinrich (H) layers, as they are called, record the massive discharge of icebergs into the North Atlantic resulting from the collapse of the Laurentide ice sheet (Bond et al., 1992), which, according to some estimates, may have taken place in <100 yrs (Broecker et al., 1992). Such rapid purges of ice into the North Atlantic would cause CREs that drowned fast-growing *Acropora* reefs in the Caribbean.

To link ice-sheet collapse and CREs we matched deglacial ocean-volume changes, recorded by the Caribbean sea-level curve, with patterns of climatic and oceanic change recorded in ice and deep-ocean cores (Fig. 3.4). This matching shows that episodes of ice-sheet collapse, marked by H-1 and H-0 layers in core V23-81 (Bond et al., 1993) correlate closely with CRE-1 and 2, and with reorganization of the ocean-atmosphere system. A survey of ^{14}C dates for H-1, however, shows that calendar ages range from 14.5 ka (Broecker et al., 1992) to 16.9 ka (Bond et al., 1992; Andrews et al. 1994). For an event that is considered to be almost instantaneous (Bond et al., 1992), such poor resolution suggests that dates are affected by bioturbation and this is confirmed by numerous date reversals in the best dated cores (e.g., DSDP 609 in Bond et al. 1992). Nevertheless, the youngest date in the H-1 age range correlates with CRE-1.

More reliable temporal evidence of ice sheet collapse has been identified from dated terrestrial and submerged diamictons on the north shore of the Hudson Strait (Miller and Kaufman 1990; Kaufman et al., 1993). These, together with ice-direction and source indicators, show that a major ice-stream from the Labrador Dome (the single largest centre of the Laurentide ice sheet) underwent large magnitude surge-and-retreat events consistent with collapse at ~14 and 11.5 ka—dates that closely match CRE-1 and 2.

Although ice-sheet collapse provides a compelling explanation for CRE-1 and 2, a link between CRE-3 at 7.6 ka and Northern Hemisphere ice-sheet instability is improbable because of the small volume of ice remaining at that time. Corrected ^{14}C dates on marine cores from the shelves adjacent to the Antarctic ice sheet, however, show that significant expansion of marine-ice extent took place between 7 and 8 ka (Herron and Anderson, 1990; Domack et al., 1991). Thus, Antarctic ice sheet instability could account for CRE 3.

Ice Sheet—Ocean-Atmosphere Link

The correlation between CREs and ocean-atmosphere reorganization (Fig. 3.4) implies that ice-sheet collapse had a significant impact on climate. This is best illustrated by covariant trends in ice-sheet and ocean records. Abrupt changes occurred in all records at ~14.5 ka

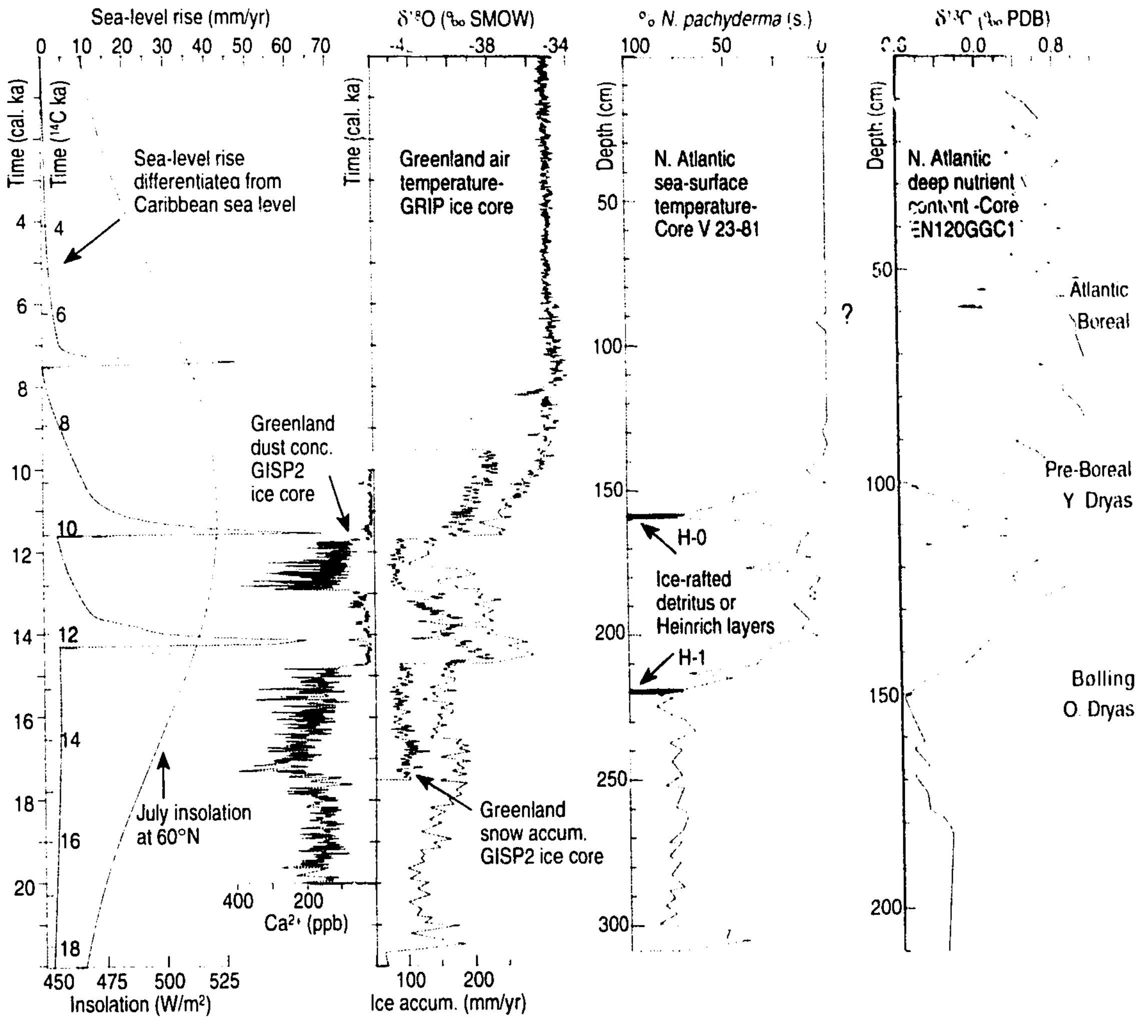

Figure 3.4. Rate of sea-level rise, differentiated from curve in Fig. 3, correlated with Northern-Hemisphere insolation and ocean-atmosphere changes during last deglaciation. Ice-core records of dust (Mayewski et al., 1993), snow (Alley et al., 1993), and temperature (Dansgaard et al., 1993) are dated by layer counting with an estimated accuracy of 3%. With these errors considered, CREs in sea-level record are synchronous with atmospheric-reorganization events recorded in ice cores. Note how Heinrich layers immediately precede oceanic reorganization events which correlate with CREs and atmospheric reorganizations.

(Fig. 3.4): glacial air masses over Greenland were suddenly (<10 yrs) replaced by warmer, moister, and less dusty conditions (Alley et al., 1993; Dansgaard et al., 1993; Mayewski et al., 1993), the sea-ice-covered North Atlantic was invaded by warm-water masses (Koç et al., 1993; Lehman and Keigwin, 1992), and ocean waters overturned in response to North Atlantic deep-water formation (Charles and Fairbanks, 1992) . Although temperature and snow-accumulation trends show gradual deterioration (Dansgaard et al., 1993; Alley et al., 1993), these conditions persisted until the next major reorganization at 2.9 ka—the onset of the Younger Dryas — when ocean-atmosphere circulation abruptly reverted to former glacial-type patterns (Lehman and Keigwin, 1992; Taylor et al., 1993). Finally, at ~11.5 ka, the Younger Dryas was terminated by an abrupt (<3 yr) reorganization in ocean-atmosphere circulation that heralded the start of present interglacial conditions (Alley et al., 1993; Koç et al., 1993).

On the basis of the coincident timing of CREs, we propose that this covariant pattern of dramatic ocean-atmosphere reorganization resulted from atmospheric threshold changes induced by a rapid decrease in the elevation (collapse) of the Laurentide ice sheet. In this view, the first collapse event—marked by CRE-1 at 14.2 ka—lowered the ice-sheet surface sufficiently to change tropospheric boundary conditions, weakening the ridge in the upper Westerlies over the ice sheet and causing the split polar-front jet stream to unite and rapidly retreat northward (COHMAP, 1988). This retreat facilitated the expansion of subtropical air masses and westerly winds, which in turn caused retreat of North Atlantic sea ice and allowed the warm western-boundary current to flow unrestricted into the northeast Atlantic, reactivating deep-water formation. In addition to abrupt circum-Atlantic warming, this rapid influx of subtropical water caused a dramatic increase in evaporation rates and delivered large amounts of moisture to Laurentide margins. Snow accumulation rates doubled (Alley et al., 1993), and over the next few thousand years, the ice sheet began to regain lost elevation, aided to some extent by glacio-isostatic recovery. By 12.9 ka it had regained sufficient height to split and divert the polar-front jet stream once more, causing an arm to shift southward. By restricting the subtropical air masses and westerly winds, this shift in the jet caused sea-ice formation, which effectively blocked the transport of warm water into the North Atlantic and plunged the global climate back into glacial mode during the Younger Dryas. The second ice-sheet collapse event—marked by CRE-2 at 11.5 ka—had the same effect as the first, switching the cold Younger Dryas ocean-atmosphere system back into warm interglacial mode, but this time melting induced by peak insolation—and perhaps reduced glacio-isostatic uplift—offset the increase in snow accumulation, and the Laurentide ice sheet was never able to recover.

A key point in this explanation is the expansion of subtropical air masses and the activation of thermohaline circulation by North Atlantic sea-ice retreat. Thermohaline activation was previously attributed to an over balanced North Atlantic salt budget related to a gradual increase in salinity during glacial conditions (Broecker et al., 1990; Lehman and Keigwin, 1992). Carbon and oxygen isotope records from the Greenland, Iceland, and Norwegian seas show that these sensitive areas of North Atlantic deep-water formation were dominated by low-salinity waters during glacial conditions, therefore throwing doubt on this mechanism (Veum et al., 1992). Our suggestion—that thermohaline activation resulted from air-mass expansion during atmospheric reorganization—finds support from evidence of rapid and synchronous climate change at ~14 and 11.5 ka in tropical African lake systems (Street-Perrott and Perrott, 1990) and Younger-Dryas advance of mountain glaciers in New Zealand (Denton and Hendy, 1994). Such parallel and synchronous trends from distant areas of the globe, suggest that atmospheric circulation flipped between glacial and interglacial modes, regulating deglaciation by switching the North Atlantic thermohaline heat pump on and off.

Triggering Ice-Sheet Collapse

Although ice-sheet collapse accounts for CREs and ocean-atmosphere reorganization, a fundamental problem remains: What initially triggered ice-sheet collapse? MacAyeal (1993) proposed that collapse was related to the onset of warm-based conditions in a previously cold-based ice sheet and, by estimating accumulation rates and geothermal flux, he calculated a collapse-recurrence interval of ~7 ka. Although this mechanism might explain collapse during the last glaciation, with 7 to 14 ka between events (Bond et al., 1993), it can not account for events <3 ka apart during deglaciation. Furthermore, if internal processes forced collapse and triggered deglaciation, it is difficult to explain why previous collapse events did not have the same effect. In short, deglaciation must have been triggered by an external mechanism that affected ice-sheet stability.

Atmospheric cooling has been proposed as a cause of ice-sheet instability (Bond et al., 1992), but no simple relation between climate change and ice-sheet response exists

TABLE 1.
MELTWATER MEGAFLOODS DURING DEGLACIATION

Megaflood (Ref.*)	Calendar age† (ka)	Discharge ($km^3 x 10^4$)	Eustatic rise (mm)
Livingstone Lake (1)	~14-15§	8.4	230
Baltic Ice Lake (2)	12.1	2.9	~100
Lake Agassiz (3)	11.4	2.3	70
Lake Agassiz (4)	11.5	2.9	~100
Lake Agassiz-Ojibway (5)	8.0	7.5-15	200-420

*1—Shaw (1989), 2—Bjorck and Digerfeldt (1986), 3 Broecker et al. (1989), 4—Smith and Fisher (1993), 5—Hillaire-Marcel et al. (1981).

†Calculated from ^{14}C dates (Bard et al., 1993).

§Products of the Livingstone Lake megaflood (see Rains et al., 1993) extend to position of last glacial maximum, limiting their timing to onset of deglaciation. Estimates of discharge and eustatic rise are from Shaw (1989) and Dawson (1992).

(Oerlemans, 1993). The only other viable mechanisms for destabilizing ice-sheets over a relatively short time are either delayed glacio-isostatic subsidence along the ice-sheet grounding line, or rapid sea-level rise (Hughes, 1987).

We propose that ice-sheet instability during the last deglaciation was triggered by the catastrophic release of meltwater megafloods from glacial and proglacial reservoirs. Such megafloods were released close to the times of ice-sheet collapse and CREs (Table 1). For instance, a large meltwater reservoir associated with the Laurentide ice sheet was catastrophically released sometime after the ice-sheet reached its maximum extent during early deglaciation (Shaw, 1989). The volume of water discharged produced regional-scale fields of drumlins, giant flutings, and extensive tracts of scoured bedrock (see Rains et al., 1993). Furthermore, faunally derived records of meltwater influx into the Gulf of Mexico—an expected megaflood outflow site—demonstrate an exceptionally large meltwater spike at ~14 ka (Broecker et al., 1989) that may have diluted a 1112-m water column in the Gulf (Aharon, 1992). Such large amounts of meltwater could potentially destabilize ice sheets grounded below sea level.

Although the link between gradual meltwater input, sea-level rise, and ice-sheet collapse has been suggested (Denton and Hughes, 1983) it is considered ineffective because of compensation by glacio-isostatic rebound (Lingle and Clark, 1985). The rapid influx of meltwater megafloods at times coincident with ice-sheet collapse and CREs is a more effective mechanism for triggering ice-sheet instability. It also provides the link between deglacial mechanisms and insolation—a forcing function acknowledged to be a major player in the glacial-interglacial cycle (Hays et al., 1976).

Deglacial Mechanisms

By identifying CREs from drowned *Acropora* reefs, we provide a critical piece of evidence that links insolation, large meltwater influxes, ice-sheet collapse, and ocean-atmosphere reorganization. From these links, we conclude that Northern-Hemisphere summer-insolation maxima forced deglaciation because greater land area in mid-latitudes allowed more extensive ice sheets. Stronger insolation over these latitudes generated large volumes of meltwater that, when catastrophically released, provided a trigger for subsequent interactions among ice sheets, oceans, and atmosphere. These interactions were dramatic. Collapse of one ice sheet affected them all (Denton et al., 1986), producing CREs that drowned reefs and other coastal features (Blanchon and Jones 1995). Collapse also abruptly switched atmospheric circulation patterns from glacial to interglacial modes, and consequently initiated changes in oceanic circulation that activated the Atlantic thermohaline heat pump — the mechanism ultimately responsible for Northern Hemisphere warming.

More importantly, we show that ice-sheet collapse and CREs were an integral part of deglaciation. Given that two large ice sheets over Greenland and Antarctica still exist, there is further potential for collapse. Indeed, if global atmospheric and surface-ocean warming continues at its present rate (Intergovernmental Panel on Climate Change, 1992), collapse of the West Antarctic ice sheet is a distinct possibility (Bindschadler, 1990). Consequently, despite the prediction of gradual sea-level rise by the IPCC (1992), the potential for future catastrophic sea-level rise also exists—especially now that catastrophic rises have been recognized from the recent past.

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4

Marine-Planation Terraces on the Shelf Around Grand Cayman: A Result of Stepped Holocene Sea-Level Rise.

Introduction

Marine terraces are important geomorphic elements of coastal areas and have been instrumental in reconstructing sea-level history during the last interglacial period (Mesolella et al., 1969; Bloom et al., 1974; Chappell, 1974; Dodge et al., 1983). Although studies have also confirmed that terraces are common elements of submarine topography (Emery, 1961; Newell, 1961; Stanley and Swift, 1968; Zankl and Schroeder, 1972; Schwartz and Lingbloom, 1973; Pratt and Dill, 1974; Focke, 1978; Twitchell et al., 1991), there is relatively little consensus regarding their genesis or significance.

On shallow shelves, the most commonly invoked mechanism of submarine terrace formation is intertidal marine planation during sea-level stillstands (Dietz and Menard, 1951; Emery, 1961; Dietz 1963). Although this erosional mechanism is widely accepted, a major problem exists in identifying the timing of terrace formation. Did terraces form during stillstands associated with the last glacial cycle, are they remnants of older cycles, or are they the result of incremental erosion during successive sea-level cycles? An alternative terrace-forming mechanism has been identified in tectonically uplifting areas, as a result of the interactions between fringing-reef accretion, episodic tectonic uplift, and eustatic sea-level oscillations (Mesolella et al., 1970). This constructional mechanism has been applied to stable shelf areas, where it has been suggested that submarine terraces form as the result of incremental reef growth during successive sea-level cycles (Hopley, 1982; Paulay and McEdward, 1990).

Constraining the timing and mechanism of shelf-terrace formation is of fundamental importance for understanding the developmental history of modern sedimentary shelf environments, and for documenting sea-level changes during the last glacial hemicycle (Blanchon

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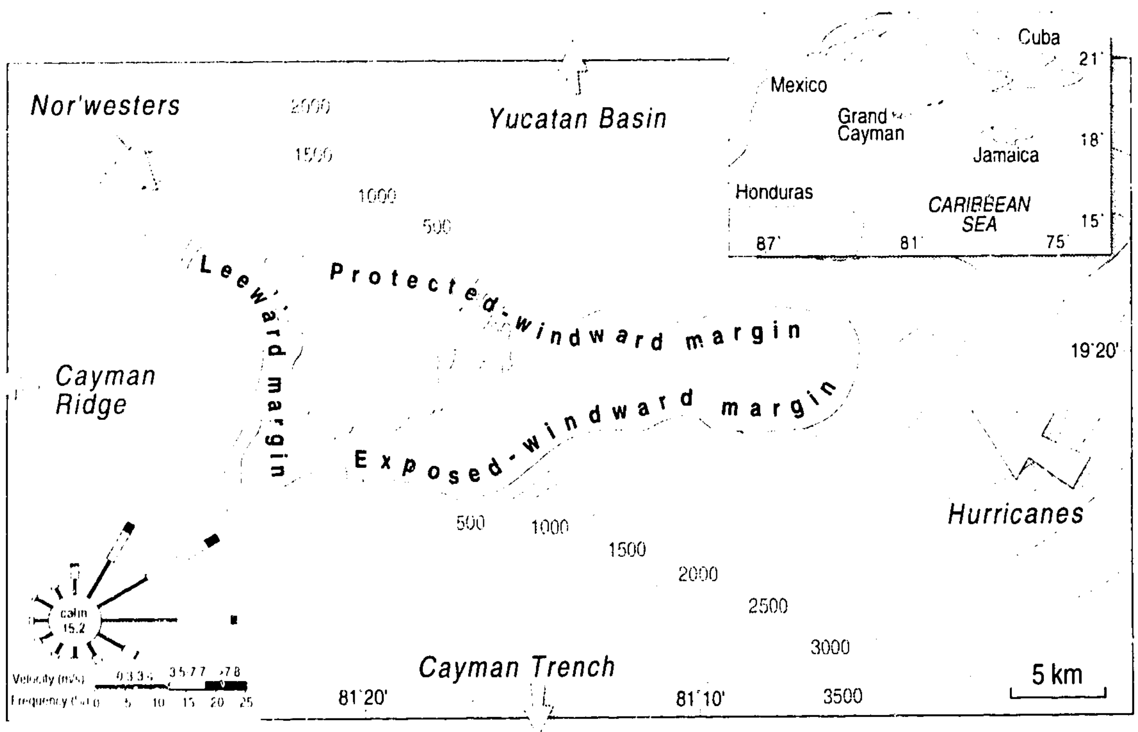

Figure 4.1. Grand Cayman. Location, bathymetry in metres (after LeBlanc, 1979), extent of shelf (dashed line), position of fringing reefs (shaded lines), and wind/storm directions (from Darbyshire et al., 1976). Shelf divided according to wave regime; exposed-windward margin (south and east sides), protected-windward margin (north side), and leeward margin (west side).

and Shaw, 1993a, 1993b). As part of a wider investigation on modern reef growth and sea-level change, this study was initiated to provide evidence on the mechanism and timing of terrace formation on the shelf around Grand Cayman (Fig. 4.1). By integrating data from submarine drill cores, scuba transects, depth-sounding profiles, and seismic profiles, we describe the bio-ge morphic zonation of shelf-terraces and link this zonation to the marine processes responsible for terrace formation. Then, supported by data from other shelf areas, we constrain the timing of the sea-level changes responsible. Finally, by providing insights into terrace formation around Grand Cayman, we propose a general model of terrace formation that relates different terrace types to different phases of the sea-level cycle.

Setting

Climate

Grand Cayman is a small, low-lying island located in the northern part of the Caribbean Sea (Fig. 4.1). Being sheltered from high-latitude storm swells by islands of the Greater

Antilles chain, its climate and wave regime are controlled by tropical, moisture-laden air masses of the northeast Trade Wind system. Under fairweather conditions, prevailing winds blow from a general easterly direction and commonly average 3-7 m s^{-1} (Darbyshire et al., 1976). As a result, only the west side of the island is a leeward margin, all others are windward (Fig. 4.1). Due to variations in fetch, however, the windward margin is subject to significant variations in wave energy. The south and east sides, which constitute the *exposed-windward margin*, have the greatest fetch and typically experience moderate sea states (1.25-2.5 m wave heights). In contrast the north side, which constitutes the *protected-windward margin*, has a limited fetch and experiences slight to calm sea states (< 1.25 m wave heights). Superimposed on these larger-scale variations in wave energy are changes induced by smaller-scale variations in coastal orientation. Thus, sections of the exposed- or protected-windward margins can be described as either *wind-facing* if they have an orientation with an easterly component, or *sheltered* if they have an orientation with a westerly component.

These fairweather wind-and-wave patterns are seasonally disrupted by atmospheric disturbances, including storms associated with cold fronts from the northwest, and hurricanes from the southeast (Fig. 4.1). Both types of disturbances regularly inflict serious damage on the island's marine and terrestrial ecosystems.

Bathymetry

The shelf around Grand Cayman is a gently sloping platform, typically less than a kilometre wide, that extends from coastal and reefal strandlines to a vertical shelf-edge escarpment. Like many open shelves in the Caribbean-Atlantic reef province, it consists of a regular zonation of reef-dominated carbonate depositional environments arranged in a concentric series of belts around the island (described in detail by Rigby and Roberts, 1976). The island shelf is bounded along its seaward margin by a spectacular escarpment that typically begins between 55-80 m, and extends vertically into waters 115-145 m deep (Messing, 1987). From there, the island slope extends into abyssal waters of the Cayman Trench, where depths exceed 7000 m. Similarly, along the north side, the slope extends into the Yucatan Basin where waters depths reach 4500 m. This shelf-scarp-slope configuration is common in other areas and has been reported from the Belize Barrier Reef (James and Ginsburg, 1979), Jamaica (Moore et al., 1976), and Tongue-of-the-Ocean, Bahamas (Grammer et al., 1989).

Tectonics

Grand Cayman is situated in a complex tectonic zone that separates the North American Plate from the Caribbean Plate. This zone, which extends from Central America to the Virgin Islands, consists of an east-west trending Mesozoic orogenic belt, that is dissected by a major system of Cenozoic strike-slip transform faults (Lewis et al., 1990). These faults define an extensive rhomb-shaped graben known as the Cayman Trough, which has undergone ~1000 km of lateral extension over the last 50 Ma (Pindell and Barrett, 1990). Grand Cayman sits on the northern flank of this trough, along a narrow east-west-trending horst called the Cayman Ridge.

Although the northern Caribbean area has undergone extensive tectonic activity, the Cayman Ridge has been stable since the middle Miocene when the last episode of regional subsidence came to an end (Lewis et al., 1990). The cessation of activity coincided with the transfer of Caribbean-Plate movement to the northern transform fault in the Cayman Trough (Lewis et al., 1990). Activity along this fault continues today but uplift and subsidence are restricted to areas where fault-bends or kinks produce transtensional or transpressional stresses (Mann et al., 1990). The proximity of Grand Cayman to planar faults has consequently isolated the island from tectonic uplift or subsidence for the last 5 Ma, in spite of the fact that these faults have accommodated considerable lateral strike-slip movement (Ladd et al., 1990; Pindell and Barrett, 1990). Island stability is also confirmed by morphological and sedimentological evidence (Emery, 1981; Jones and Hunter, 1990).

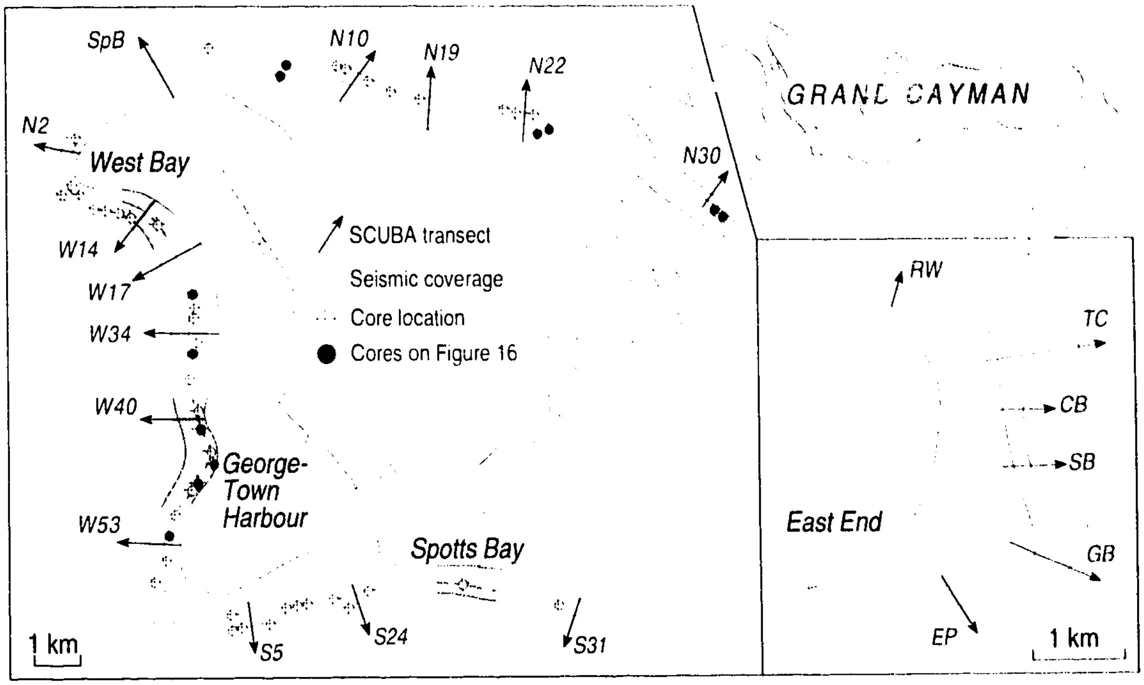

Figure 4.2. Location of scuba transects, areas of seismic coverage, and diamond-drill core locations on the shelf around Grand Cayman. (Dashed line shows position of mid-shelf scarp and shaded lines are fringing reefs).

METHODS

Direct observations and measurements of underwater features were made on 20 scuba transects from shore to the edge of the island shelf (Fig. 4.2). These observations, recorded using a combination of video and photographs, provided the basis for the interpretation of features from depth-sounding and seismic profiles.

Depth-sounding-profile grids were run across the shelf at regular intervals with shore-parallel lines spaced at 100 m intervals and shore-perpendicular lines spaced at 50 m intervals (Fig. 4.3). Line locations were made using shore-based markers and plotted using dead reckoning. Profiles were recorded on a print-out-style bottom profiler, powered from the onboard battery of a shallow-draft boat. For optimum sensitivity the transducer was externally mounted on the stern of the vessel, the paper speed adjusted to its fastest setting, and the boat speed reduced to a constant 4 knots. This set-up produced high-resolution profiles that recorded substrate depth and changes in substrate type.

Seismic data, shot by the Pelagos Corporation in April 1987, consisted of ~50 km of high-resolution seismic profiles taken perpendicular to shore at 50 to 60 m intervals (Fig.

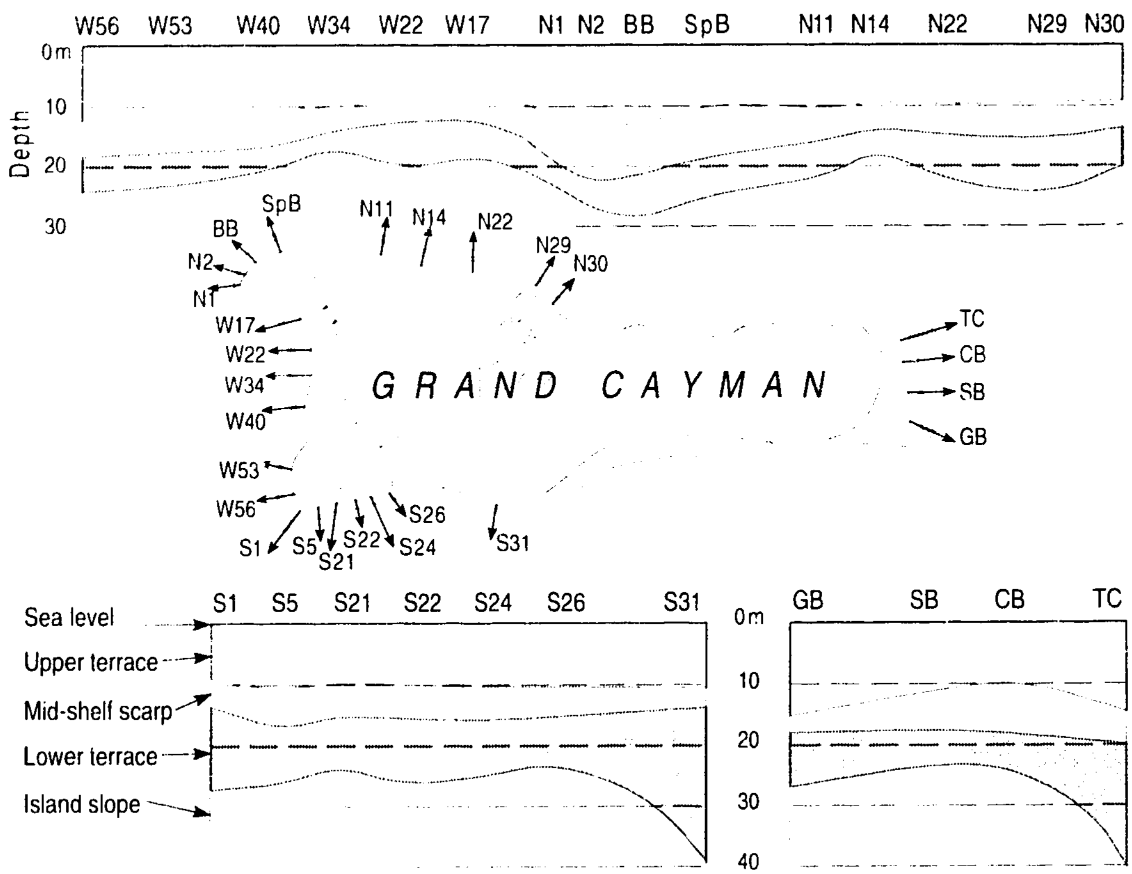

Figure 4.3. Locations of depth-sounding-profile grids (arrows) and positions of the upper terrace, mid-shelf scarp, and lower terrace at those locations. Position of slope breaks determined from profiles over non-reefal substrates. Note how the lower terrace is flatter on the protected-windward and leeward margins, but sloping on the exposed-windward margin. This is due to differential sediment buildup, with greater buildup along sheltered areas (see transects W34, W22, and W17) and less buildup in areas exposed to storms (see transects S31and TC).

4.2). Navigation and line locations were determined using a microwave-pulse positioning system that utilized a mobile interrogator and shore-based transponders. Subbottom-profiling surveys were carried out on the west and south sides of the island using a boomer-type (0.5-3 kHz at 175 joules), single-channel, analog acquisition system. The maximum depth penetration of this system below the sea floor was ~25 m, and the minimum resolution of individual features within the sediment was 1.5 m. All interpretations of sediment thickness from the seismic profiles were based on the velocity of sound in water-saturated, unconsolidated, sand-sized sediment, arbitrarily taken as 1750 m s^{-1} (average of empirical values from 1650-1850 m s^{-1}; Akal, 1972; Morton, 1975; Hamilton and Bachman, 1982).

To confirm interpretations made from seismic profiles, a stainless steel probe was used to determine the thickness of unconsolidated sediment along the scuba transects. The probe, which consisted of screw together sections of stainless-steel rod fitted with a 5/8 inch masonry bit, was driven into the sediment using an air-powered drill supplied from a scuba tank (see Jones et al., 1992 for specifications). With the addition of a masonry bit, the probe was able to drill through cobble/cemented layers that would have otherwise limited the depth of penetration. This method provided a minimum thickness of unconsolidated sediment and confirmed the accuracy of estimates from the seismic profiles.

Cores from hard substrates were obtained using a diver-operated hydraulic drilling system, similar to the one described by Macintyre (1975, 1978). The system, powered by a 700 cc/18 hp hydraulic motor unit (3600 rpm), was fitted with a 10 cm diameter, 1 m core barrel. Fifteen cores were obtained from the west, north, and south sides of the island in waters 2 to 25 m deep (Fig. 4.2). A further 40 cores, cut during a dive-site mooring-installation program, were donated by the Natural Resources Unit of the Cayman Islands Government (Fig. 4.2). In addition, a smaller pneumatic drill, sourced from an ordinary scuba tank (see Bonem and Pershouse, 1981), was used to sample vertical substrates. Cores were slabbed, logged and analyzed using conventional thin-section and SEM techniques.

To determine suitability of core samples for dating, the degree of aragonite to calcite inversion was analyzed using X-Ray Powder Diffraction (Davies and Hooper, 1962). Radiocarbon determinations, made by conventional gas-proportional counting and using the Libby half life (5568 yrs. BP), yielded an average analytical uncertainty (1 σ) of $\pm$1150 years for samples older than 17 ka, and $\pm$90 years for those younger than 1.7 ka. It should be noted, however, that all samples in the older group contained calcite and 'ages' are therefore severely rejuvenated (Olsson, 1974). Samples in the younger group were 100% aragonite but were not corrected for the oceanic reservoir effect (Bard, 1988) or secular variations in the atmospheric radiocarbon production (Bard et al., 1993).

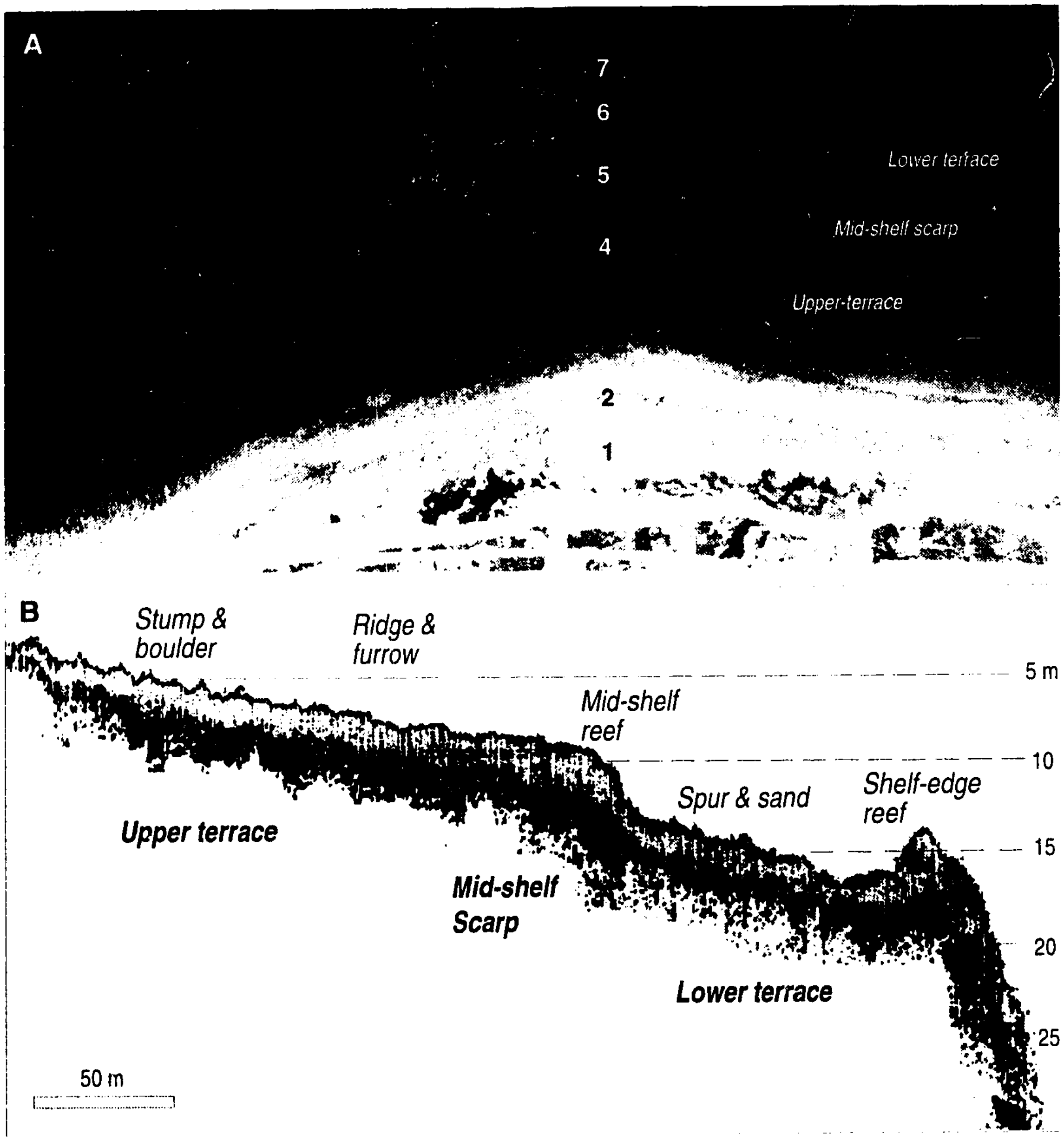

Figure 4.4. Aerial photograph (A) and sonar profile (B) showing terraces and biogeomorphic zonation of protected-windward shelf off Grand Cayman (North Side village, 2 km east of marine site N30). Upper-shelf terrace consists of; (1) narrow lagoon, (2) fringing-reef crest and flat, (3) stump-and-boulder zone, (4) ridge-and-furrow zone, (5) mid-shelf reef. Lower terrace consists of; (6) sand-and-spur zone, (7) shelf-edge reef. Note how lower terrace shows a typical exposure trend in sand-and-spur zone: wind-facing areas have frequent spurs whereas sheltered areas have less spurs and are dominated by sand. Note also the change in spur orientation.

Shelf Morphology and Zonation

The shelf around Grand Cayman consists of two distinct terraces separated by a mid-shelf scarp (Fig. 4.4). Although consistently present, their positions vary as a result of sediment deposition and reef growth (Fig. 4.3). For the purpose of discussion, the terraces of the

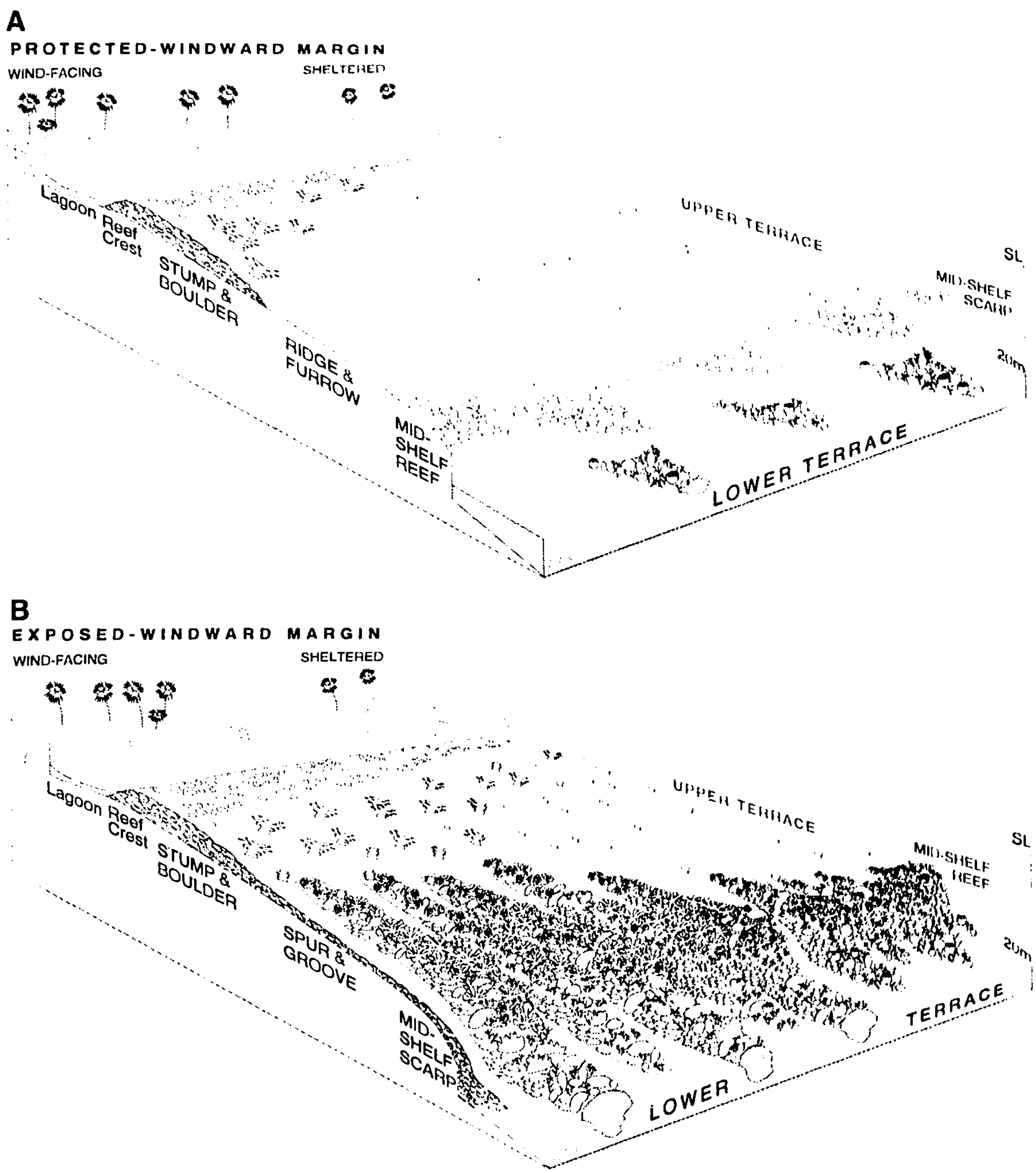

Figure 4.5. Schematic summary of exposure trends in upper-terrace zonation and substrate composition along (A) protected-windward margin and (B) exposed-windward margin. Note how enhanced surf action along wind-facing parts of the shelf produces wider stump-and-boulder and ridge-and-furrow zones than along sheltered parts. Along the exposed-windward margin, where the surf-action is strongest, coral growth is most prolific with spur and groove and better mid-shelf reef development. Note how spur-and-groove passes laterally into ridge-and-furrow.

leeward and protected-windward margins are considered together and are henceforth denoted by the term 'protected'. Zonation of the protected terraces is then compared with zonation of the exposed-windward-margin terraces, which are henceforth denoted by the term 'exposed' (see Fig. 4.5A and 5B respectively).

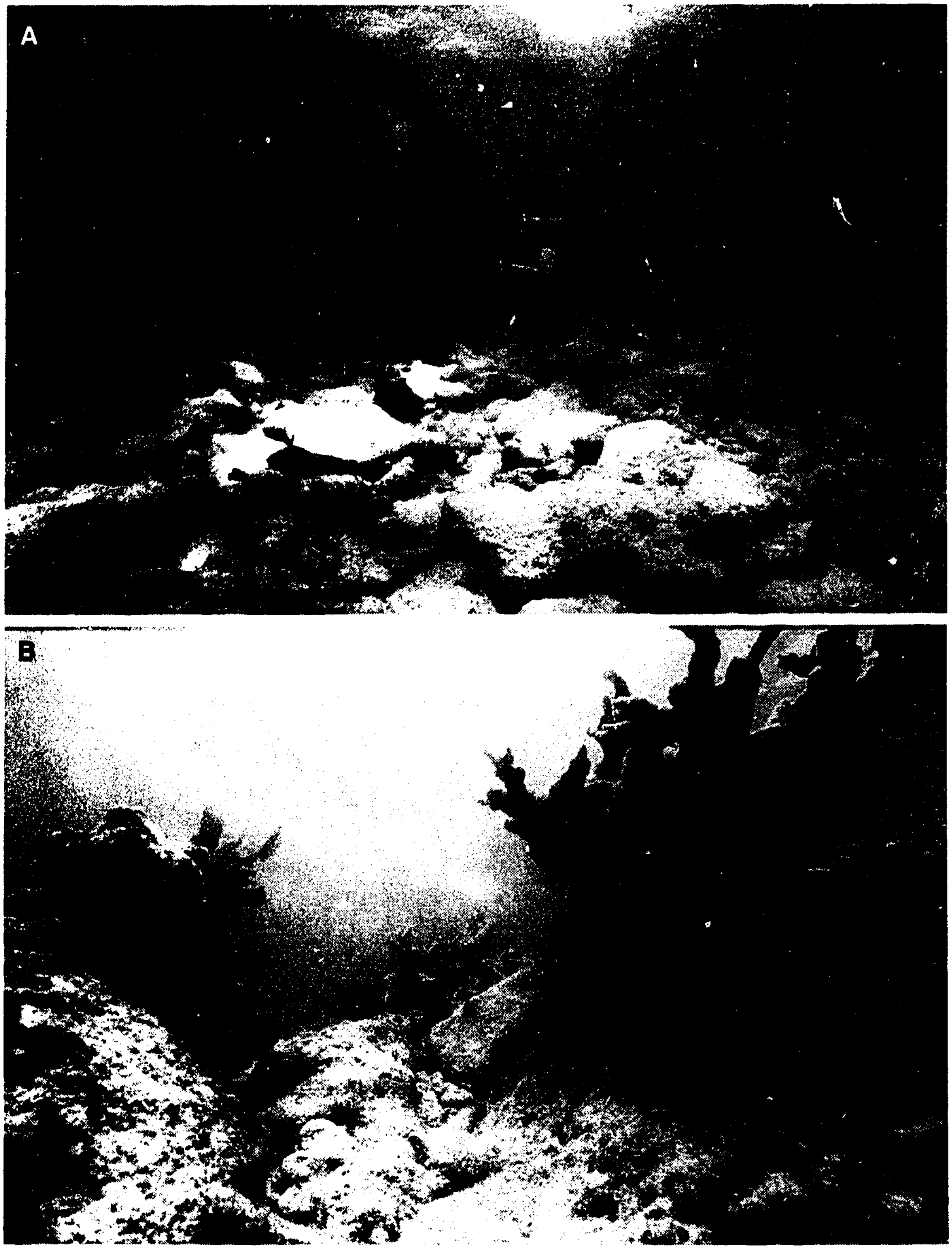

Figure 4.6. Photographs of the stump-and-boulder zone along protected-windward margin (site N30). (A) Well developed wind facing boulder field in 3.5 m of water. Note boulders are not stabilized. Diver's fins are 60 cm long. (B) Regenerated stumps of *Acropora palmata* (4 m water depth) showing evidence of cyclic growth and destruction (irregular knots in branches; compare with Fig. 4.10B). Boulder in centre is 50 cm wide.

Protected upper-shelf terrace (0-10 m)

The upper terrace along the protected margins extends from shore or reefal strandlines to the 10 m isobath. At that point, there is a distinct break-in-slope which marks the top of the mid-shelf scarp (Fig. 4.3 and Fig. 4.4). Along rocky, high-gradient shorelines, the terrace is characteristically narrow, averaging ~300 m in width, however, along low gradient shores it can reach up to ~900 m wide. The terrace surface has a concentric bio-geomorphic zonation that consists of a stump-and-boulder zone, a ridge-and-furrow zone, and a mid-shelf-reef zone. These zones vary systematically in response to local variations in wave energy (Fig. 4.5A).

Stump-and-boulder zone (0-3 m): This zone, which extends seaward from the fringing-reef crest to the 3 m isobath, is characterized by the production of detritus from, and the growth of, *Acropora palmata*. As well as *in-situ* colonies, this coral also occurs as broken *in-situ* stumps and loose cobble- to boulder-sized clasts. Zone width and characteristics vary according to shelf orientation; wind-facing parts of the terrace are dominated by wide, well developed boulder fields with clasts of *A. palmata* commonly reaching medium to large boulder-size (Fig. 4.6A). Robust colonies and broken stumps of *in-situ A. palmata* are common but sparsely distributed (Fig. 4.6B). As shelf orientation changes and becomes more sheltered, the stump and boulder zone narrows, becomes increasingly discontinuous, and eventually disappears along the leeward margin where the fringing reef is absent. As the zone narrows, *A. palmata* becomes less robust and migrates upslope onto the reef crest, eventually clustering into small patch reefs along the leeward terrace.

Ridge-and-furrow zone (0-10 m): Seaward from the stump-and-boulder zone is a broad, sculptured rock pavement that extends to the mid-shelf scarp at the terrace edge. This barren pavement is largely inhabited by a sparse, low diversity assemblage of small corals, stunted gorgonians, sponges, and brown algae (Rigby and Roberts, 1976). Its rocky surface is characteristically traversed by linear ridges and furrows (Fig. 4.7A). Along shallow, wind-facing parts of the terrace, ridges and furrows are closely spaced (< 5 m), high-amplitude

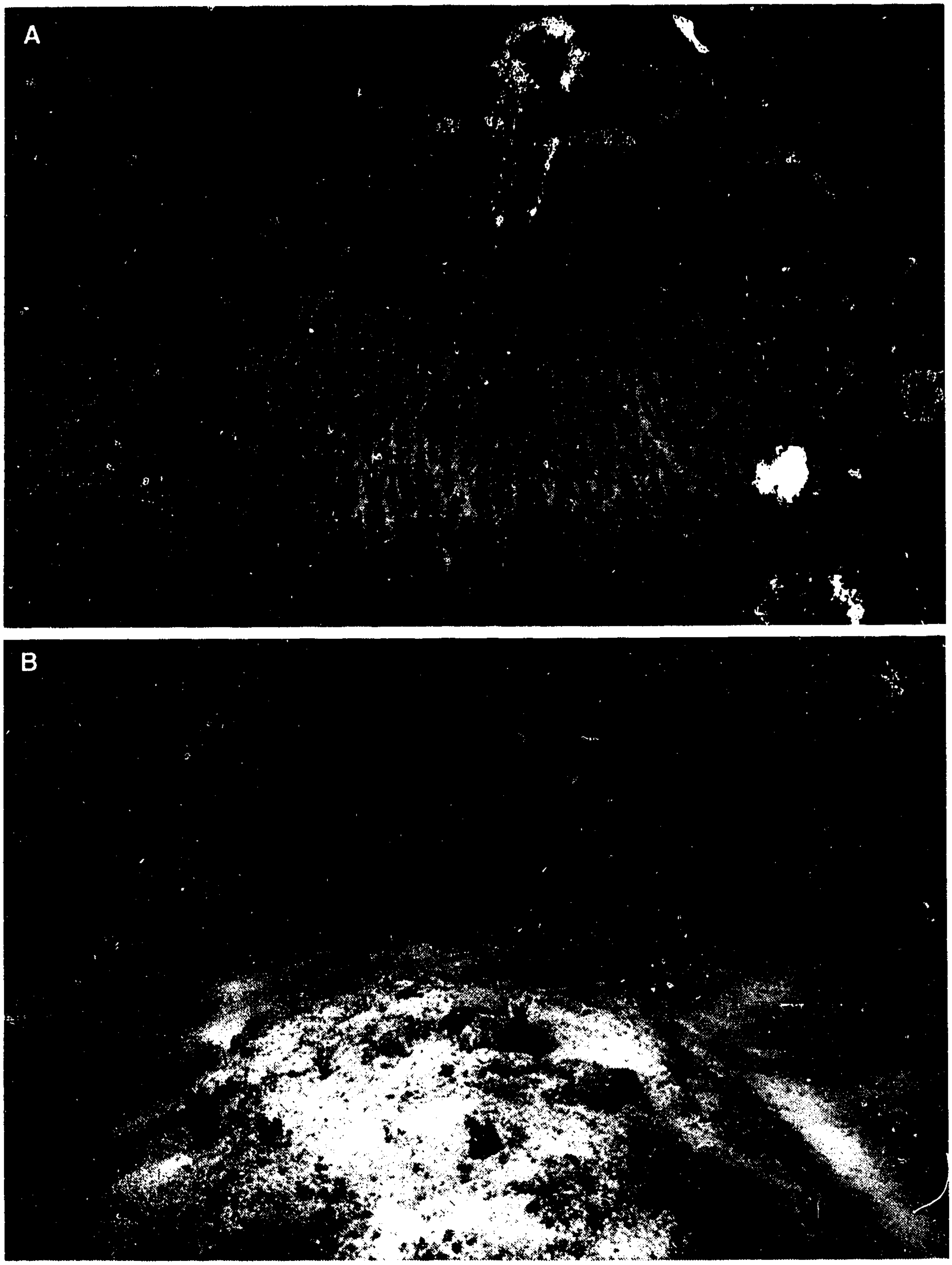

Figure 4.7. Photographs of the ridge-and-furrow zone along the protected-windward margin (transect N30). (A) Oblique aerial photograph of the ridge-and-furrow zone covering the upper terrace in a wind-facing area. Upper terrace is ~200 m wide. (B) Underwater view of the same area showing a ridge flanked by furrows in 5 m of water. Note barren nature of furrows and sickly corals and gorgonians on ridge. Ridge is ~3 m across.

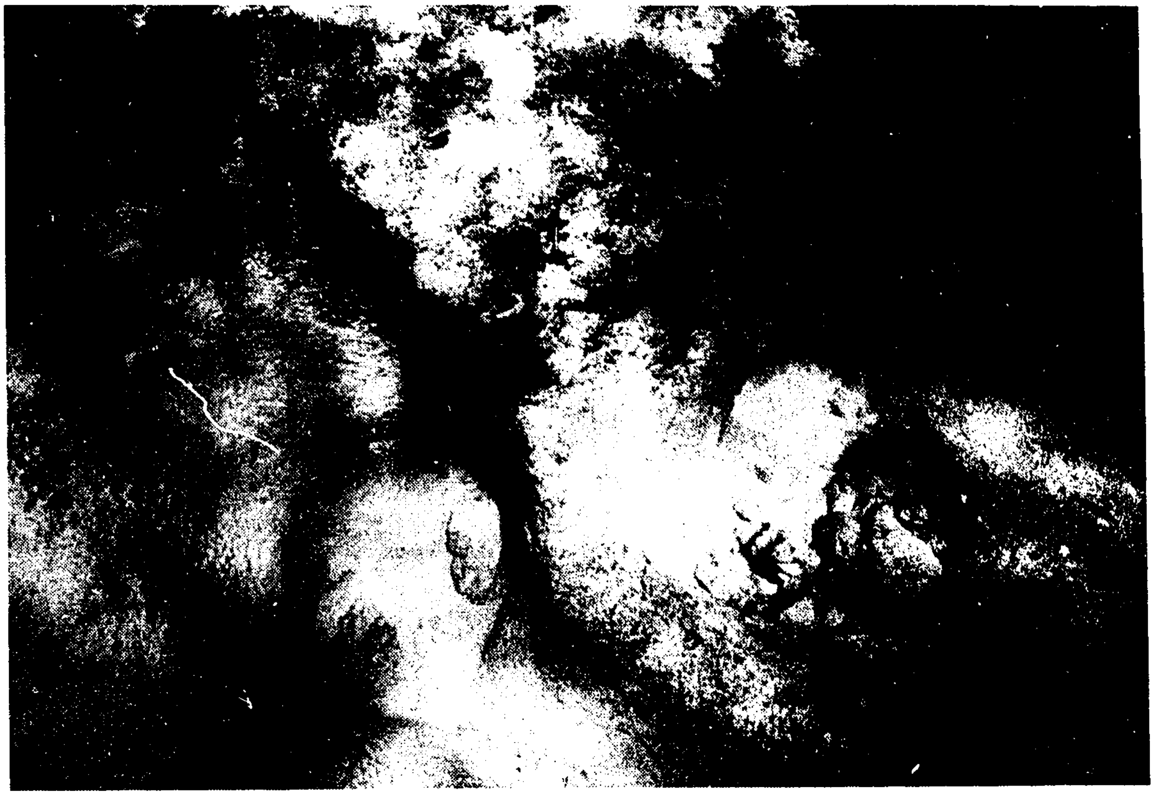

Figure 4.8. Photograph of marine potholes in the ridge-and-furrow zone along a narrow headland of the leeward margin (transect N2; water depth 3 m). The absence of fringing reefs along the leeward margin means that storm wave energy is reflected and amplified by coastal cliffs enhancing nearshore erosion. Potholes are 80 cm in diameter at the base and 2 m deep.

features (> 0.5 m), (Fig. 4.7B). whereas in deeper parts of the terrace, ridges become lower in amplitude (<0.5 m) and spacing increases (> 5 m). In more sheltered areas, the ridge-and-furrow zone is only significantly developed beneath rocky coastal cliffs, where increased erosion commonly accentuates these features, producing blind-ended furrows and pot-holes up to 2 m deep (Fig. 4.8). In all other areas the zone consists of a flat pavement with subdued furrow development occurring only along the edge of the terrace, adjacent to the mid-shelf reef (Fig. 4.5a).

Mid-shelf-reef zone (8-12 m): Along the outer parts of the upper terrace is a narrow zone of coral growth that occupies a transitional area between the ridge-and-furrow zone, the mid-shelf scarp, and the lower terrace (Fig. 4.5A). Along wind-facing areas, a diverse fauna of corals, gorgonians, and sponges tends to preferentially colonize the broad, flat-topped ridges between furrows, and in some areas extends down and across the mid-shelf scarp onto the lower terrace. In contrast, sheltered areas of the terrace generally have a more

Figure 4.9. Mid-shelf-reef development along the leeward margin (W40) showing typical sparse coral-coverage concentrated near the shoulder of the mid-shelf scarp (large *M. annularis* coral is 1.5 m tall; water depth is 18 m at base of mid-shelf scarp and 12 m at the shoulder). Contrast with Fig. 4.12.

patchy reefal development, and the mid-shelf scarp is generally better exposed (Fig. 4.9). In both areas furrows remain uncolonized and commonly extend into steep gullies which dissect the mid-shelf scarp at regular intervals.

Exposed upper-shelf terrace (0-10 m)

The upper terrace along the exposed margin is generally an area of luxuriant coral growth. As a result, the position of the terrace surface, and the break-in-slope that delineates its outer edge, are not as well defined as they are in protected areas. Nevertheless, terrace depth ranges are similar to protected upper terrace and range down to the 10 m isobath. The exposed upper terrace is divided into concentric bio-geomorphic zones consisting of the stump-and-boulder, spur-and-groove, ridge-and-furrow, and mid-shelf-reef zones. Like the protected margin, the characteristics of these zones vary systematically in response to local variations in wave energy (Fig. 4.5B).

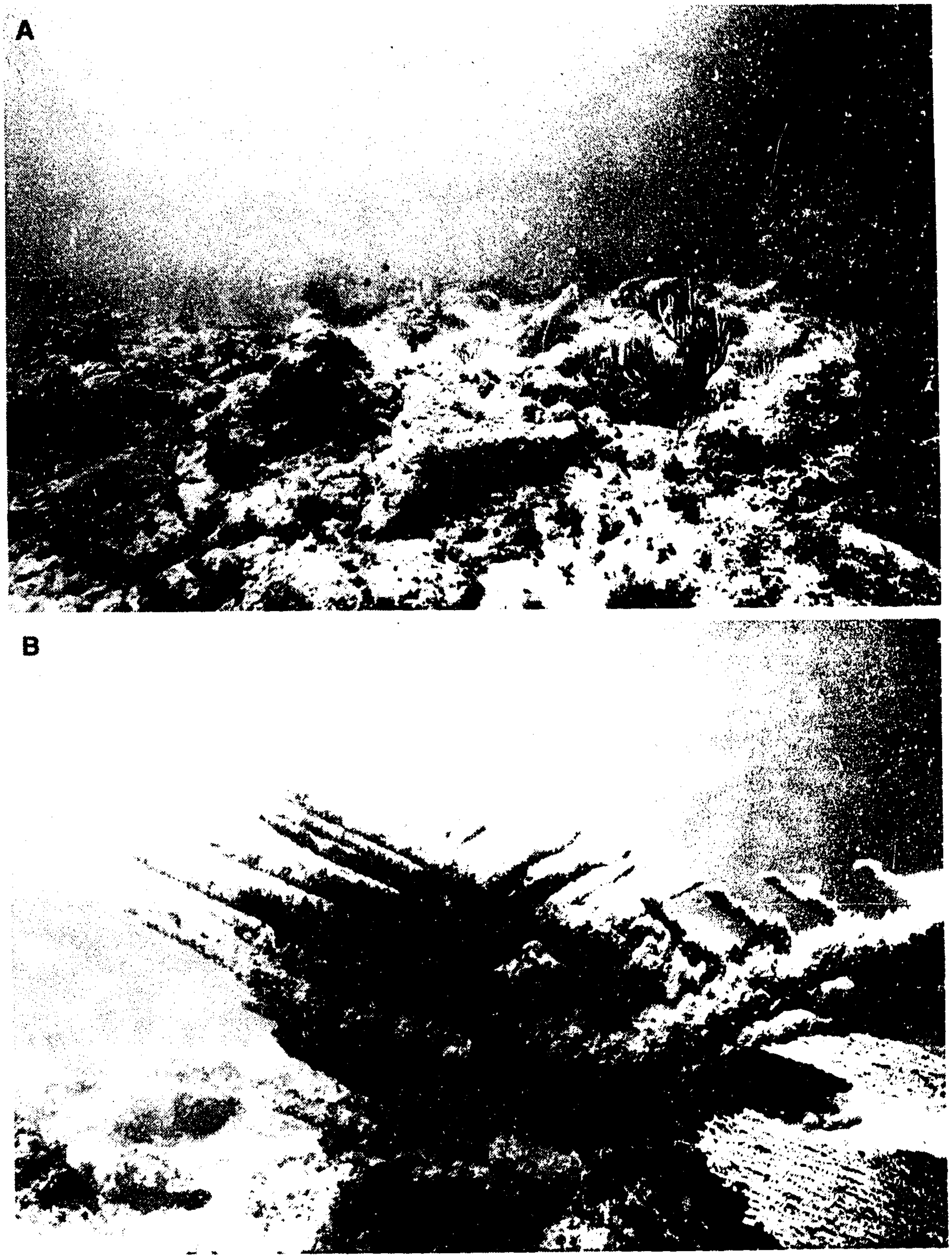

Figure 4.10. Photographs of stump-and-boulder zone along exposed-windward margin (sites S5 and S24). A) Typical boulder field in 4 m of water along a wind-facing area. Boulders are stabilized by extensive crusts of coralline red algae (boulders in foreground are 1 m in length). B) Live colony of *A. palmata* showing morphology adapted to strong fairweather surf-action. Colony is 1 m tall and is in 3 m of water.

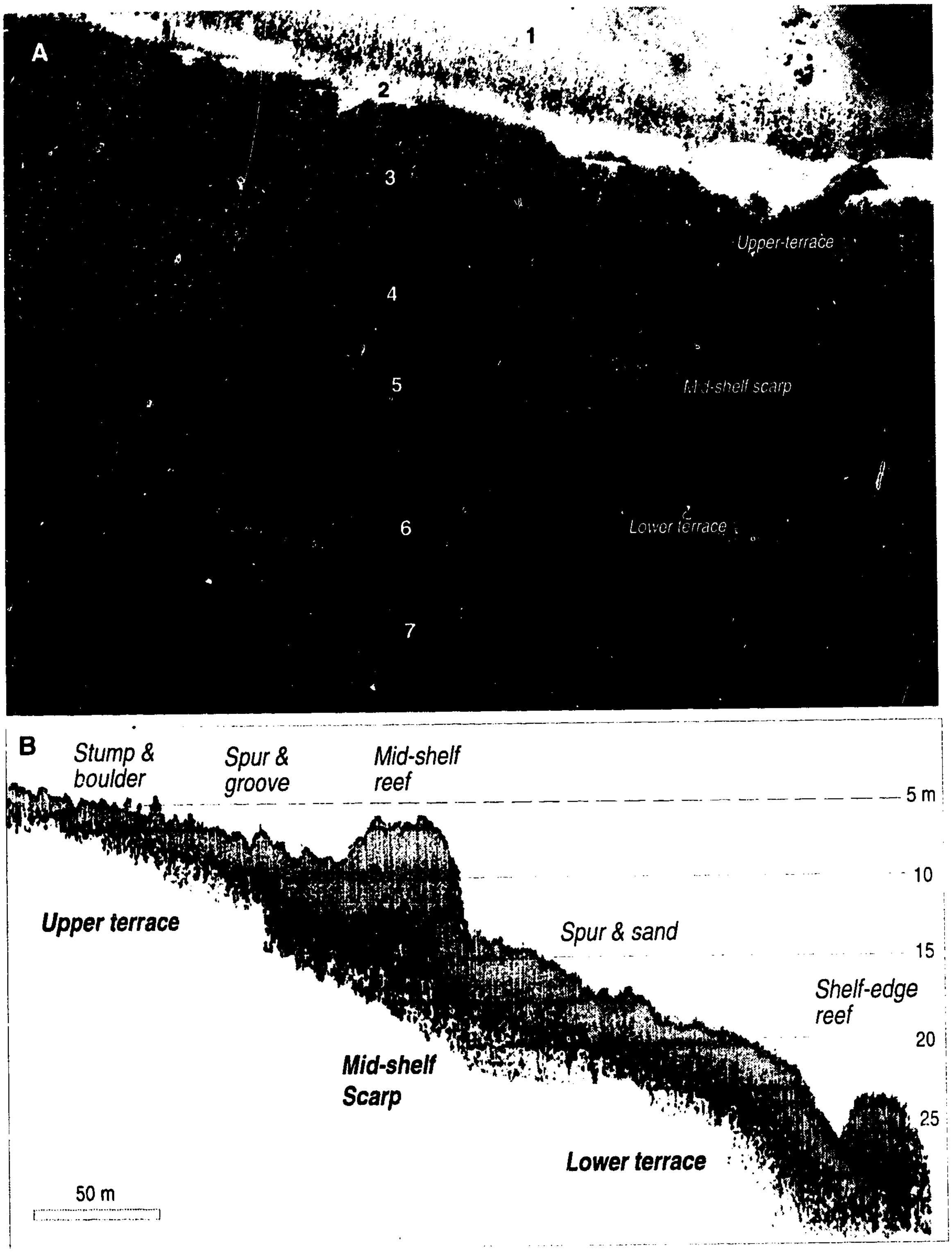

Figure 4.11. Aerial photograph (A) and depth-sound profile (B) showing terraces and bio-geomorphic zonation along the exposed-windward shelf of Grand Cayman (transect S5). Upper-shelf terrace; lagoon (1), fringing-reef crest and flat (2), stump-and-boulder zone (3), spur-and-groove zone (4), mid-shelf-reef zone (5). Mid-shelf scarp mostly buried. Lower-shelf terrace; sand-and-spur zone (6), shelf-edge reef (7).

Figure 4.12. Mid-shelf reef development along exposed-windward (S24) showing typical dense coral coverage obscuring mid-shelf scarp. Note roofed-over groove below diver (who is 2 m tall in fins; water depth 8 m). Contrast with Fig. 9.

Figure 4.13. Photograph of drowned inter-tidal notch at -18.5 m cut into mid-shelf scarp (notch is 5 m high; site N2). Lateral persistence and drill cores demonstrate notch is an erosional feature cut into the Pleistocene Ironshore Formation. Preservation of notch in such pristine condition implies it was drowned by rapid sea-level-rise event (see Fig. 4.19). So far, notches at this depth have been identified around five other islands (see Fig. 4.20).

Stump-and-boulder zone (0-5 m): In wind-facing areas of the exposed upper-terrace, clasts, stumps, and colonies of *A. palmata* cover a wide area in front of the fringing reef to a depth of 5 m. In the vigorous shallow water adjacent to the reef crest, cobble and boulder-sized clasts are stabilized by coralline red algae, which form undulating crusts over broad areas (Fig. 4.10A). These coralline crusts also cover broken stumps of *A. palmata* that are common, along with robust live colonies, in the deeper parts of the zone (Fig. 4.10B). In more sheltered areas of the terrace, the zone is narrower and not as well developed, with stumps and live colonies being less abundant and clasts more patchily distributed. This exposed-terrace stump-and-boulder zone is equivalent to the *palmata* zone of Goreau (1959).

Spur-and-groove zone (5-15 m): Seaward of the stump-and-boulder zone, along wind-facing areas of the exposed upper-terrace, is a zone of regimented coral spurs (Fig. 4.5B and

Figure 4.14. Lower terrace spur-and-sand zone along wind-facing section of the exposed-windward margin at a depth of ~20 m looking seaward (transect TC). Shows typical high amplitude (3.5 m) and closely-spaced spurs (diver in a 5-m wide sand channel).

11A). In shallow turbulent areas spurs start as an alignment of dead stumps, juvenile *A. palmata* colonies, and small head-corals. These proto-spurs, arranged in rows projecting downslope, are separated by narrow <3-m-wide grooves floored with cobble-sized clasts of *A. palmata*. Further downslope, these proto-spurs quickly develop into large spurs (up to 3 m relief and 3 m across) topped by dense thickets of *A. palmata* and separated by 5 m wide grooves. The vertical to overhanging sides of the spurs are reinforced by large, shingled head-corals such as *Montastrea annularis*. In deeper parts of the zone, spur tops are increasingly colonized by the delicately branched coral *Acropora cerviconis* and massive *M. annularis*. This zone corresponds to the spur-and-groove zone described by Shinn (1963) and the buttress zone of Goreau (1959).

Ridge-and-furrow zone (3-8 m): In more sheltered locations of the exposed upper-terrace, the spur-and-groove zone is replaced laterally by the ridge-and-furrow zone (Fig. 4.5B).

The form of the ridges and furrows is similar to that found on wind-facing parts of the protected upper-terrace. Faunal characteristics, however, are slightly different and ridges have a dense population of gorgonians as well as scattered corals (small heads and *A. palmata* stumps). Furrows are typically barren with a smooth U-shaped rocky surface.

Mid-shelf-reef zone (5-12 m): Along the outer parts of the ridge-and-furrow and spur-and-groove, is a zone of increased coral growth that extends down the mid-shelf scarp onto the lower terrace (Fig. 4.5B). In wind-facing areas this mid-shelf reef accentuates the development of spurs that extend down from the spur-and-groove zone. In sheltered areas, where spur-and-groove is replaced by ridge-and-furrow, reef development is concentrated in a series of coalescing patch reefs along the terrace edge (Fig. 4.12). Coral growth is typically luxuriant and commonly grows to within 5 m of sea level where thickets of *A. palmata* and large head corals dominate (Fig. 4.12). Grooves maintained by storm-wave surge are commonly roofed-over and form anastomosing tunnel systems. This zone is equivalent to the *annularis* zone of Goreau (1959).

Mid-shelf scarp (10-20 m)

The upper-shelf terrace along both exposed and protected margins is terminated by a significant break-in-slope that separates it from the lower terrace (Fig. 4.3). Generally expressed as a small scarp, this slope break is a site of profuse coral growth associated with the development of the mid-shelf reef (Fig. 4.5). The mid-shelf scarp is equivalent to the "step" of James and Ginsburg (1979) and the "fore-reef escarpment" of Goreau and Land (1974). In wind-facing areas of the exposed margin, the mid-shelf scarp is commonly obscurred by the mid-shelf reef which grows down onto the lower terrace. In such areas, the scarp's position is only marked by a subtle gradient change. In contrast, along sheltered areas of the exposed margin, coral growth is concentrated on top of the scarp and greatly accentuates its relief (Fig. 4.5 and 4.11B).

Along the protected margins, in addition to being overgrown with coral, the scarp is also buried by accumulations of skeletal sand. These deposits almost completely bury the scarp

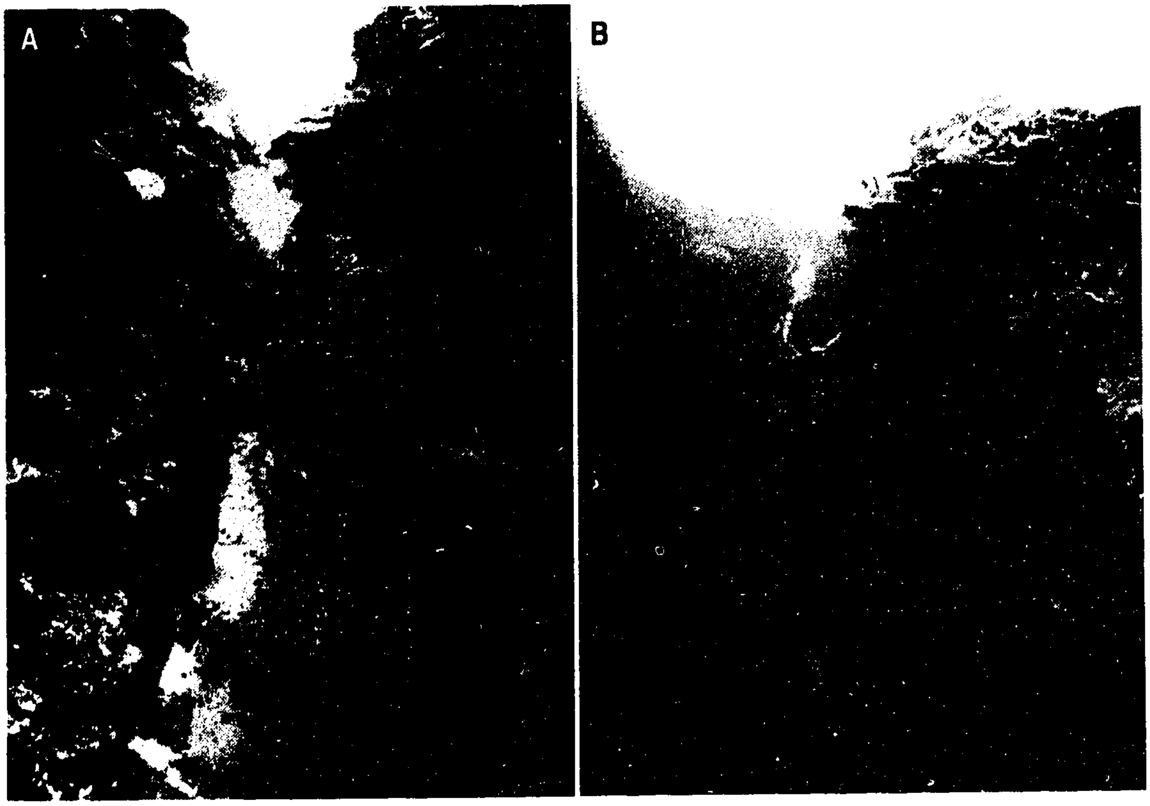

Figure 4.15. Lower-terrace shelf-edge-reef zone. (A) Steep canyon through shelf-edge reef along exposed-windward margin in 36 m of water (site GB; *M. cavernosa* in foreground is 50 cm tall). (B) Vertical to over-hanging shelf-edge-reef wall along the protected-windward margin in 35 m of water (site N30). Diver in background is 2 m in fins.

in sheltered sites, and it is only along headlands and wind-facing areas that it is exposed to any appreciable extent and displays up to 10 m of relief. In one of these headland areas, a 2 km section of scarp is exposed sufficiently to display an intertidal notch at a consistent depth of -18.5 m (Fig. 4.13). This drowned notch is in pristine condition and differs little in morphology from notches along the adjacent modern shoreline.

Lower-shelf terrace (12-40 m)

The lower terrace extends from the base of the mid-shelf scarp to the edge of the shelf (Fig. 4.4 and Fig. 4.11). Along rocky coasts it is typically only ~150 m wide, but increases to ~300 m along low-gradient coasts. In all areas the terrace is a site of active sediment/reef accumulation and, as a result, there is a significant degree of variation in the position of the terrace surface (Fig. 4.3). In general, this surface is shallow and flatter along protected margins. Along exposed-windward areas, however, the terrace surface starts in relatively deeper water and slopes more steeply before plunging into the shelf-edge escarpment (which typically starts between 60 to 70 m). The terrace surface has two zones, the spur-and-sand zone and the shelf-edge-reef zone. The characteristics of these zones vary systematically in response to changes in shelf margin orientation (Fig. 4.4).

Spur-and-sand zone (12-30 m): The inner part of the lower terrace is composed of a broad sandy plain crossed at regular intervals by seaward-projecting coral spurs (Fig. 4.4). In general, sand accumulation dominates the leeward margin and sheltered areas of the protected-windward margin, whereas spur formation dominates wind-facing areas of the windward margin. Along the exposed-windward shelf, wind-facing areas are characterized by narrow, closely spaced spurs (<25 m crest to crest) that extend across the full width of the terrace (Fig. 4.14). Individual spurs have relatively high amplitudes (averaging 4 m of relief) and slope from 15 to 25 m depth. As the shelf becomes more protected, spurs broaden and develop increasingly wider spacing(>25 m). Ultimately, in sheltered areas of the exposed-windward terrace, spurs give way to broad expanses of skeletal sand.

Spurs along wind-facing areas of the protected-windward shelf, are closely spaced (<25 m) and extend across most of the terrace (Fig. 4.4). Spur amplitudes and gradients, however, are much lower than those on the exposed-windward shelf, and they tend to have less than 2 m of relief. As the shelf becomes more sheltered, spurs are rapidly replaced by broad expanses of skeletal sand (Fig. 4.4).

Shelf-edge-reef zone (15-50 m): The outer part of the lower-terrace is rimmed by a significant and continuous coral reef. This shelf-edge reef is composed of an array of coral-mantled buttresses aligned perpendicular to shore, and regularly dissected by sediment-floored canyons (Blanchon and Jones, 1993). Individual buttresses, which rise from ~50 m, have steeply sloping to overhanging fronts and sides that are armored by platy corals (*Agaricia* sp., *Montastrea annularis*). The upper parts of the buttresses, which grow into waters as shallow as 15 m, are characterized by a diverse, head-coral-dominated biota (including

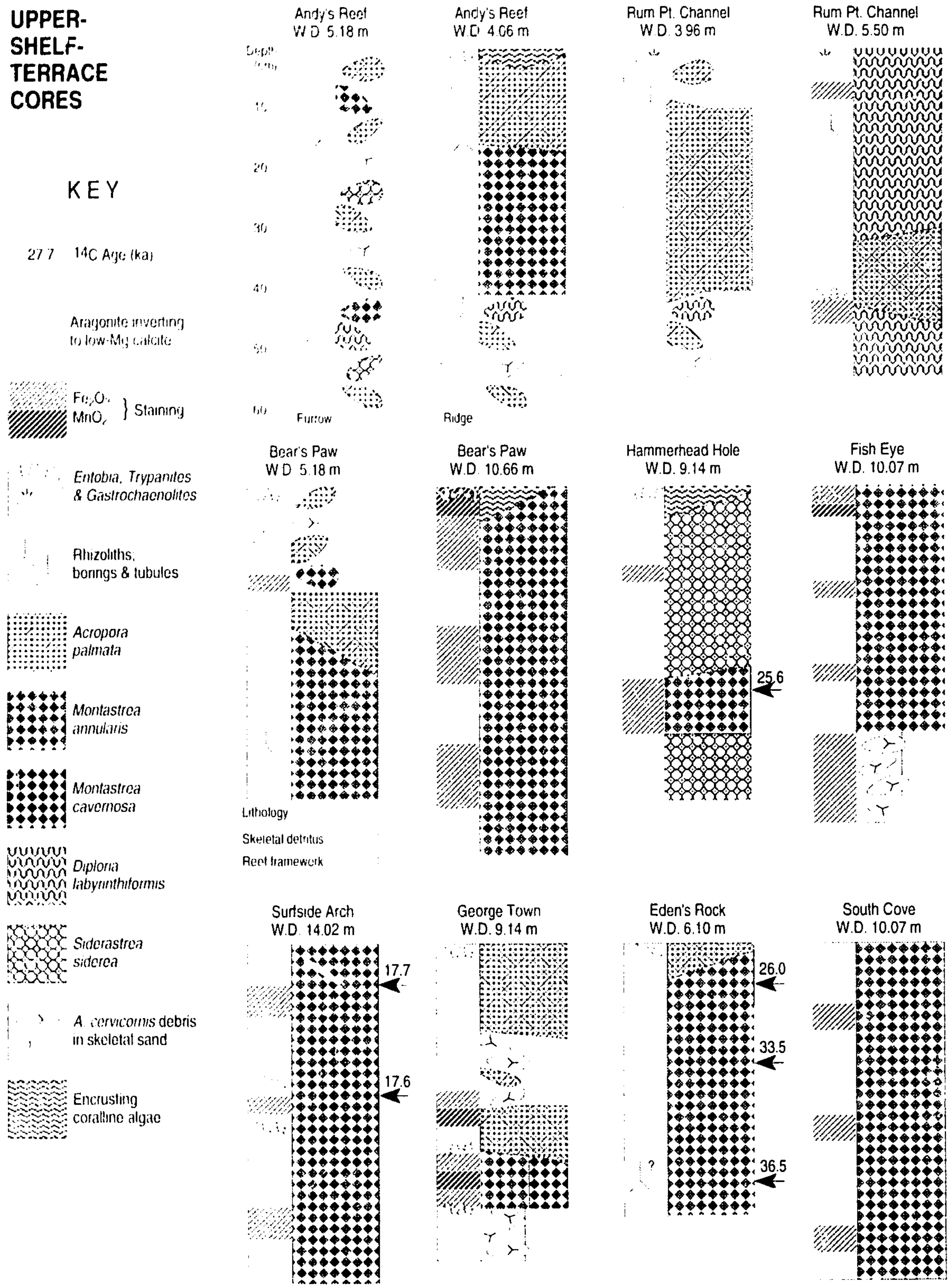

Figure 4.16. Selected cores from upper-terrace ridge-and-furrow zone along north and west sides of Grand Cayman. All cores show evidence of exposure to meteoric conditions including ferric and manganese oxide staining which coats coral framework, skeletal detritus, and meteoric cements. These oxides could only be sourced from the island's terra-rosa soils and, since they coat meteoric cements, were probably deposited during an episode of falling sea level.

Montastrea annularis, *M. cavernosa*, *Diploria labyrinthiformis*, and *Colpophylia natans*). The buttresses typically extend into shoreward-projecting coral spurs.

The width, amplitude, and depth of the buttresses and their spurs show a significant exposure trend as shelf orientation changes. Along the exposed-windward shelf, buttresses and spurs tend to develop in relatively deeper positions (>25 m) and are dissected by vertical sided, narrow canyons (Fig. 4.15A) that slope steeply into deeper waters (canyon outflow is commonly >50 m). In contrast, along the protected-windward shelf, buttresses grow in relatively shallow positions (<25 m) and develop spectacular frontal escarpments that overhang in many locations (Fig. 4.15B). These shallow buttresses and their low amplitude spurs are frequently dissected by vertical-sided, shallow canyons that slope gently down to the terrace edge (canyon outflow typically <50 m). Along the leeward shelf, buttresses grow in the most shallow positions of all, and are commonly found in ~15 m water depth. There, they are typically wider and are dissected by broad channels at less frequent intervals.

Substrate composition and thickness

The upper-shelf terrace is an erosional limestone pavement that provides the foundation for modern coral growth. Cores from the ridge-and-furrow zone (particularly those drilled on adjacent ridges and furrows; see Andy's-Reef on Fig. 4.16) and the mid-shelf scarp (including the -18.5 m notch), confirm that these features are erosional and show that the terrace substrate is composed of shallow-water reefal limestones (Fig. 4.16). These deposits are dominated by compact coral-headstones and mega-grainstones (see classification of Blanchon, 1992) and display evidence of pervasive meteoric diagenesis, including vadose cements, speleothems, rhizoliths, ferric- and manganese-oxide staining, aragonite to low-Mg calcite inversion, and solution-enlarged intraskeletal pore networks.

Most of these upper-terrace limestones are lithologically and diagenetically akin to limestones of the Ironshore Formation, which commonly crop out along the coast of Grand Cayman (Matley, 1926; Brunt et al., 1973; Hunter and Jones, 1988). The upper part of this formation was formed during the last (Sangamon) interglacial, and has been dated at 121 ±6

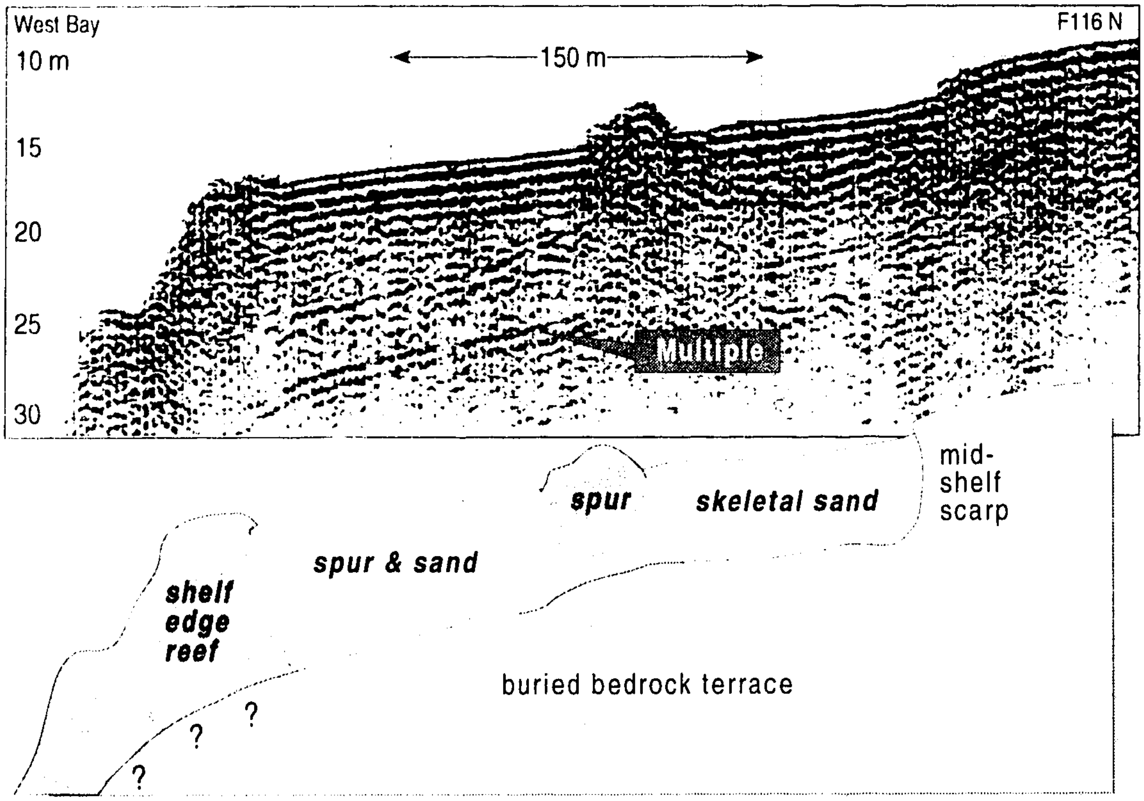

Figure 4.17. Seismic profile across the lower-shelf terrace on the west side of Grand Cayman (near transect W14). Interpretation below shows the typical architecture of modern deposits on the lower terrace, and the geometry of the bedrock terrace and mid-shelf scarp underlying them.

ka (Woodroffe et al., 1983; Jones and Hunter, 1990). Radiocarbon dates from corals in the ridge-and-furrow-zone cores give 'ages' between 17.5 and 36.5 ka. These dates are considered to be anomalous and represent the age of meteoric-cementation events (cf. Newell, 1961; McCullough and Land, 1992). This contamination by younger carbon causes a rejuvenation of the true coral ages that, on the basis of similarity with Ironshore deposits, are suspected to be at least 121 ka old.

The growth and accumulation of modern deposits on the eroded upper-terrace limestones is limited by water-depth (<10 m) and varies with shelf orientation. Along the protected-windward shelf, the stump-and-boulder zone is ~3 m thick, but increases to ~5 m on the exposed-windward shelf. Cores from both areas consist of a mega-grainstone facies (Blanchon, 1992) composed of cobble- to boulder-sized clasts of *A. palmata* set in a marine-cemented, skeletal-sand matrix. The same facies also occurs in grooves of the spur-and-groove zone. Although cores did not fully penetrate a complete sequence of this facies, it is estimated to be <2 m thick. Spurs in the spur-and-groove zone, which are up to 5 m thick, are composed of a grainy, compact-headstone facies (Blanchon, 1992) dominated by *in-situ* heads of *Montastrea annularis* with *Millepora* mega-grainstones filling interstices. Cores from the mid-shelf-reef zone are composed of similar *in-situ* growth textures but poor core coverage prevented the facies from being fully characterized.

The lower-shelf terrace is an area of active sediment accumulation and reef growth, and, without exception, cores from the sand-and-spur and shelf-edge-reef zones yielded radiocarbon ages of <1000 yrs BP. The thickness of these modern deposits increases toward the shelf edge where they are seismically estimated to be ~25 m thick, ranging up to 40 m in some exposed-windward areas. All modern deposits on the lower terrace are underlain by a smooth, gently sloping surface that produces a distinct reflector on seismic profiles (Fig. 4.17). Although it could not be cored, drill-probing, seismic geometry, and regional continuity confirm that the reflector is a sediment/bedrock interface. On the leeward shelf, the reflector extends from the base of the mid-shelf scarp (known to be bedrock) at 20 m, and slopes gently to 30 m before disappearing beneath the shelf-edge reef. The reflector is also apparent along the reef-dominated exposed-windward shelf, but its continuity is poor due to refraction and attenuation of the seismic signal by the overlying reef deposits. The geometry of this surface on the exposed-windward margin is confirmed, however, by depth-sounding profiles in areas with unusually limited reef development (Fig. 4.3, localities S31 and TC). In such areas, the lower-terrace surface slopes gently from 20 to 40 m, reflecting the geometry of the underlying bedrock terrace, before rapidly steepening into the shelf-edge escarpment.

Formation of Shelf Terraces

Morphological features on the upper-shelf terrace around Grand Cayman are the result of fairweather and storm processes. Under fairweather conditions, coral growth, sediment deposition and bioerosion dominate. Although fairweather wave action causes some sediment movement, scuba observations suggest little mechanical erosion takes place. The erosional ridge-and-furrow zone on the upper-shelf terrace must, therefore, be produced during seasonal storm conditions.

Historical records show that Grand Cayman is affected by severe storms and hurricanes on a regular basis. Hurricanes pass directly over the island on average once every ten years (Clark, 1988), although secular variations in storm-recurrence intervals are common. Eyewitness accounts of a particularly severe hurricane that affected the Cayman Islands in 1932, provides some insight into marine processes:

> ...early on Wednesday the 7th the wind whipped up, and in twenty minutes grew from a calm to hurricane intensity...The sea swept high over the coast, carrying huge rocks on its crest and the wind hurled rocks, some weighing tons, through the air...Everything lay buried beneath a mass of broken coral, boulders, trees and the dismal wreckage of houses. (Williams, 1970).

Such accounts, which serve as chilling testimony to the awesome power of hurricane-generated waves, demonstrate that storm waves entrain a wide spectrum of sediment and clasts from various shelf and coastal sources (e.g., Jones and Hunter, 1992). As these sediment-charged storm waves reach the mid-shelf scarp, the abrupt change in depth causes them to spill onto the upper terrace, transforming it into an expansive surf-zone. These powerful breakers have three main effects on the upper terrace. First, they destroy stands of live coral in the stump-and-boulder and spur-and-groove zones (Fig. 4.18), providing a copious source of large clasts. Second, this debris is entrained in saltation and traction loads and repeatedly scours back-and-forth along the terrace surface, eroding furrows and drilling-out deep potholes into the soft limestone bedrock (Fig. 4.7 and 4.8). Third, sand and pebble-sized sediment in suspension erodes the terrace surface by sand-blasting. The efficacy of these storm-abrasion mechanisms is related to the supply of sediment and clasts. Sand and pebble-sized grains are widely available from the sandy lower terrace, and sand-blasting consequently affects large areas of the upper-terrace surface. In contrast, cobble and boulder-sized clasts are only available in appreciable quantities from areas with significant coral growth, such as the stump-and-boulder and spur-and-groove zones. Consequently, areas proximal to these zones show well developed, high amplitude ridge-and-furrows that

Figure 4.18. Before (A) and after (B) photographs illustrating destructive effects of Hurricane Gilbert (1988) on spur-top coral communities (~5 m water depth) along exposed-windward margin (site S5). Triggerfish is ~40 cm long. Courtesy of Phil Bush.

grade into low-amplitude features as distance from the clast source increases. A similar relationship occurs along the edge of the upper terrace where the mid-shelf reef is the source of clasts for furrow formation; furrows are well developed close to the reefal source but die-out away from it (Fig. 4.5A).

Erosion during storms has clearly played an important role in the formation of the upper-she[illegible] and coastal cliffs around the island of Grand Cayman. During fairweather, bioerosion weakens and undermines non-growth substrates, rendering them susceptible to mechanical erosion during storms and hurricanes. Over extended periods of time, substrate degradation and cliff collapse produce a smooth seaward-sloping terrace backed by coastal cliffs. The upper-terrace/coastal-cliff geomorphic unit are the product of ongoing erosional processes that started when sea-level first reached the position of the upper terrace. The lower-terrace and mid-shelf scarp, however, are an older geomorphic unit related to an earlier sea-level stillstand. Although blanketed by a thick accumulation of modern sediments, the lower-terrace is underlain by a smooth, seaward-sloping bedrock surface that represents an older marine planation surface. At the head of this buried planation surface is a drowned sea-cliff (mid-shelf-scarp) cut during this older episode of erosion. Strong evidence supporting this terrace/cliff interpretation is provided by the -18.5 m intertidal notch developed in the mid-shelf scarp (Fig. 4.13). This feature records the position of sea-level during the earlier episode of stabilized sea level.

Origin of Shelf Terraces

The terraced nature of the shelf around Grand Cayman owes its origin to variations in the rate of sea-level rise. During slow-rise or stillstand episodes, terrace/cliff units were formed by marine erosion. These units were then drowned by a rapid sea-level-rise event. The rate and magnitude of this rise event is illustrated by the preservation of an intertidal notch developed at -18.5 m on the mid-shelf scarp. Measured profiles of the notch and visor suggest that it suffered little erosion during the rise event (Fig. 4.13). Because cores from the notch are composed of soft limestone prone to relatively high rates of intertidal erosion,

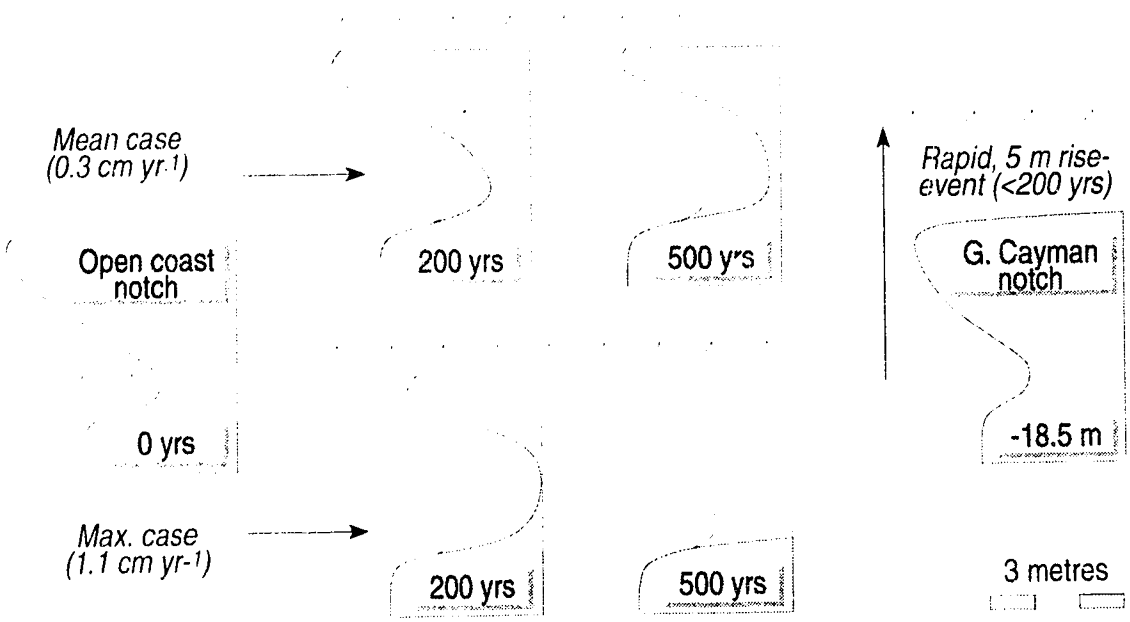

Figure 4.19. Simulated erosion of an intertidal notch at different rates of sea-level rise, using conservative erosion rates measured from Grand Cayman (from Spencer 1985). The morphology of the starting-point notch (0 yrs) is based on notches from open, microtidal coastal areas with regular wave action (cf. Pirazzoli et al., 1991). The morphology of the drowned notch at -18.5 m (Figure 13) is shown at the same scale to illustrate the lack of erosion suffered during the actual sea-level rise. This implies that the actual sea-level-rise event must have occurred rapidly, in <200 years, and completely submerged the mid-shelf scarp, i.e., 5 m plus.

notch preservation must be attributed to a rapid rate and large magnitude of sea-level rise. Simulated marine erosion using conservative modern intertidal and subtidal erosion rates (from Spencer, 1985), suggest that the notch and mid-shelf scarp were drowned by a 5 to 8 m sea-level rise in less than 200 years (Fig. 4.1

Available evidence suggests such variations in the rate of sea-level rise were eustatic in origin. Regional investigations into the tectonic history of the North American-Caribbean Plate Boundary Zone indicate that the last phase of tectonic activity to affect the Cayman Ridge ended ~5 Ma (Lewis et al., 1990; Pindell and Barrett, 1990). This stability is also supported by the consistent positions of the shelf terraces around the island (Fig. 4.3). More importantly, however, the shelf terraces around Grand Cayman correlate with submerged terraces from islands in the Atlantic, Pacific, and Indian Oceans (Fig. 4.20), thereby suggesting that terraces on these islands were also produced by the same eustatic sea-level changes that affected Grand Cayman.

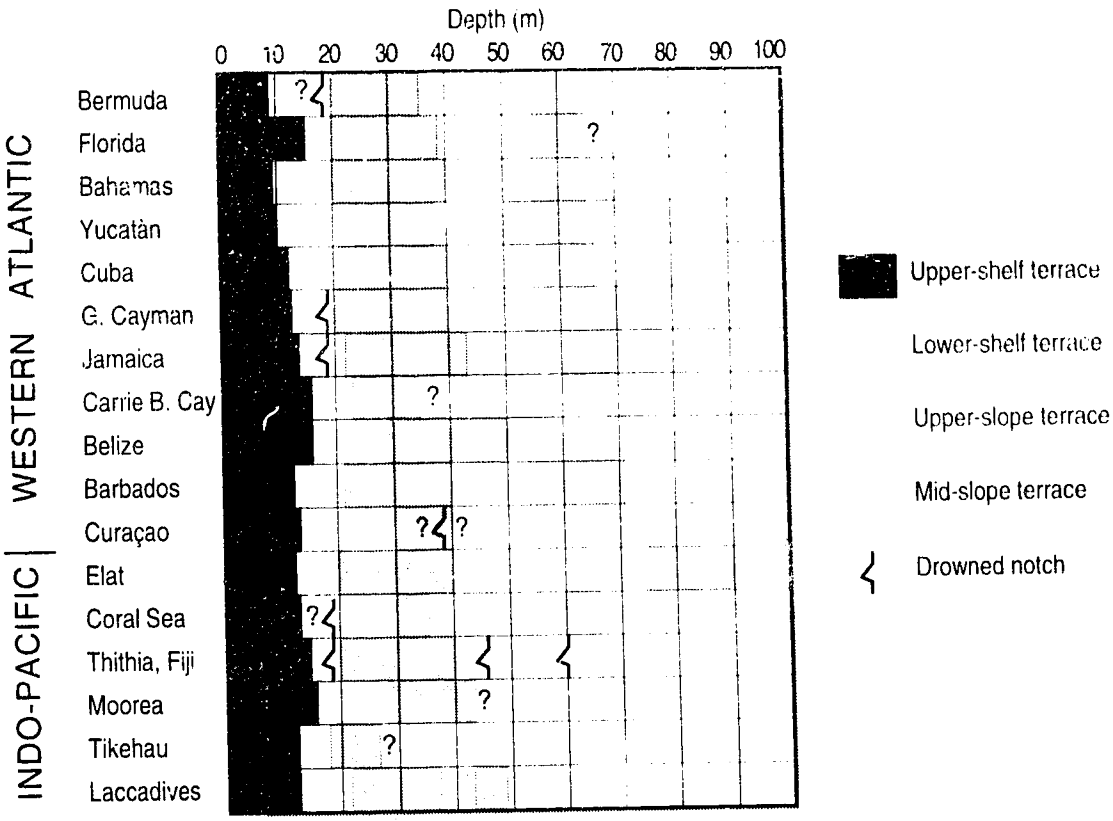

Figure 4.20. Depth range of terrace surfaces and positions of intertidal notches from islands in the western Atlantic, western and south Pacific, Red Sea and Indian Ocean. Variations in the depth range of terrace surfaces is due to accumulation of modern sediments and reefs. Bermuda—Meischner & Meischner (1977); Stanley & Swift (1968). Florida—Lidz et al. (1991). Bahamas—Hine & Neumann (1977); Wilber et al. (1990). Yucatán—Logan et al. (1969). Cuba—Kühlmann (1970). Grand Cayman—this study; Raymont et al. (1976). Jamaica—Goreau & Land (1974); Liddell et al. (1984); Digerfeldt & Hendry (1987). Carrie Bow Cay—Rützler & Macintyre (1982). Belize—James & Ginsburg (1979). Barbados—Acker (1987); Macintyre (1967). Curaçao—Focke (1978). Elat, Red Sea—Reiss & Hottinger (1984). Osprey Reef, Coral Sea—Sarano & Pichon (1988). Thithia, Fiji—Phipps & Preobrazhensky (1977). Moorea, Polynesia—Vénec-Peyré (1991). Tikehau, Polynesia—Harmelin-Vivien (1985). Laccadive Islands, Indian Ocean—Siddique (1975).

Age of Shelf Terraces

The age of the eustatic stillstands that produced the terraces around Grand Cayman can be constrained by two independent lines of evidence. First, the terraces are cut into the Ironshore Formation which consists of reefal limestones deposited ~121 ka ago (Woodroffe et al., 1983). Thus, terrace formation must have occurred after these deposits were formed, and consequently must be the product of eustatic sea-level oscillations associated with the last glacial cycle. Second, a correlation between shelf terraces in areas with a history of tectonic uplift suggest terraces around Grand Cayman formed during the last deglaciation.

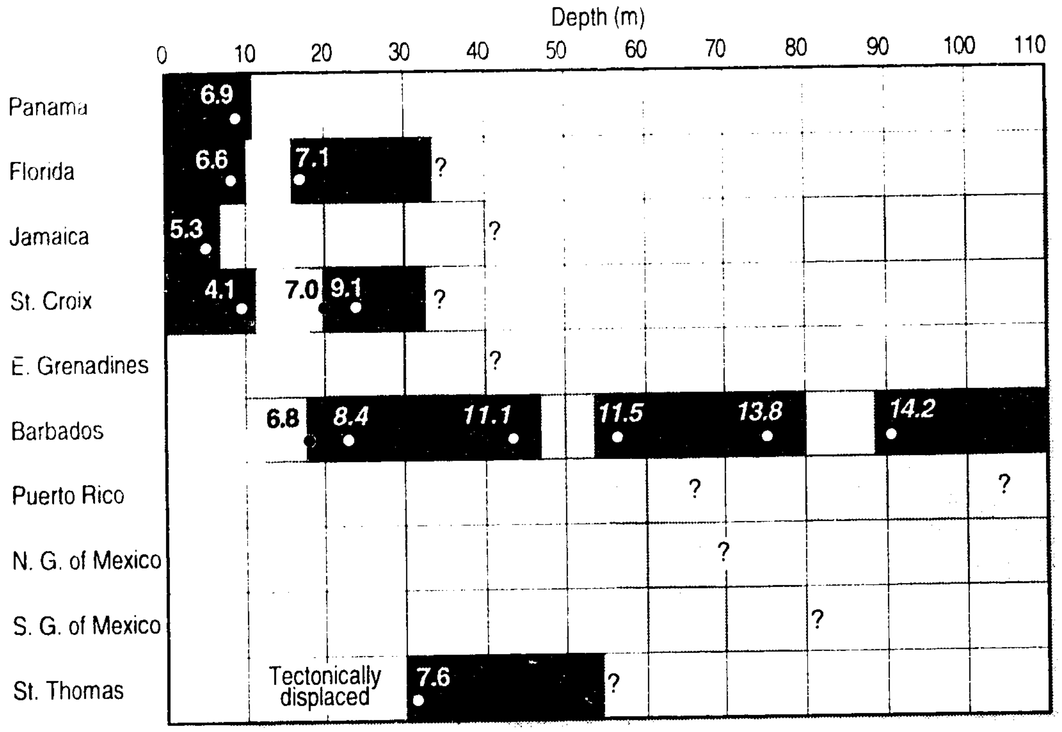

Figure 4.21. Reported positions and ages (ka) of drowned reefs in the Caribbean-Atlantic reef province. *Acropora palmata* framework shown by dark shade with dates in white; other coral and unknown framework shown by light shade with dates in black. Dates are radiocarbon except those in italics which are U/Th (TIMS). (Radiocarbon dates uncorrected for ocean reservoir effect, see Bard, 1988, or secular variation in atmospheric radiogenic carbon, see Bard et al. 1993). Note, reefs at ~18 m all ceased growing ~7 ka, with reef growth initiating at least 5 m upslope by ~6.9 ka. Panama—Macintyre & Glynn (1976). Florida—Lighty *et al.* (1978); Lidz et al. (1985). Jamaica—Goreau & Burke (1966); Land (1974). St. Croix—Adey et al. (1977); Macintyre and Adey (1990). East Grenadines—d'Anglejan & Mountjoy (1973). Barbados—Fairbanks (1989). Puerto Rico—Seiglie (1968). Northern Gulf of Mexico—Rezak (1977). Southern Gulf of Mexico—Bright (1977). St. Thomas—Holmes & Kindinger (1985).

Barbados, for example, is one such area that was tectonically active throughout the Pleistocene (Mesolella et al., 1969), with average uplift rates of between 25-45 cm ka^{-1} (Bender et al., 1979; Fairbanks, 1989). If terraces on this island formed during earlier Pleistocene stillstands, they would have been uplifted and should therefore occur at higher elevations than those on stable islands. Depth profiles across the Barbados shelf (Acker, 1986; Acker and Stearn, 1990), however, clearly identify terraces in positions coincident with stable islands such as Grand Cayman (Fig. 4.20). This correlation demonstrates that such shelf terraces are relatively young and must have formed during stillstands or slow-rise episodes associated with the last deglacial sea-level rise.

Constraining the timing of terrace formation on Grand Cayman any further is difficult because episodes of erosion cannot be dated directly. However, these episodes can be dated indirectly using constructional features associated with terraces. Reefs composed of *Acropora palmata*, for example, are ideal for establishing slow-rise or stillstand episodes because they are good indicators of sea-level, they are not subject to post-depositional compaction or transportation (Lighty et al., 1982), and they provide reliable radiometric dates (Edwards et al., 1987a; Edwards et al., 1987b). *Acropora* reefs which grew on the lower-shelf terrace and drowned when sea-level rose to its present position, have been investigated on the Florida shelf (Lighty et al., 1978; Lighty, 1985), off southern Barbados (Fairbanks, 1989) and around the Virgin Islands (Adey et al., 1977; Holmes and Kindinger, 1985). The reefs all started to grow ~11 ka ago at a depth of 30-40 m below present sea level (Fig. 4.21). As sea-level gradually rose, the rapidly-growing *Acropora* reefs tracked the rise until 7 ka, when it reached a stillstand at ~18 m below present sea level. Reefs established close to this stillstand position then suddenly stopped growing, and backstepped to new upslope positions (Fig. 4.21). This regional, and perhaps global, reef-die-off event at 7 ka was clearly related to the rapid sea-level-rise event that drowned the lower terrace, mid-shelf scarp, and -18.5 m intertidal notch on Grand Cayman and other islands (Fig. 4.20). Estimates of the rate and magnitude of this rise event from both the reef data and the drowned -18.5 m notch, independently suggest that sea level rose at least 5 m sometime between 7.1 and 6.9 ka ago (Fig. 4.21; see Panama, Florida and Barbados).

The dates of *Acropora* reef growth consequently demonstrate that the lower terrace and the mid-shelf scarp formed during an episode of slow sea-level rise from 11 to 7 ka (see Bard et al., 1990). This lower-terrace/mid-shelf scarp unit was then drowned by the extremely rapid 5 m sea-level-rise at ~7 ka. Shortly after this event (circa. 6.8 ka) sea-level slowed sufficiently to allow reefs to re-establish and marine erosion to re-initiate terrace cutting.

Discussion

In addition to the evidence provided by drowned reef and terrace positions, several morphological relationships support the interpretation that contemporary intertidal and subtidal erosion is responsible for the formation of terraces on the shelf around Grand Cayman. First, along the leeward and protected-windward margins, there is a positive correlation between small- and large-scale variations in shelf exposure and the aerial extent of the erosional ridge-and-furrow zone. In wind-facing areas the ridge-and-furrow zone extends across the upper terrace, but as the shelf becomes more sheltered these erosional features become increasingly restricted in aerial extent (Fig. 4.5). This suggests that such features are produced by, and are in equilibrium with, contemporary erosional processes. Second, a positive correlation also exists between shelf exposure and upper-terrace coral cover. This trend is directly related to changes in the aerial influence of the surf zone. Along wind-facing areas of the exposed-windward margin the fairweather surf-zone influences most of the upper terrace allowing surf-adapted corals, such as *Acropora palmata*, to colonize large areas, producing the spur-and-groove zone. In sheltered areas and along protected margins, however, the fairweather surf-zone has a much smaller influence, and surf-resistant coral growth is absent from most parts of the upper terrace. These barren areas remain uncolonized by other corals as a result of episodic erosion. During storms the upper terrace becomes an expansive surf-zone, and non-surf-adapted organisms that do initiate growth during fairweather are destroyed by abrasion.

This lack of coral growth, coupled with the ridge-and-furrow distribution trends, clearly demonstrates that contemporary marine erosion is the principle agent in the terrace-forming process. This conclusion is supported by two independent studies which document subtidal and intertidal erosion on the upper-terrace of Grand Cayman. On the leeward margin, Acker and Risk (1985) demonstrated that the boring sponge, *Cliona caribbaea*, caused up to 5 mm yr^{-1} of downwearing on the rocky pavement of the upper-terrace. As a result, they suggested that sponge boring alone could account for between 5-10 m of downwearing in a few thousand years (Acker and Risk, 1985). In another study along coastal areas of the exposed windward margin, Jones and Hunter (1992) documented the occurrence of large boulder-clusters with individual boulders weighing up to 40 tonnes. These boulders, quarried from the coastal cliff by storm waves, were either thrown onto the adjacent coastal area, or accumulated at the base of the cliff. Several of the onshore boulders were encrusted by modern corals (Jones and Hunter, 1992), confirming that undercutting and mass wasting of sea-cliffs is an active process on Grand Cayman.

The formation and origin of shelf-terraces in other areas has been attributed to constructional or erosional processes. By estimating how much time sea-level spent at a particular depth over the last 140 ka, Hopley (1982) proposed that shelf terraces on the Great Barrier Reef represented common levels of reef growth during interstadial highstands. This incremental-growth hypothesis explained why such terraces could not be correlated precisely; not all reefs grew to sea level at any one time (Hopley, 1982). In support of his hypothesis Hopley (1982) linked the formation of the shelf terraces to emergent reef terraces found in tectonically uplifted areas, such as Barbados and the Huon Peninsula (cf. Mesolella et al., 1969; Chappell, 1974). Uplift in these areas prevented incremental growth in the same position during successive sea-level rises, producing many small terraces. In stable areas, however, reef growth would be concentrated over a narrower depth range and old reef positions would become re-occupied during successive interstadials producing fewer, but wider terraces. A similar incremental explanation of shelf-terraces was suggested by Paulay and McEdward (1990), who simulated reef growth over the last 125 ka using computer modeling. They suggested that terraces developed by their model were incremental growth features resulting from the interaction between eustacy, subaerial erosion, subsidence, and reef growth.

The fundamental problem with the incremental-growth hypothesis is that it denies marine erosion a role in terrace formation. Hopley (1982) and Paulay and McEdward (1990) both dismissed marine erosion as insignificant in the terrace forming process. Hopley (1982, p. 165) stated: "As no stillstand of more than 5000 years is indicated during this [Glacial] period, the maximum width of intertidal erosional platforms that could be cut into carbonate substrate based on Trudgill's (1967) figures, ...is 35 m and probably much less." This statement reveals critical assumptions made by Hopley (1982) and Paulay and McEdward (1990). Both assumed low rates of contemporary marine erosion based on micro-erosion-meter measurements (from Trudgill, 1967 and Spencer, 1985). This instrument, although capable of precise measurements of substrate elevation, cannot record the long-term effects of wave quarrying (Trenhaile, 1987) which is the dominant erosional process along many, if not most, coastal areas (Trenhaile, 1980, 1989). Consequently, rates of coastal erosion measured with this instrument reflect fairweather processes and therefore underestimate both present day and time-averaged marine-planation rates. In addition, marine planation rates are likely to be non-linear through time as a result of temporal variations in climate and the reduced efficiency of erosion as terrace width, and bottom friction, increases (Trenhaile, 1980, 1987, 1989).

Mathematical modeling using more realistic erosion rates has demonstrated that erosional marine planation is capable of producing terraces in all but the most resistant rocks over relatively short periods of time (Trenhaile, 1989). During a series of modeling runs Trenhaile (1989) demonstrated that most simulated terraces had reached their equilibrium profiles in less than 5000 years, and in many cases little change had occurred after 2500 years. It is interesting to note that the width of terraces produced by this modeling (Trenhaile 1989, Table 9.2) are similar to the average widths of the terraces around Grand Cayman and other areas, such as the Bahamas (cf. Newell, 1961).

Rates of subtidal and intertidal erosion along Grand Cayman's rocky coasts have been measured at between 0.05 to 1.1 cm yr^{-1} (Spencer, 1985). These are considered to be significant underestimates of the rate of contemporary marine erosion for several reasons. First, the rates were taken over a six-month period using a micro-erosion meter (Spencer, 1985), and as a result they do not consider the effects of wave quarrying, known to be an active process on Grand Cayman (Jones and Hunter, 1992). Second, such a limited-duration study

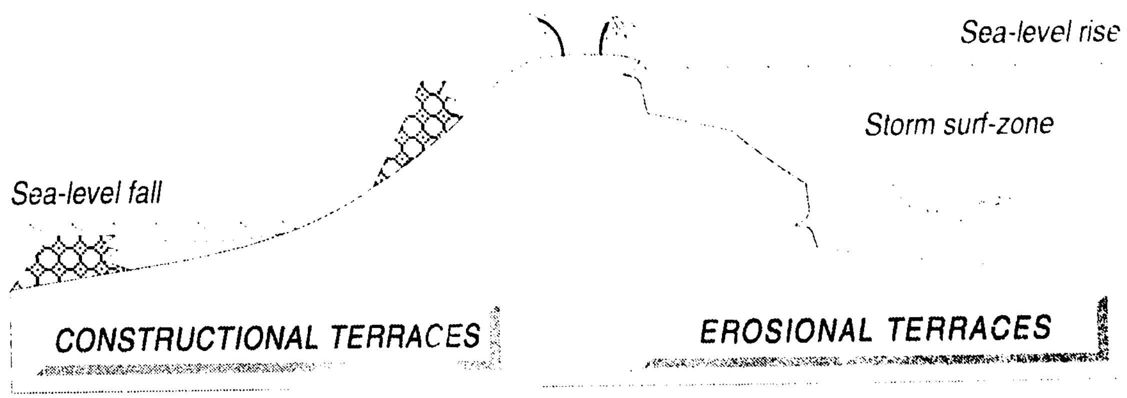

Figure 4.22. Erosional versus c... ...s. During slowly rising sea level marine abrasion during storms sculptures se... ...cliffs and notches into the island bedrock. During sea-level fall, terraces are formed ... ...ree... and the accumulation of lagoonal sediments. Note, erosional terraces ... sea... whereas con...tional terraces are horizontal or slope gently landward.

could not addre... ...so... ...l variations in erosion rat...s (e.g., seasonal storms), let alone longer-term variations due t... c...ma... change. By calcul...ating the average widths of the upper and lower terraces around G...nd ...ayman it is possible to determine the actual rate of marine erosion over the last 11 k... (when se... level ... ...ched the position of the lower terrace; Bard et al., 1990). Along rocky ...o...sts the planation rate is consistent for both the upper and lower terraces at ~4.0 cm yr^{-1}, whereas along beach-fronted coasts it is higher at ~7.0 cm yr^{-1}. In light of the mechanically weak nature of Grand Cayman's coastal limestones, and the abundant evidence of contemporary intertidal and subtidal erosion, marine planation rates of between 4 and 7 cm yr^{-1} seem quite reasonable, and compare favorably with long-term coastal erosion rates documented from other Pleistocene-limestone islands (cf. Cambers, 1988).

The above relationships clearly establish a firm role for contemporary marine erosion in the formation of shelf terraces around Grand Cayman. Hence, it is probable that erosion played a significant role in the formation of shelf terraces in other areas. An equally plausible suggestion is that shelf terraces formed by reef growth (Hopley, 1982). A close inspection of the characteristics of each type of terrace reveals some subtle differences that provide clues to the different origins of these features. Erosional shelf terraces are characterized by a gentle seaward gradient, whereas constructional, reef-formed terraces commonly have a landward sloping component (Fig. 4.22). From the depth-ranges shown in Fig. 4.20 it is apparent that most shelf terraces slope seawards and are therefore probably erosional.

We propose that the formation of erosional and constructional shelf terraces are related to separate terrace-forming processes at different stages of the sea-level cycle (Fig. 4.22). Erosional terraces are produced during sea-level rise as a result of *dynamic drowning*. This process occurs when the submergence of land areas is accompanied by significant marine erosion (in the opposite process — *passive drowning* — submergence is not associated with significant erosion). The efficiency of this type of marine erosion is directly related to the residence time of the storm surf-zone in any one position. During episodes of slow sea-level rise or stillstand, long residence times will allow the topography of the drowned land surface to be planated, and will produce a smooth seaward-sloping terrace. During more rapid rises, erosional modification by the storm surf-zone will be short lived, and relict topography may be preserved (although this depends on the magnitude of sea-level rise). Constructional terraces will not be produced during sea-level rise unless the drowning event is rapid enough and of sufficient magnitude to completely remove the reef from the optimum growth window (i.e. a sea-level rise of 20-30 m in a few hundred years!). Constructional terraces are much more likely to form during sea-level fall, because when emergent, they are isolated from additional coral growth or re-working by marine erosion.

If this hypothesis is correct, then erosional terraces cut during the last deglacial sea-level rise should be present at the same depth-ranges in stable areas. This information would be critical in confirming the rate and magnitude of rapid sea-level rise events recently documented by the positions of drowned reefs (Fairbanks, 1989; Bard et al., 1990; Chappell and Polach, 1991; Edwards et al., 1993; Blanchon and Shaw, 1993b). Unfortunately, attempts at correlating shelf terraces are hindered by three factors; neotectonic activity, modern sedimentation, and terrace description techniques (e.g., the practice of assigning a single depth value to a sloping surface). These sources of error easily account for the variation in terrace positions reported from the Great Barrier Reef (cf. Hopley, 1982) and the western Atlantic

(Fig. 4.20). Thus, to successfully correlate erosional terraces from different parts of the world, future work will need to employ high-resolution seismic-profiling techniques in areas where the tectonic history is well constrained.

Although much work on identifying terrace position [illegible] needs to be done, the terraces around Grand Cayman demonstrate that Holocene sea-level rise had a stepped geometry with long episodes of slow rise or stillstand being punctuated by a rapid (<200 yrs.), metre-scale rise event. This stepped pattern is consistent with pre-Holocene records of sea-level rise (Fairbanks, 1989; Bard et al., 1990; Chappell and Polach, 1991; Edwards et al., 1993) which also show slow-rise episodes punctuated by rapid, metre-scale rise events. While slow-rises or stillstands have been postulated by other sea-level investigators (e.g., Mörner, 1971), catastrophic-rise events are a relatively new concept (Blanchon and Shaw, 1993a, 1993b). If the 7 ka catastrophic-rise event recognized in the Caribbean-Atlantic reef province can be confirmed from other areas of the world, it would provide important clues to the processes involved in deglaciation and have profound implications for future sea-level rise (Blanchon and Shaw, 1993b).

Conclusions

1) Distribution of erosional ridges-and-furrows and coral growth on the upper bedrock terrace, and the presence of terraces at similar elevations around recently uplifted islands, demonstrates that shelf-terraces on Grand Cayman are erosional marine-planation surfaces cut by storm-wave abrasion during the last deglacial sea-level rise.

2) The coincident elevations of shelf terraces from other areas, and dates from associated *Acropora palmata* reefs, suggests that the lower-shelf terrace (20-40 m below msl) on Grand Cayman was cut during a slow sea-level rise episode from ~11 to 7 ka. The preservation of an -18.5 m notch, and the synchronous die-off and backstepping of *Acropora palmata* reefs, strongly suggest that an extremely rapid (<200 year) sea-level-rise event of at least 5 m in magnitude drowned the lower terrace about 7 ka ago. An-

other episode of slow sea-level rise ensued and produced the upper-shelf terrace (0-10 m below msl), which continues to form at the present day.

3) The cutting and drowning of shelf-terraces around Grand Cayman and other stable islands, resulted from the staircase geometry of Holocene glacio-eustatic sea-level rise. While slow-rise episodes or stillstands have been previously documented, rapid-rise events of the duration and magnitude demonstrated by this study are unprecedented. If such catastrophic-rise events can be confirmed from outside the Caribbean-Atlantic reef province, it will have profound implications for the mechanisms responsible for deglaciation (Blanchon and Shaw, 1993b).

4) The seaward-sloping nature of the shelf terraces and the landward-sloping nature of emergent reef-formed terraces, suggests there is a genetic relationship between the sea-level cycle and terrace-forming processes. During sea-level-rise episodes bedrock terraces are cut by erosional marine-planation, whereas during falling sea-level sedimentary terraces are formed by reef accretion.

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5

Fringing Reef around Grand Cayman: Coral Framework or Storm Rubble?

Introduction

Their ubiquitous distribution around tropical shores might suggest that fringing reefs are well studied and the controls on their development well understood. Yet relatively recent reviews of reef development have concluded that "...of the main types, fringing reefs are the simplest, apparently the least in need of complex explanations, and also the least studied." (Steers and Stoddart, 1977, p.27). Although lack of study has been true in the past, increased work over the last 20 years or so has shed much light on factors that influence fringing-reef development. In particular the physical and biological processes that govern coral growth and distribution have been widely documented (e.g., Goreau, 1959; Roberts et al., 1975; Geister, 1977; Adey and Burke, 1977; Done, 1983; Glynn, 1990). Also, the internal structure of fringing reefs has been investigated using submersible drilling units (Adey and Burke, 1976; Macintyre and Glynn, 1976; Easton and Olson, 1977; Shinn et al., 1977). The view that has emerged is that fringing reefs result from simple, in-place, coral growth controlled by fairweather wave energy and the location of antecedent terraces (Macintyre, 1988; Hubbard, 1988).

Yet these observations on fringing reef development have been made over limited human timescales and the influence of processes with longer recurrence intervals have been largely overlooked. Woodley (1992), for example, recently suggested that hurricane incidence is more important than previously recognized, and the classic view of reefs as being composed of luxuriant stands of coral may be one extreme of a variable condition. Woodley also suggested that over long time scales, hurricanes can be considered as a continuous force because no reef escapes their influence. Such suggestions are clearly important for reef

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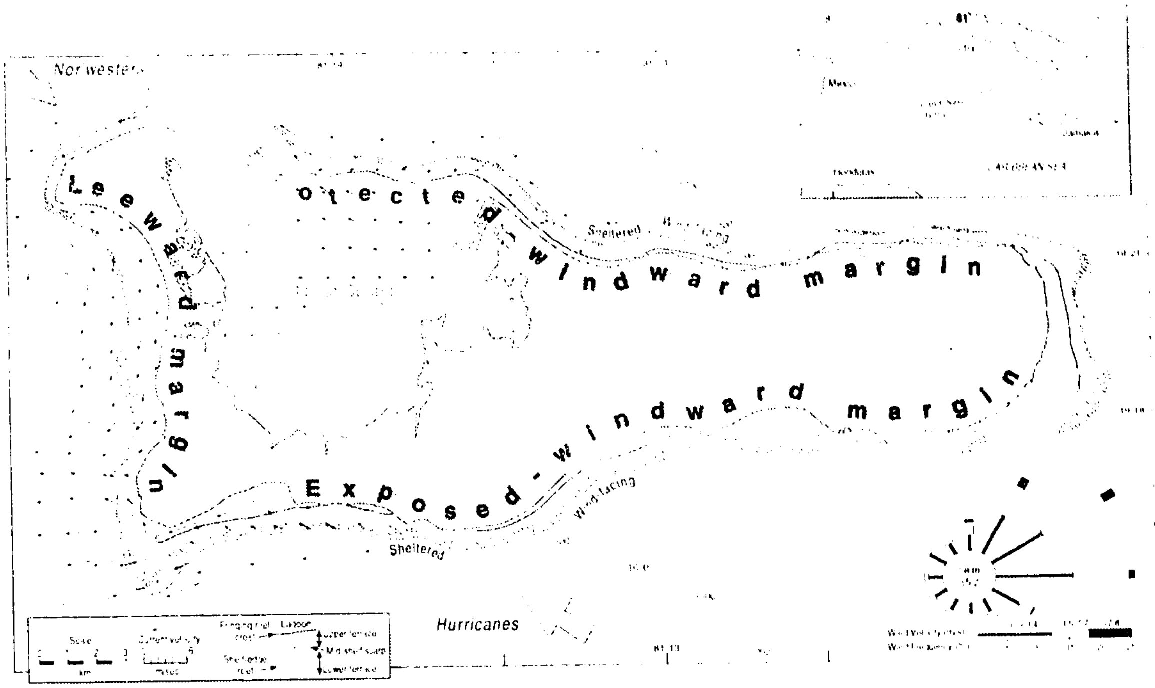


Figure 5.1. Grand Cayman. Location (inset), bathymetry in metres, wind and storm directions, surface currents, and details of the shelf, including energy classification of island margins, position of the mid-shelf scarp, and distribution of fringing and shelf-edge reefs (modified from Blanchon and Jones, 1995).

development and imply that the influence of fairweather processes has been overstated.

To assess the relative importance of fairweather versus storm processes, we examine the zonation, anatomy, configuration, and architecture of a fringing reef complex around Grand Cayman and attempt to establish a process-oriented model of reef development. By integrating sediment analysis, aerial and sonar profiles, cores and sections detailing internal anatomy, and previously published data on physical and biological processes, we show that the fringing reef around Grand Cayman consists of a ridge-like pile of coral rubble and sand that was deposited by hurricanes. Apart from a surficial cover, we found no in-place coral framework. We did, however, identify a previously undocumented relation between the reef configuration and shelf width which suggests that the initiation and architecture of fringing reefs result from an interplay between shelf gradient and sea-level rise. This interplay has important implications concerning reef development for it provides a mechanism by which fringing reefs can develop into barrier reefs and eventually into atolls during sea-level rise.

SETTING

Climate

Grand Cayman, located in the northwest Caribbean Sea between Jamaica and Cuba, is a small (197 km^2), low-lying (max. 17 m above msl), riverless island (Fig. 5.1). It enjoys a subhumid, tropical, and orographically unmodified ocean climate that is dominated by moisture-laden air masses of the North-East Trade Wind System. Like many other Caribbean islands subject to this system, its climate is distinctly seasonal (Burton, 1994). During the wet season (May to November), the island is subject to high temperatures (averaging ~29°C) frequent showers (averaging 4-8 mm/day), high humidity, and easterly or southeasterly winds (averaging 4-5 m sec^{-1}). During the dry season (December to April), tempe[illegible]es fall slightly (averaging 25°C), showers are less frequent (<3 mm/day), and winds move round to the east and north east (averaging 5-6 m sec^{-1}).

Cyclonic disturbances, which provide a large proportion of the annual rainfall, are common during both seasons. Tropical storms and hurricanes track east to northeast during the wet season, and storms associated with continental cold fronts track west to northwest during the dry season. The high historical frequency of severe storms affecting Grand Cayman suggests that they are the primary physical agent impacting the island's marine environments (Fig. 5.2). Although the destructive effects of such storms are widely documented, their long-term influence on tropical marine ecosystems is poorly known (Connell, 1978).

Marine Hydrology

Its micro-tidal setting means that large-scale oceanic currents and waves dominate fairweather water movement around Grand Cayman. Sheltered from high-latitude storm swells by islands of the Greater Antilles, the island's wave field is a product of the Northeast Trades and storm swells generated in the southwest Caribbean. Annual mean wave-power values hindcast for different sections of the coast show that at a depth of 10 m, east- and southeast-facing margins receive the highest and most enduring wave energy (~4 x 10^9 ergs/sec), north and northeast-facing margins receive large to moderate energies (~0.9 x 10^9 ergs/

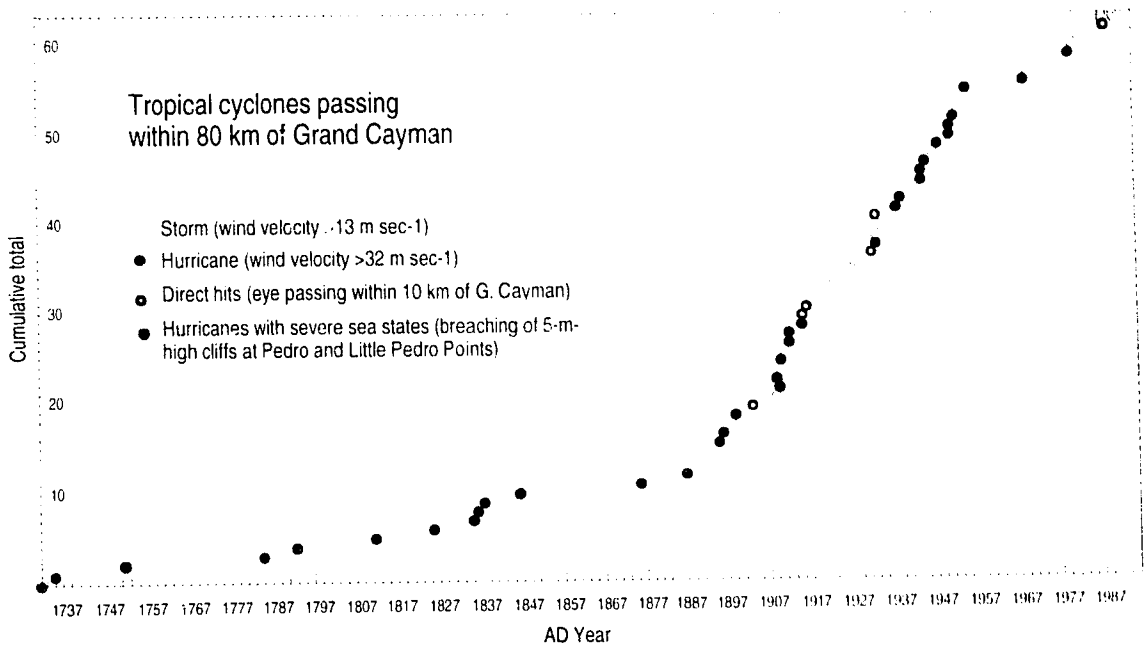


Figure 5.2. Historical record of tropical cyclones passing within 80 km of Grand Cayman. Hurricane recurence interval is ~7 years (38 hurricanes in 264 years). Recurrence of hurricanes passing within 10 km of the island is ~20 years (5 hurricanes in last 100 years), but recurrence time gaps range from 1 to 55 years. Since 1731, when records began, 4 hurricanes have breached the 5-m-high cliffs on the south side of the island between Great and Little Pedro Points (these cliffs are fronted by a deep, narrow, reefless shelf that allows deep-water waves to reach the shore). These large magnitude hurricanes have a 66-year recurrence interval with time gaps from 64 to 95 years. (Data compiled from Hirst, 1910; Williams, 1970; Clark, 1988; and Burton, 1994. Note, only hurricanes confirmed by at least two of these sources were used; prior to 1887 only hurricanes are recorded).

sec), and west-facing margins receive the least energy (~0.08 x 10^9 ergs/sec) (Roberts, 1974). This variation has been used to delineate three margin types (Fig. 5.1): a high energy *exposed-windward margin*, moderate energy *protected-windward margin*, and a low energy *leeward margin* (Blanchon and Jones, 1995).

Marine Environments

The marine shelf surrounding Grand Cayman is typically narrow, usually less than a kilometer wide, and slopes gradually from shore to the 20 m isobath where it is abruptly terminated by a vertical wall that forms the upper-island slope. The narrow shelf is characterized by two seaward-sloping terraces separated by a small scarp (Rigby and Roberts, 1976; Blanchon and Jones, 1995). These terraces are covered by a series of concentric coral-dominated environments that vary systematically as the orientation of the shelf, and its exposure to major ocean swells, changes (Fig. 5.3). This variation, as well as the configuration

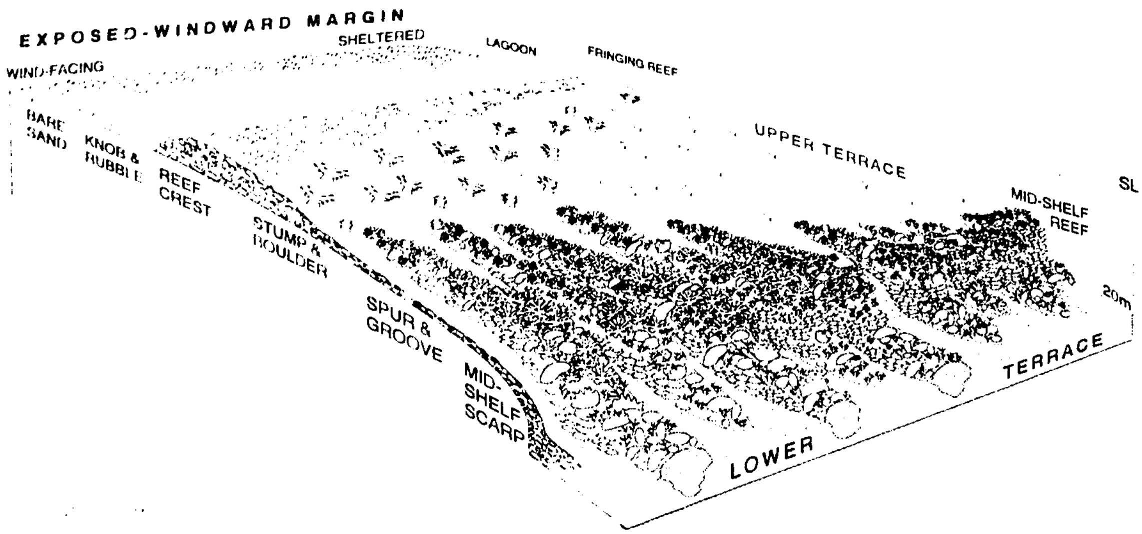


Figure 5.3. Schematic summary of Grand Cayman's shelf morphology and zonation showing how this varies as margin orientation changes. Note how, as margin becomes more sheltered, stump-and-boulder and reef-crest zones get narrower and spur-and-groove is replaced by ridge-and-furrow. For a more detailed de iption of these zones and terraces refer to Blanchon and Jones (1995).

of the terraces and scarp have been outlined recently by Blanchon and Jones (1995). Although details of the biota and environments have been well documented (Roberts, 1971; Rigby and Roberts, 1976; Raymont et al., 1976; Logan, 1981; Acker and Risk; 1985; Tongpenyai and Jones, 1991; Hunter, 1994; Kalbfleisch, 1995), little is known about the thickness or anatomy of deposits that underly these environments.

Methods

Five scuba transects were made across the fringing-reef complex on the south and east sides of the island (Fig. 5.4). Along each transect, sediment was collected and substrate character, including bedforms and biota, was recorded at 20 m intervals. Sediment thickness was determined using either a stainless steel push-probe or a drill-probe powered by compressed air and fitted with a masonry bit (Jones et al., 1992). Transect locations were estimated from triangulation on shore-based markers and plotted on aerial photographs. Where lagoon depths permitted, profiles were made using a shipboard depth sounder (accurate to within ±15 cm), and located using a Magellan GPS Nav 5000. In waters too shallow for the boat, profiles were completed using depth gauges on scuba.

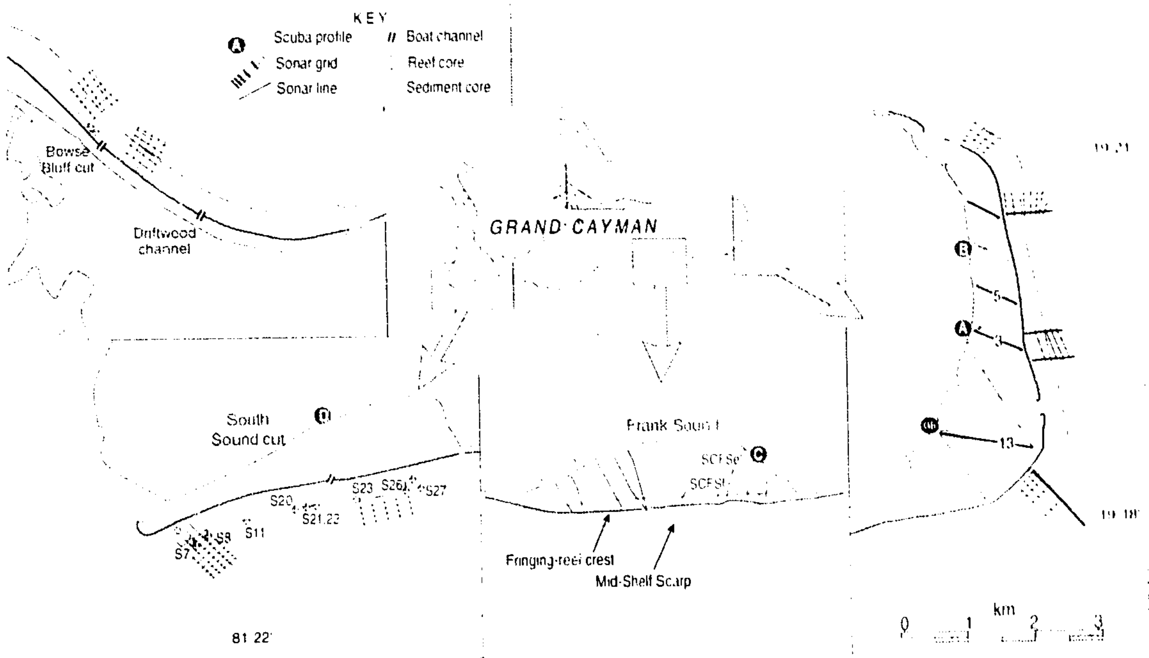


Figure 5.4. Transect map, showing location of scuba profiles, sonar profiles, boat channels, and cores. Reef cores are numbered after Natural Resources Unit dive-site mooring installations.

In addition to the transects, all breaks and man-made excavations through the reef crest were measured, logged, and sampled to determine reef anatomy (Fig. 5.4). These observations were supplemented by sediment and hard substrate cores from unbroken sections of the reef (Fig. 5.4). Sediment cores were collected by driving a 10-cm diameter PVC pipe into soft substrate using an air hammer sourced from an ordinary scuba tank (Jones et al., 1992). The longest core was 1.65 m long and recovery was only limited by the sediment thickness. Cores from hard substrates were obtained using a diver-operated hydraulic drilling system similar to the one described by Macintyre (1975, 1978). Back-reef zones and much of the reef crest were largely unconsolidated and so could not be cored, but cores were taken from reef-front zones in waters 2-12 m deep.

Sediment size-analysis at 1/4 phi (ϕ) intervals was conducted on 40 sediment samples from 4 out of the 5 transects using the procedures described by Folk (1974). All samples contained less than 5 wt. % silt and clay; this was removed by wet sieving samples, concentrated, dried, and weighed prior to sieving the remainder of the samples. Statistical parameters—graphical mean and standard deviation—were derived using the formulas of

Folk and Ward (1957) on the sand- and very-fine-pebble-sized fraction (-1.75 to 4.0 ϕ). Errors due to variations in sieve-screen openings and aggregation of sediment during drying were eliminated using tests described by Folk (1966). Sediments collected contained a variable weight fraction of gravel but statistical analysis of this fraction was not attempted because samples were too small for the gravel fraction to be statistically meaningful.

Sediment composition was determined on both the gravel and sand fractions. Analysis of the gravel, which consisted largely of cobble-sized fragments of coral, was made by random selection of 100 clasts across a 50-100 m^2 area of the reef surface and on the walls of the excavations cut through the reef. Each clast was split, coral genus identified, and the condition estimated (% bored). Analysis of the sand components was made using thin-sections and the point-counting method (e.g., Harwood, 1988).

FRINGING-REEF COMPLEX

Parallel to the coast along Grand Cayman's windward margin, less than 1 km offshore, is a well developed fringing-reef complex (Fig. 5.1). Along the exposed-windward margin, this complex covers the entire inner half of the shelf and merges with reef development on the small scarp that separates the upper- and lower-shelf terraces. But along the protected-windward margin, the aerial extent of the fringing-reef complex is reduced, covering less than a third of the inner shelf (Fig. 5.3).

Zonation

Transects and aerial reconnaissance show that the fringing-reef complex consists of 5 shore-parallel zones centered about the reef crest (Fig. 5.3). From the shoreward side these are 1) the bare-sand zone, 2) the knob-and-rubble zone, 3) the reef-crest zone, 4) the stump-and-boulder zone and, 5) the spur-and-groove zone (Fig. 5.3). The bare-sand and knob-and-rubble zones largely consist of unconsolidated sediment that slope back into the lagoon, whereas the reef-crest, stump-and-boulder, and spur-and-groove zones, are seaward-sloping substrates dominated by a surface cover of crustose coralline algae and/or corals.

ZONE depth/width	CHARACTERISTICS	EQUIV. ZONES
	Boundary: spurs extend over mid-shelf scarp onto lower-terrace (spur-and-sand zone)	
SPUR & GROOVE ZONE (SGZ) 5-10 m bmsl 150-200 m *See Fig. 12*	Seaward-projecting coral spurs separated by cobble floored grooves. **Spurs**: topped by *Acropora palmata* and *Acropora cervicornis* thickets, sides reinforced by *Montastrea annularis*. **Grooves**: cobble to boulder-sized clasts (mean -8ϕ) of abraded *A. palmata*. **Variation**: spur-and-groove zone only developed along wind facing sections of the exposed-windward margin.	**Buttress Zone** (Goreau, 1959) **Spur & Groove** (Shinn, 1963).
	Boundary: stump-and-boulder passes into spur-and-groove along wind-facing exposed-windward margin, but in other areas there is a distinct edge to the deposit (see Fig. 5D).	
STUMP & BOULDER ZONE (SBZ) 0-5 m bmsl 10-100 m *See Figs. 5A, 5B, and 5C.*	Broad, seaward-sloping field of coral boulders colonized by sparse *A. palmata* colonies and stumps. **Stumps**: robust, surf oriented *A. palmata* colonies, broken *A. palmata* stumps, and thickets of *Millepora complanata*. **Boulders**: large cobble to boulder-sized clasts of abraded *A. palmata* fragments (mean -9ϕ), stabilized by a mm-thick crust of coralline algae. **Variation**: see Blanchon and Jones (1995).	**Lower *Palmata*** (Goreau, 1959).
	Boundary: reef-crest zone slopes [illegible]ard into stump-and-boulder zone.	
REEF-CREST ZONE (RCZ) 0-1.5 m bmsl 10-20 m *See Figs. 5E and 5F.*	Thickets of *A.* [illegible] ridge of stabilized coral rubble. **Thickets**: in [illegible] deep, robust *A. palmata* colonies (up to 1.5 m [illegible]story of *M. complanata, P. astreoides*, and near spher[illegible] *Diploria strigosa*. In waters [illegible] m, encrusting forms of *A. palmata, M. complanata*, and coralline algae dominate. **Rubble ridge**: cobble to boulder-sized clasts (mean -8ϕ) of abraded *A. palmata* stabilized by crust of coralline algae, particularly *Porolithon pachydermum*. **Variation**: exposed-windward margin has predominantly shallow rubble-dominated ridges, whereas protected-windward margin has deeper coral-dominated ridges.	**Upper *Palmata*** (Goreau, 1959). ***A. palmata, M. alcicornis, P. astreoides* and *Lithothamnium* communities** (Rigby & Roberts, 1976). **Coralline-coral-*Dictyota* pavement** (Macintyre *et al.*, 1987).
	Boundary: knob and rubble shallows into the reef crest zone.	
KNOB & RUBBLE ZONE (KRZ) 1-2 m bmsl 50-120 m *See Fig. 5G.*	Shoreward-sloping field of coral rubble sparsely colonised by coral knobs. **Knobs**: monospecific knobs of *M. annularis* dominate, but others include *Diploria* spp. *Siderastrea* spp. and *A. cervicornis*. Understory species of *Agaricia agaricites*. **Rubble**: cobble to boulder-sized clasts (mean -7ϕ) of abraded *A. palmata*, encrusted by *Homotrema rubrum, Carpentaria utricularis*, bryozoa, boring sponges, and coralline algae. Interstitial sand and pebbles. **Variation**: more A. cervicornis knobs on protected windward margin; rubble commonly stabilized by corallines on exposed-windward margin	***M. annularis* community** (Rigby & Roberts, 1976). ***Turbinaria-Sargassum* rubble and *Laurencia-Acanthophora* sand and gravel** (Macintyre *et al.*, 1987)
	Boundary: bare sand grade into knob and rubble zone.	
BARE-SAND ZONE (BSZ) 1-8 m bmsl 50-300 m *See Fig. 5H.*	Shoreward-sloping field of rippled and burrowed sand. **Infauna**: includes shrimp (*Callianassa major*), fish (*Malacanthus plumieri*), sea urchins (*Meoma ventricosa, Clypeaster* sp.), worms (*Arenicola* sp.), and pelecypods (*Tellina radiata*). **Surface biota**: conch (*Strombus gigas*), stingrays (*Dasyatis americana*), sea-grass (*Thalassia, Syringodium*), green algae (*Halimeda, Penicillus*). **Variation**: sand stabilized by sea grass on protected -windward margin, bare on exposed-windward.	**Barren sand sheets** (Rigby & Roberts, 1976). **Bare sand zone** (Macintyre *et al.*, 1987).
	Boundary: bare sand zone slopes (5-20°) into *Thallasia*-sand zone in lagoon.	

Table 5.1. Characteristics of fringing-reef zones around Grand Cayman (see Fig. 5.5 for views of each zone).

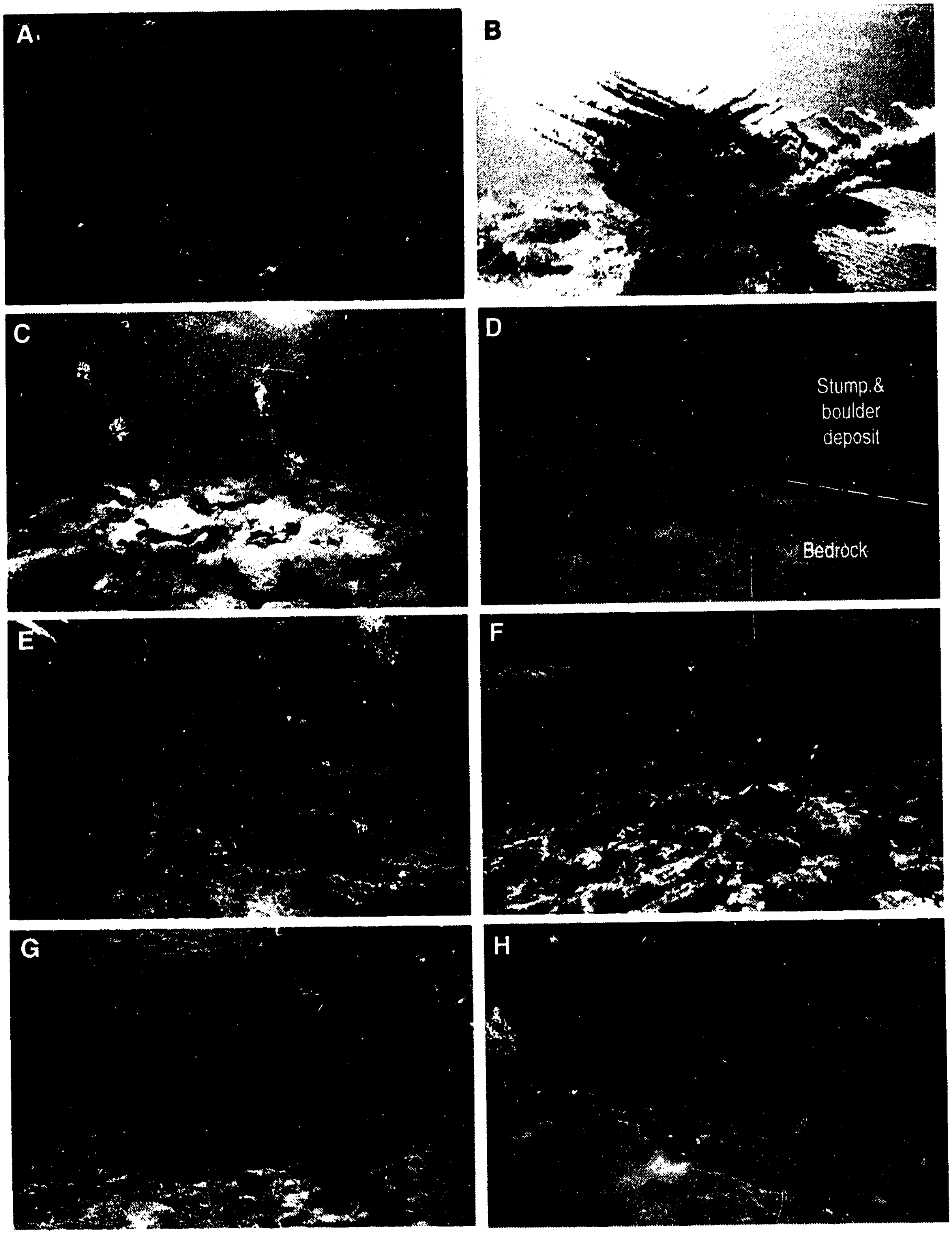


Figure 5.5. Zones of fringing-reef complex. (A), sparsely-distributed stumps of *A. palmata* (-3 m, transect B). (B), robust colony of *A. palmata* with surf-adapted form (-3 m, site S23). (C), large field of *A. palmata* boulders (-3.5 m, site N32). (D), seaward edge of stump-and-boulder zone with stabilized clasts overlying bedrock of upper terrace (-5 m, Old Man Bay). (E), reef-crest coral community dominated by *A. palmata* (-1.5 m, Grape Tree Point). (F), shallow reef-crest zone dominated by *A. palmata* cobbles (site: Driftwood Village). (G), coral knob (*M. annularis*) in knob-and-rubble zone (knob is 1.5 m high, site GB). (H) lagoonal edge of bare-sand zone with 20° slope into lagoon (-3 to -9 m, site GB).

Sedimentological and biological characteristics of the fringing-reef zones are detailed in Table 1 and illustrated in Fig. 5.5.

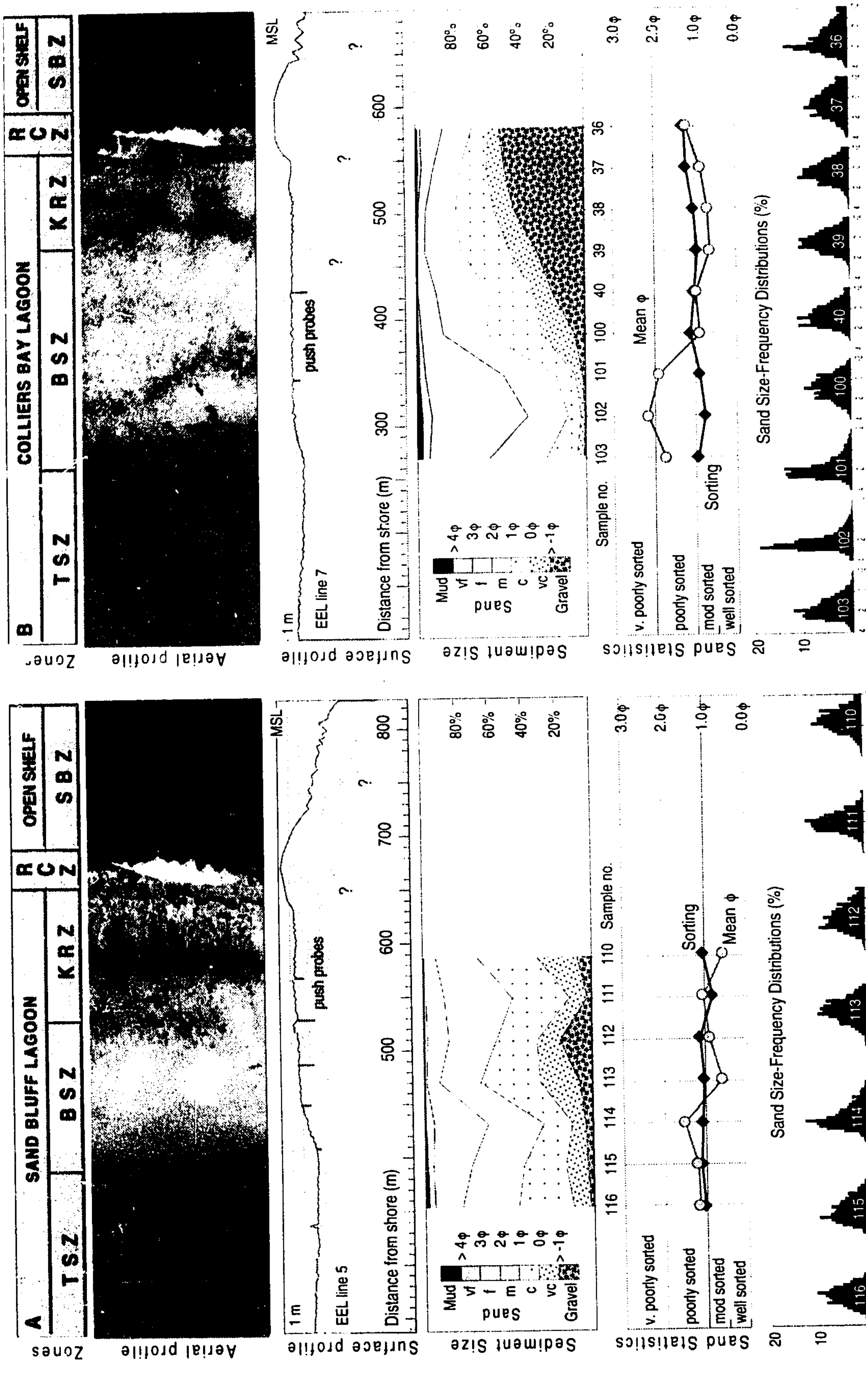
Sediment Analysis

As zones seaward of the reef crest usually lack loose sand and gravel, sediment analysis was only carried out in the knob-and-rubble and bare-sand zones. Gravel and sand fractions in these zones show a distinctive size decrease away from the reef crest (Fig. 5.6). In the knob-and-rubble zone close to the reef crest the gravel fraction constitutes >50 wt.% of the surface sediment and is medium-cobble in size, whereas the interstitial sand has a mean size of coarse sand (0 to 1ϕ). Some 50 m into the knob-and-rubble zone the amount of gravel abruptly decreases to <30 wt. %, and drops to medium-pebble size. By the time the bare-sand-zone is reached, the sand fraction dominates, having a mean size of coarse to medium sand (0 to 2ϕ), and a gravel content of <10 wt.%. Across the bare-sand-zone this gradually decreases to a medium to fine-sand (1 to 3ϕ) with <5% gravel (Fig. 5.6).

The sand fraction in all zones is typically poorly-sorted and the size distributions are almost invariably unimodal (Fig. 5.6). Only where sea-grass colonizes the bare-sand zone does the size distribution become polymodal (as in transect D; Fig. 5.6D).

Composition of cobbles on the reef crest is dominated by fragments of *A. palmata* (Fig. 5.7), but as the clast size decreases in a shoreward direction from the reef crest *A. cervicornis* becomes more dominant. Composition of sand-sized sediment is fairly uniform being dominated by coral detritus with subordinate foraminifera (Fig. 5.7). The only changes observed back from the reef crest is an increase in micritized grains in the bare-sand zone.

Figure 5.6. (pages 108-109) Transect profiles over the fringing-reef complex showing zonation, sediment thickness, and sediment size characteristics (TSZ: *Thalassia*-sand zone. BSZ, bare-sand zone; KRZ, knob-and-rubble zone; RCZ, reef-crest zone; SBZ, stump-and-boulder zone). (A) and (B) are transects across the East End lagoon showing typical decrease in sediment size into the lagoon and unimodal sand-size frequency distributions. (C) is the transect across Frank Sound showing similar profile and sediment characteristics (locations of the sediment cores are projected onto this profile). (D) shows the South Sound transect where sea grass has rapidly overgrown the bare-sand zone. This has produced a distinct increase in very-fine sand and mud, and has resulted in a typical bimodal sand-size frequency distribution. Despite this 'lagoonal overprint,' the decreasing size trend and poor sorting typical of other transects can still be recognized. (Note, boat channel through knob-and-rubble, reef-crest and stump-and-boulder zones; see Figure 4 for its location).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Sand Statistics
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Sorting
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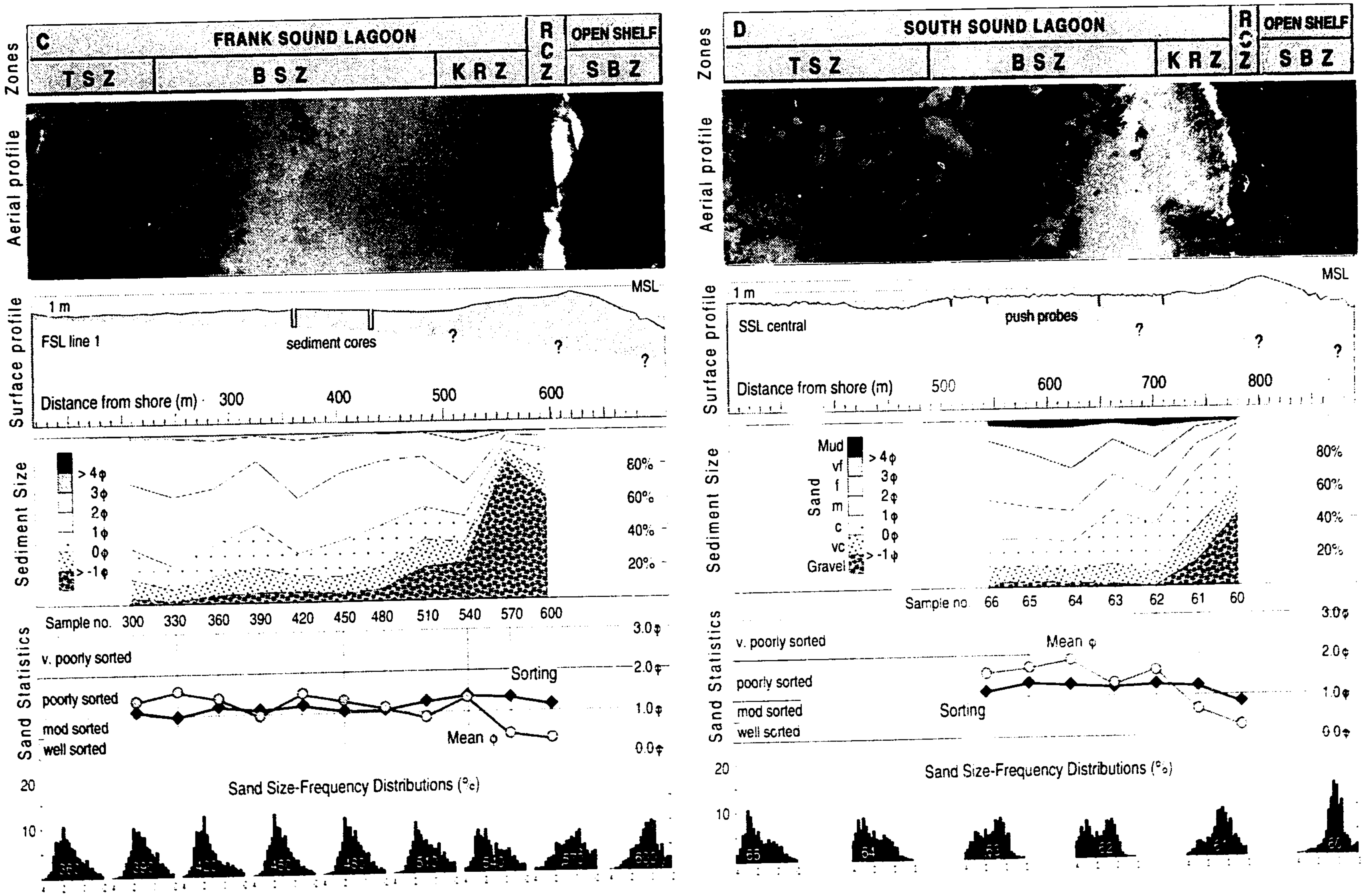

C
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Sand Statistics
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Sand Statistics
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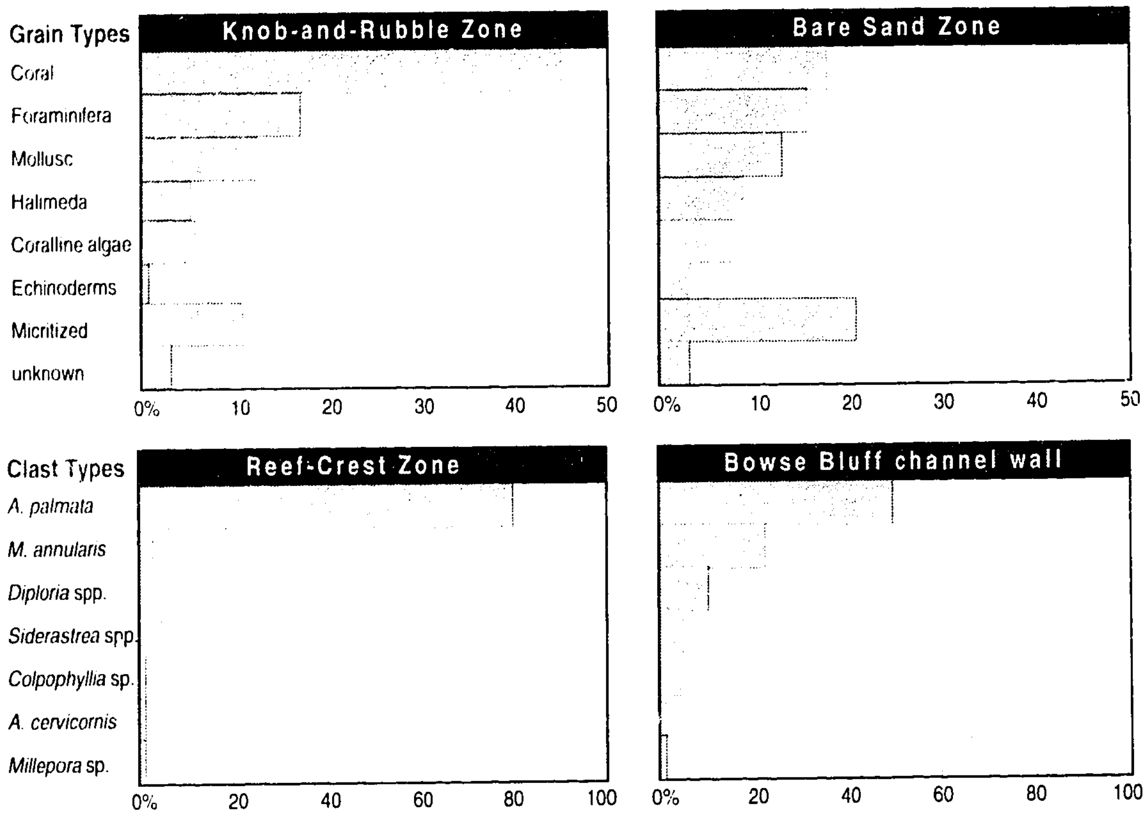

Figure 5.7. Composition of clasts and sand in different back-reef zones of fringing-reef complex (transect C) and from boat channel walls cut through reef (Bowse Bluff cut).

Anatomy

Examination of cores and submarine sections (boat channels) through the fringing-reef complex show that, with the exception of spur-and-groove, all zones consist exclusively of detrital deposits—in-place coral framework is conspicuous by its absence (Fig. 5.8).

Sediment probings show that the bare-sand zone is a wedge of sediment that thins from as much as 9 m near the reef crest (but more commonly 3-5 m) to <1 m along its shoreward edge. Cores up to ~1.5 m in length show the deposit is a structureless coral sand with a grainstone texture and rare pebble stringers (Fig. 5.8). The mean sediment size, sorting, and constituent abundances, are uniform down-core largely reflecting surface sediment characteristics.

Boat channels cut through the fringing reef complex provided vertical exposures of the upper 3 m of the knob-and-rubble, reef crest, and stump-and-boulder zones (Fig. 5.8 and

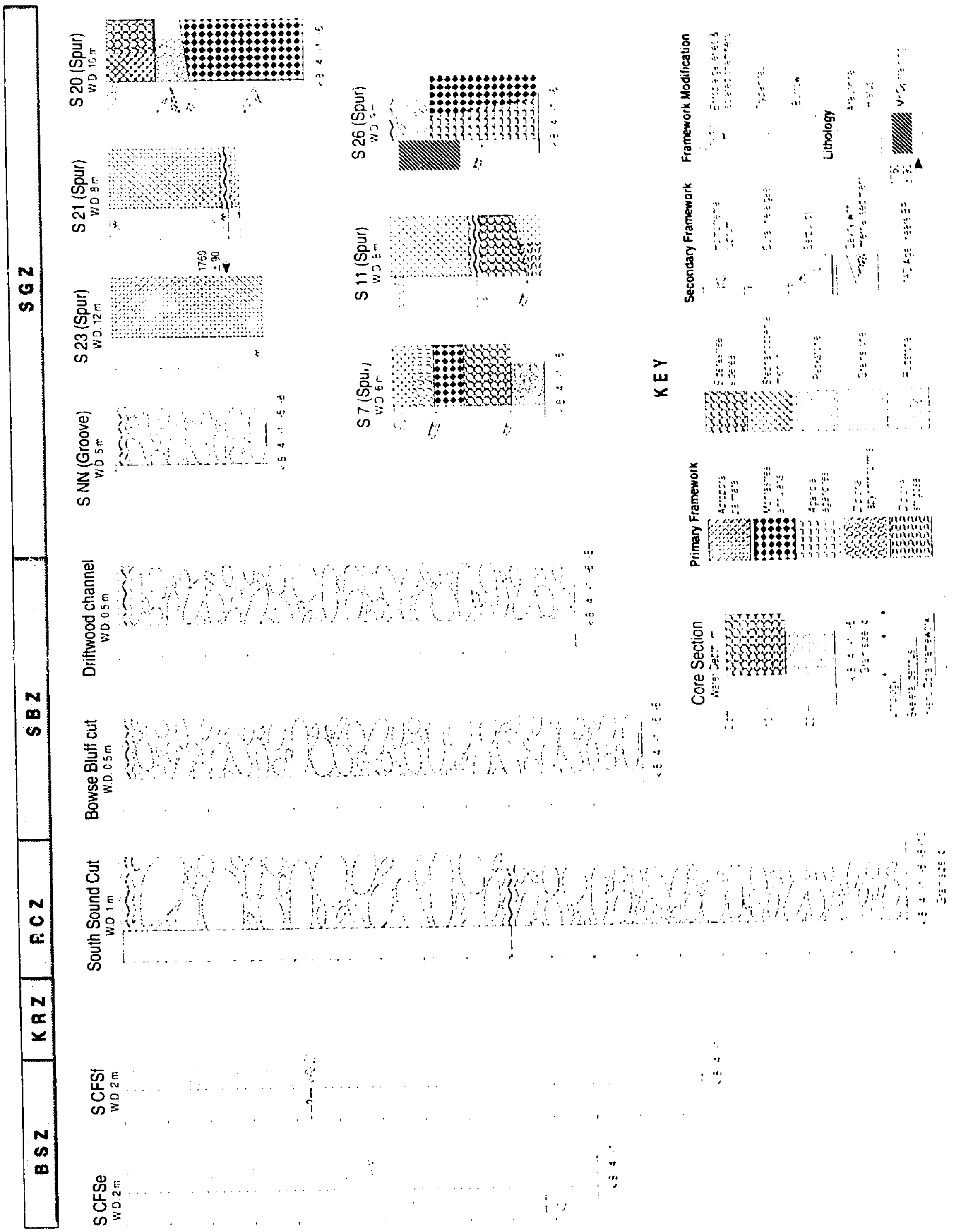

Figure 5.8. Boat-channel sections, drill cores, and sediment cores, from different zones of the friniging-reef complex around Grand Cayman (BSZ, bare-sand zone; KRZ, knob-and-rubble zone; RCZ, reef-crest zone; SBZ, stump-and-boulder zone; SGZ, spur-and-groove zone). Deposits underlying each zone are detrital and, except for spurs of spur-and-groove, no in-place cora. framework was encountered.

Figure 5.9. View of boat-channel section through the fringing-reef complex on the protected-windward margin. View is taken beneath the stump-and-boulder zone looking seaward (3 m below msl, site Bowse Bluff cut).

5.9). These exposures, without exception, show that all three zones are underlain by a coral-cobble conglomerate with a skeletal pebbly-sand matrix. From visual estimates along the exposures, clast size tends to increase in a seaward direction from small-medium cobble to medium cobble-small boulder size. Clast composition, however, is more consistent being dominated at all sites by *A. palmata* (e.g., Fig. 5.7A). In the deepest cut through the fringing-reef complex (shown on Fig. 5.6D), two distinct coral-cobble-conglomerate layers are exposed. Each layer is capped by a cm-thick crust of coralline algae, with the crust on the upper layer representing the present reef surface and the crust on the lower presumably representing a fossil surface. Clast size in each layer is consistent but at any particular site, the sizes of clasts in the upper layer are larger than clasts in the underlying layer. Both layers dip ~5° in a seaward direction from the reef crest.

Short cores collected from reef-front zones away from boat channels, confirm that the coral-cobble conglomerate deposit extends the full width of the stump-and-boulder zone

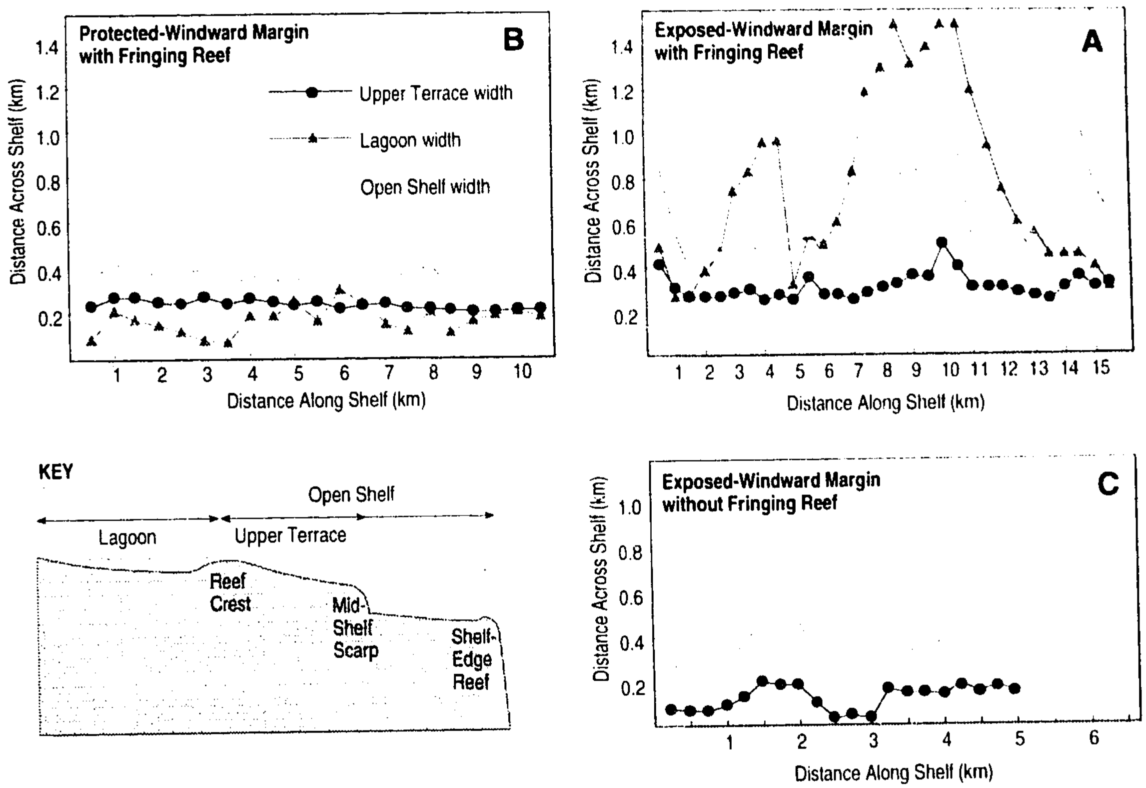

Figure 5.10. Bivariate plots showing variation in width of lagoons, open shelf, and upper terrace. (A) Variation along exposed-windward margin shows the distance from reef crest to mid-shelf scarp (upper terrace width) is relatively uniform at ~350 (±50) m when compared to lagoon or open shelf widths. (B) Along the protected-windward margin, the reef-crest to mid-shelf scarp distance is slightly less at 250 (±50) m. (C) In areas of the exposed-windward margin that lack a fringing reef, the distance from shore to mid-shelf scarp is ~200 m.

and even underlies the grooves of the spur-and-groove zone (Fig. 5.8). Cores into the spurs, however, generally encounter in-place coral growth frameworks dominated by heads of *M. annularis*, *Diploria* spp., *Siderastrea siderea*, and irregular stumps of *A. palmata*. Interstices between these corals are filled with *Millepora* pebble gravels with grainstone or packstone textures. All interstitial sediment in the cores from stump-and-boulder and spur-and-groove zones is cemented by bladed circumgranular crusts of Mg calcite.

Configuration and Architecture

The position of the reef crest compared to other features on the shelf shows some distinct relations (Fig. 5.10). Although it shows no correlation with lagoon or open-shelf widths, the reef crest is 300 (± 100) m from the mid-shelf scarp. This distance varies slightly on the different margins (Fig. 5.10). On the exposed-windward margin the reef crest is 350 (±50) m from the scarp, whereas on the protected-windward margin the distance is only 250 (±50) m. Along sections of the coast where the fringing-reef complex is absent, the distance from shore to the mid-shelf scarp is invariably less than 200 m (Fig. 5.10C). This previously undocumented relationship between reef position and upper-terrace width is also apparent from a re-examination of shelf profiles in other areas, such as the Belize Barrier Reef (Rutzler and Macintyre, 1982; Burke, 1982).

Depth and aerial profiles over the fringing-reef complex show it has two distinct architectural styles differentiated on the basis of lagoon depth, deposit thickness, and aerial extent. The first style has a 'fringing-type' architecture, characterized by a shallow lagoon (<5 m), limited deposit thickness (<5 m), wide back-reef zones (300-400 m), and a nearshore location (Fig. 5.11A). The second has a 'barrier-type' architecture, characterized by a deeper lagoon (>5 m), greater deposit thickness (>5 m), areally-restricted back-reef zones (100-200 m), and a location farther from shore (Fig. 5.11B).

Most of the reef around the island has a 'fringing-type' architecture. Drilling and probing demonstrate that the reef foundation is a gently seaward sloping bedrock terrace that can be traced from bare-rock patches in the lagoon to the edge of the stump-and-boulder zone, (Fig. 5.5D). The difference in bedrock elevation in these locations is <5 m. Consequently, as no bedrock was encountered in either the boat channels or cores, the fringing-type reef must be <5 m thick and underlain by a relatively flat seaward sloping bedrock surface.

The 'barrier-type' reef is only developed along one section of the exposed-windward margin (Fig. 5.11B). At that location, the lagoon is wide and reaches its maximum depth of ~10 m immediately shoreward of the shallow bare-sand zone. Drill probing demonstrates that the sediment cover over most of the lagoon is <2 m but abruptly increases to at least 9 m in the bare-sand zone (Fig. 5.11B). This is consistent with the suggestion that the underlying bedrock terrace slopes gradually seaward and passes beneath the reef without any significant break in slope, as it does in the 'fringing type'. If this assessment is correct, then the 'barrier type' fringing reef is at least 9 m and possibly as much as 12 m thick.

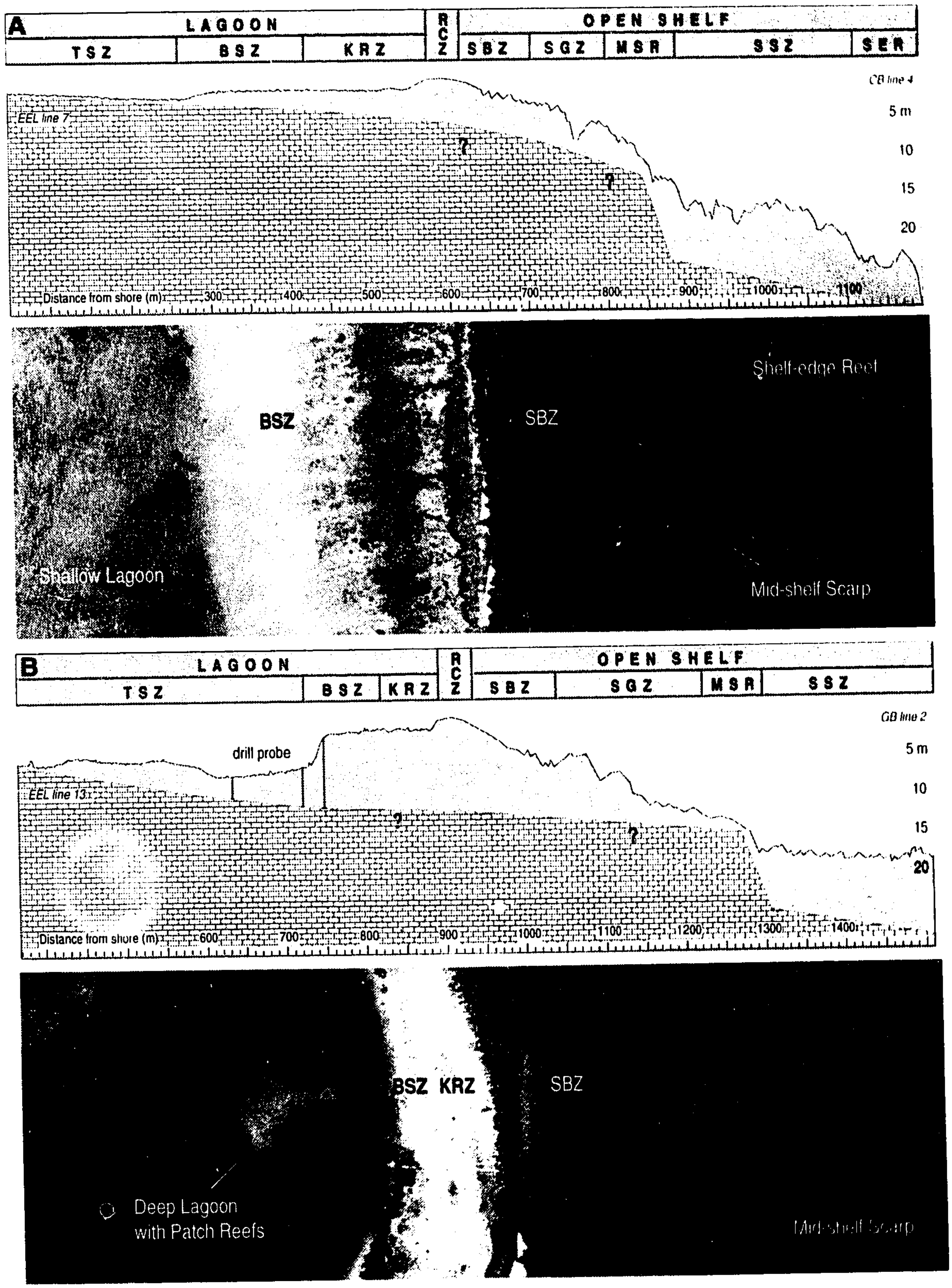

Figure 5.11. Aerial & depth profiles over exposed-windward shelf. (A), profiles over reef with fringing-type architecture (transect B). (B), profiles over reef with barrier-type architecture (transect GB). (TSZ, *Thalassia*-sand zone; BSZ, bare-sand zone; KRZ, knob-and-rubble zone; RCZ, reef-crest zone; SBZ, stump-and-boulder zone; SGZ, spur-and-groove zone; MSR, mid-shelf reef; SSZ, spur-and-sand zone; SER, shelf-edge reef).

Physical Processes

The influence of fairweather physical processes has been investigated on several fringing reefs in the Caribbean-Atlantic reef province, including Grand Cayman (Roberts et al., 1975; Suhayda and Roberts, 1977; Roberts and Suhayda, 1983; Roberts et al., 1988). Using current meters and wave-pressure sensors Roberts et al. (1988) showed that 'normal' fairweather waves overtopping a fringing-reef crest induced a bidirectional flow with a net lagoonal vector. Although mean values were <20 cm sec^{-1}, instantaneous bottom velocities in the reef-crest zone reached 180 cm sec^{-1} beneath breaking waves, and were capable of initiating movement of pebble-sized clasts (Roberts and Suhayda, 1983). Importantly, however, no sediment movement was documented over the fringing reef during fairweather, and currents in the fore-reef and back-reef zones were too weak to transport even sand-sized sediment (Roberts et al., 1975).

By comparison, storm and hurricane waves on fringing reefs are much more important for inducing sediment transport (e.g., Blumenstock, 1958; McKee, 1959; Stoddart, 1962; Maragos et al., 1973; Hernandez-Avila et al., 1977). On Grand Cayman, hurricanes have left an impressive record of sedimentation in coastal areas. Rigby and Roberts (1976) described an almost continuous ridge of sand, gravel and boulders in low coastal areas behind the fringing reef. Where the reef is close to shore, the coastal ridge consists of imbricated cobble to boulder-sized clasts of *A. palmata* deposited up to 6 m above msl (Lailey, 1976). Where the reef is farther offshore, the ridge is dominated by pebbly sand and has a consistent elevation of 3 m above msl (Lailey, 1976). Along exposed-windward coasts unprotected by fringing reefs, the cobble/sand ridge is largely absent and is replaced by spasely distributed clusters of very large boulders composed of the local coastal bedrock and encrusted with modern corals (Jones and Hunter, 1992).

Clearly, to transport such large boulders over coastal cliffs, and to deposit large sand and cobble ridges behind the island's fringing-reef complex, requires considerable wave energy. Hernandez-Avila et al. (1977) concluded that only hurricanes with breaker heights of greater than ~5 m were capable of destroying colonies of *A. palmata* and transporting their fragments onshore to produce boulder ramparts. Historical records suggest that at least 4 hurricanes with wave magnitudes >5 m have occurred since 1731 giving a recurrence interval for such major hurricanes of 66 years (Fig. 5.2).

Controls On Reef Zonation And Anatomy

Its rubble-dominated internal anatomy, surface-sediment distribution patterns, and associated coastal ramparts, indicate that the fringing-reef complex around Grand Cayman is generated and maintained by over-the-reef sediment transport during hurricanes. Most important are hurricanes with wave magnitudes >5 m because these are capable of destroying *A. palmata* coral associations in all zones seaward of, and including, the reef crest (Fig. 5.12). Coral fragments generated from these zones, together with sand and pebble-sized detritus from the lower terrace, are transported into the storm breaker zone and carried onto and over the reef crest by super-turbulent wave surf. Frictional interaction over these shallow substrates reduces the carrying capacity of the surf, and large clasts are dropped producing a rubble deposit that decreases in size from the stump-and-boulder to the knob-and-rubble zones. If back reef areas are shallow, turbulence and velocity is maintained longer, and the surf transports sand and pebble-sized sediment considerable distances into the lagoon. If the back-reef areas are deeper, however, turbulence and velocity are reduced and the surf only transports sediment a limited distance into the lagoon.

Besides explaining decreasing grain size, over-the-reef sediment transport during hurricanes also explains sorting, size-distribution, and compositional trends. Back-reef sediment is poorly sorted because hurricane-generated surf is highly turbulent and carries a broad spectrum of grain sizes that interact and interfere with the normal settling processes during transport. The sand-size frequency in back-reef zones is unimodal because this fraction represents the suspended load in equilibrium with physical processes rather than biological ones normally associated with lagoons. And the relatively uniform composition of sand in

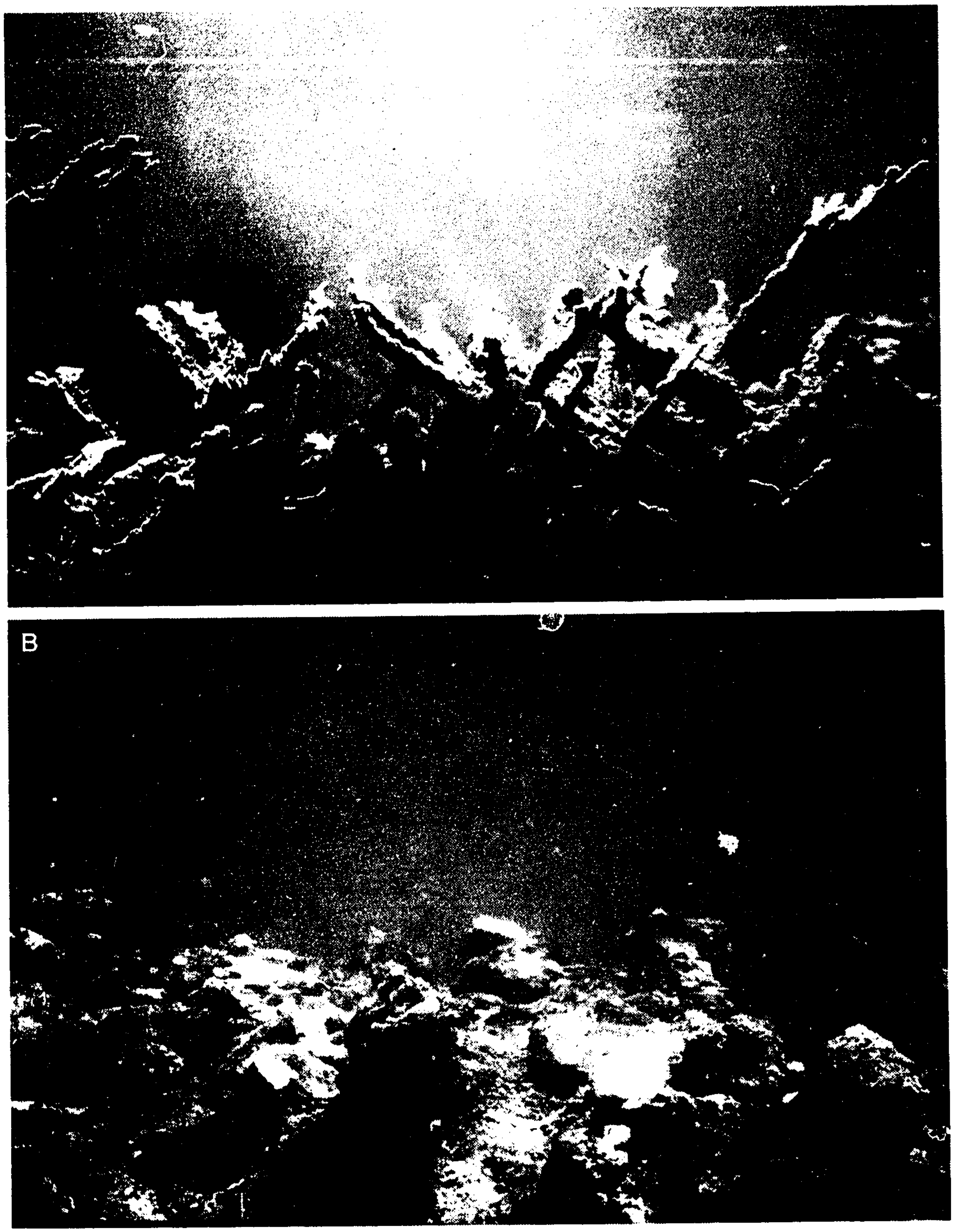

Figure 5.12. Before (A) and after (B) shots illustrating destructive effect of Hurricane Gilbert (1988) on spur-top *A. palmata* community of the spur-and-groove zone along the exposed-windward margin (~5 m below msl, site: S5). Triggerfish is ~40 cm long. Courtesy of Phil Bush.

back-reef-sediment zones reflects the single source of sediment from coral sands of the lower terrace, and the lack of sorting during transport.

Following such a hurricane, the rubble layer over the stump-and-boulder and reef-crest zones is quickly stabilized by crustose coralline algae, and new growth of *A. palmata* ensues. For a healthy *A. palmata*-dominated community, full recovery may take less than 50 years (Stoddart, 1974; Pearson, 1981). The 66 year recurrence interval of reef-destroying hurricanes on Grand Cayman is longer than this recovery interval and is consistent, therefore, with the suggestion that pre-hurricane conditions can be fully attained before the next major hurricane hits (Fig. 5.2). Even though lower intensity storms and hurricanes may recur before full recovery takes place, destruction will be less severe and localized as a result of variations in shelf orientation and angle of wave approach (Woodley et al., 1981). Consequently, individual sections of fringing reef may survive relatively unscathed for many years before being destroyed by larger hurricanes. Over several hundreds of years, therefore, the reef will undergo a cyclic pattern of destruction and renewal, producing a structure that is composed of successive layers of coral rubble, each stabilized by coralline algal crusts.

This hurricane model of reef anatomy and zonation contrasts with studies that have traditionally emphasized the dominance of in-place coral frameworks in reef complexes (e.g., Macintyre and Glynn, 1976; Adey and Burke, 1976; Easton and Olson, 1977). This emphasis is due to several factors. First, most descriptions of reefs have concentrated on the zonation and form of corals (e.g., Goreau, 1959) and have, therefore, tended to emphasize the fairweather processes that control this (e.g., Geister, 1977; Adey and Burke, 1977). Second, although the impact of destructive storms on reef communities has long been recognized (e.g., Stoddart, 1962; Connell, 1978; Woodley et al., 1981), the cumulative affect on reef anatomy over time has been overlooked and hence the importance of depositional processes largely ignored (Woodley, 1992). Even drilling investigations directly concerned with detailing reef anatomy have overemphasized the importance of in-place coral framework despite the widespread problem of poor recovery and the inherent limitations of using small diameter cores (e.g., Macintyre and Glynn, 1976; Adey and Burke, 1976; Easton and Olson, 1977). Yet, when investigators have had the opportunity to examine excavations into the reef crest they have invariably concluded that reef anatomy is detrital. Buddemeier et al. (1975), for example, used blasting to reveal the anatomy of the immediate fore-reef, reef-crest, and back-reef zones of the windward margin of Enewetak Atoll. They found that all three zones were underlain by a coral-conglomerate stabilized by a 15-cm thick surficial crust of coralline algae. From this they concluded that "..the present biological record on the atoll reef flat represents no more than a fleeting glimpse of how the reef flat grows and is destroyed through geological time." (Buddemeier et al., 1975, p. 1583).

Controls On Reef Configuration And Architecture

Previous work on reefs in general, and on fringing reefs in particular, has emphasized the role of coral growth or antecedent topography in determining reef configuration and architecture (Hopley, 1982; Braithwaite, 1987; Macintyre, 1988; Hubbard, 1988; Purdey and Bertram, 1993). Yet neither growth processes nor underlying bedrock features can reasonably account for the configuration and architecture of fringing reefs around Grand Cayman. This problem is clearly illustrated in attempting to explain the spatial relationship between the fringing-reef crest and the mid-shelf scarp (Fig. 5.10). The relatively uniform distance between these two features cannot be attributed to the interplay between coral growth and fairweather waves because much of the upper terrace along the protected windward margin is below fairweather wave base. Nor can it be explained by inferring initiation on an antecendent topographic ridge because sections through the reef show no core of bedrock.

The only physical relation between the reef crest and the mid-shelf scarp comes under storm conditions, when the width of the upper terrace becomes critical for the interaction between waves and coral detritus. As storm waves reach the abrupt break-in-slope at the edge of the upper terrace, they experience an abrupt increase in frictional attenuation due to the abrupt shallowing. This abrupt attenuation initiates spilling and causes storm waves to break shortly thereafter. For a wave of a given height, the position of the breaker point will be a constant distance from the spill point. Consequently, hurricane waves will deposit rubble a constant distance in from the edge of the upper terrace —a distance that will be proportional to their initial wave height. Thus, the average distance from the mid-shelf scarp to the reef crest along the exposed-windward margin is greater than along the protected-windward margin because the latter is impacted by smaller, fetch-limited, hurricane waves.

This interaction between hurricane waves and the upper terrace not only explains the presence of a fringing reef, but also its absence along certain sections of the shelf. Where the width of the upper terrace is less than this hurricane-transport distance (ie, ~200 m), no fringing reef will develop because the rubble will be either thrown ashore forming coastal cobble ramparts, or driven downslope by powerful currents deflected from coastal cliffs. Although this is true for all narrow parts of the upper terrace, discontinuities in fringing reef development along sheltered parts of the protected-windward margin also results from limited *A. palmata* growth (and therefore lack of sediment supply) induced by low fairweather wave energy. This is the reason why the leeward margin has no fringing reef.

If the key factor for the initiation of fringing reefs is shelf width, then the interplay between between sea-level rise and coastal gradient will control both the timing of reef initiation and the architectural style of subsequent reef development (Fig. 5.13). Where sea-level rises over a low-gradient shore, it will flood broad coastal tracts producing a wide, shallow, shelf. If this shelf is wider than the spill-point/breaker-point distance, coral debris will immediately begin to accumulate during hurricanes and a fringing-reef complex will develop. Where sea-level rises over a higher gradient shore it will only flood a narrow coastal tract producing a narrow shelf. Although coral growth will occur, the detritus produced during storms will be flushed off the narrow shelf and a fringing-reef complex will not develop until the shelf width exceeds the spill-point/breaker-point distance. This shelf-gradient hypothesis suggests, therefore, that the initiation and depth of fringing reef development will vary along the reef,

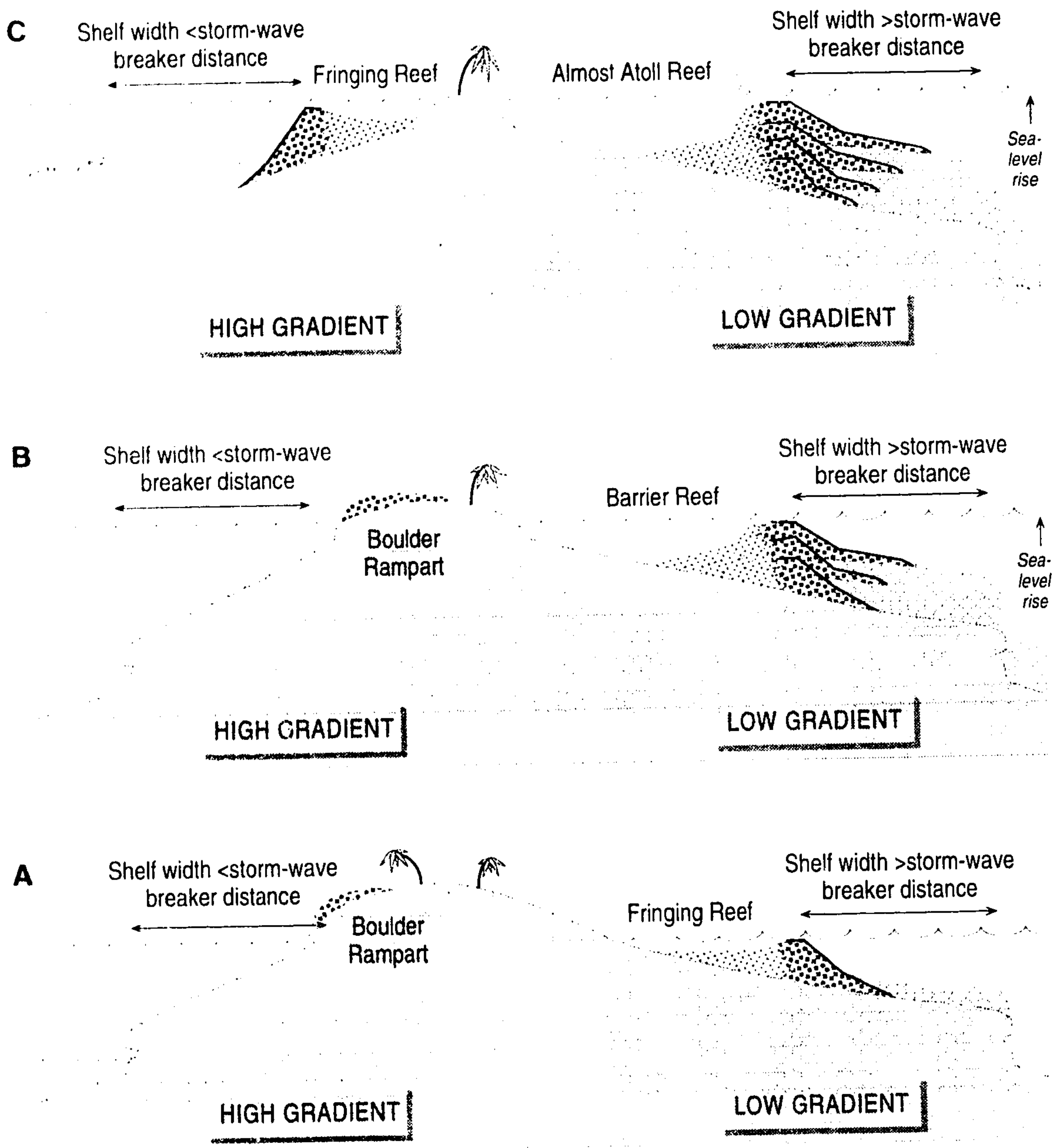

Figure 5.13. The effect of shelf gradient on configuration and architecture of fringing reefs during rising sea level. (A) Fringing reef initiates first where the shelf is wider than the hurricane spill-point/breaker-point distance. (B) As sea level rises these early reefs will accumulate vertically and the lagoon will increase in width and depth. This produces a barrier reef-type architecture. (C) As sea-level continues to rise, shelves that were previously too narrow to allow fringing reef development, will eventually attain the spill-point/breaker-point width threshold and narrow/shallow fringing reefs will develop.

being oldest and deepest along former low-gradient coasts and youngest and shallowest along former high-gradient coasts. This hypothesis is supported by a recent study from St. Croix which shows that the fringing-reef complex there initiated at a depth of ~10.5 m some

6 ka ago on a wide shelf, but only started growing along narrow shelves 1.5 ka ago at a depth of ~5 m (Burke et al., 1989). The two areas in question are only 5 km apart.

The shelf-gradient hypothesis also accounts for the variations in architectural style of the fringing reef around Grand Cayman. As sea-level rises, fringing reefs that form early on low-gradient shelves will keep pace, developing wider and deeper lagoons and thicker narrower profiles compared to those that form later on higher-gradient shelves (Fig. 5.13). Thus, early fringing reefs eventually develop into reefs with a barrier-type architecture. As they accrete, the frictional interaction of storm waves with the reef front decreases and the reef will start to retrograde over its back-reef deposits in an effort to maintain the spill-point/breaker-point distance. This retrogradation will not occur, however, if coral growth in the reef-front zones keeps pace with the sea-level rise, because the shelf will still be shallow enough to maintain the spill-point/breaker-point distance. Reefs that initiate later on higher-gradient shelves will have relatively shallow foundations, and will develop thinner, wider, profiles with narrower lagoons because they have had less time to accumulate during sea-level rise.

By highlighting this interaction between sea-level rise and hurricane-mediated fringing-reef development, we have identified a genetic succession between fringing and barrier reefs. With continued sea-level rise, and the complete inundation of Grand Cayman, the next step in this succession would probably be the development of an atoll (Fig. 5.13). This same sequence of reef development was first proposed by Darwin 150 years ago, but was suggested to be the result of simple upward coral growth during relative sea-level rise (Darwin, 1842). Perhaps if Darwin had witnessed the effects of a hurricane he might well have realized their importance for reef development.

Conclusions

We have shown that hurricanes are the dominant agent in the development of the fringing-reef complex around Grand Cayman. The surface cover of coral is merely a facade and hurricanes not only determine surface sediment character and distribution, but control reef anatomy, configuration and architecture as follows:

1) As large hurricane waves cross the shelf, they entrain sand from the lower terrace and uproot corals from reef-front zones. This detritus is transported into the storm surf zone and deposited as a poorly-sorted ridge of cobbles grading to sand over the reef. Between storms seaward parts of this rubble ridge are re-colonized and stabilized by surf-adapted crustose coralline algae and colonies of *A. palmata*. Recolonization is sufficiently rapid to allow complete reef regeneration before the next major hurricane. Consequently, the cyclic alternation of destruction, deposition, and regeneration produces a reef that is built-up of superimposed layers of hurricane-generated detritus, not by an in-place coral framework.

2) Hurricane waves also determine the configuration of fringing reefs. The height of hurricane waves control how far onto the upper terrace the detrital ridge is deposited. On margins that experience fully-developed seas this rubble ridge is deposited further across the terrace than in areas that are influenced by fetch-limited waves because of differences in power and carrying capacity. Where the terrace is narrow, a fringing reef does not develop because storm waves can transport detritus directly onshore and no rubble-ridge can accumulate.

3) During sea-level rise, the width of the shelf determines the timing of fringing-reef initiation and that, in turn, controls architecture. The reefs initiates early on wide shelves, and subsequently develops a narrower profile and deeper lagoon in a location further offshore than parts of the reef that develops later on narrower shelves. In a single reef system, therefore, older barrier-type reef sections can pass laterally into younger fringing-type sections as a result of variations in shelf gradient during sea-level rise.

4) The interplay between sea-level rise, shelf width, and hurricane mediated fringing reef development consequently produces a genetic succession from a fringing-type architecture to a barrier-type architecture and, with continued sea-level rise, could conceivably result

in an atoll-type architecture—the very sequence that Darwin's postulated nearly 150 years ago.

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6

Shelf-Edge-Reef Architecture Around Grand Cayman Controlled By Hurricanes

Introduction

The most obvious reefs on modern shelves are those that preferentially develop in the shallow turbulent waters of the surf-zone. Yet many studies have also documented the existence of substantial reef development on the edge of shelves in deeper, quieter water (Macintyre, 1967; 1972; Goreau and Goreau, 1973; Ott, 1975; Rigby and Roberts, 1976; James and Ginsburg, 1979; Burke, 1982; Hubbard et al., 1986). James and Ginsburg (1979), for example, described a submerged reef ridge on the edge of the Belize shelf rising from ~35 to 15 m below msl. They noted that it had a diverse coral cover with "a luxurience that equals that of the shallow water spurs and grooves above" (James and Ginsburg, 1979, p. 63), and concluded that this coral development was not simply a veneer but preferentially developed along the shelf edge to depths of 65 m. Subsequent drilling confirmed that this was the case and showed that the reef had undergone at least 12 m of vertical framework accretion over the last 5 ka (Macintyre et al., 1982).

Although luxuriant coral communities had been previously reported to depths of 70 m (Goreau and Wells, 1967), it was considered that significant reef accretion was restricted to depths less than 15-20 m due to the relatively slow coral growth below these depths (e.g., Macintyre, 1972). The discovery of such thick and luxuriant reefs in these supposedly deep and prohibitive shelf edge locations was therefore difficult to explain. Early attempts centred around the idea that shelf-edge reefs were immature parts of the 'main' barrier- or fringing-reef system that had not been able to keep up during sea-level rise (Macintyre, 1972; Goreau and Goreau, 1973; Adey and Burke, 1976; Hubbard, 1988). Prior to drilling, it was thought that they were originally barrier reefs that established on shelf edges during a lower sea

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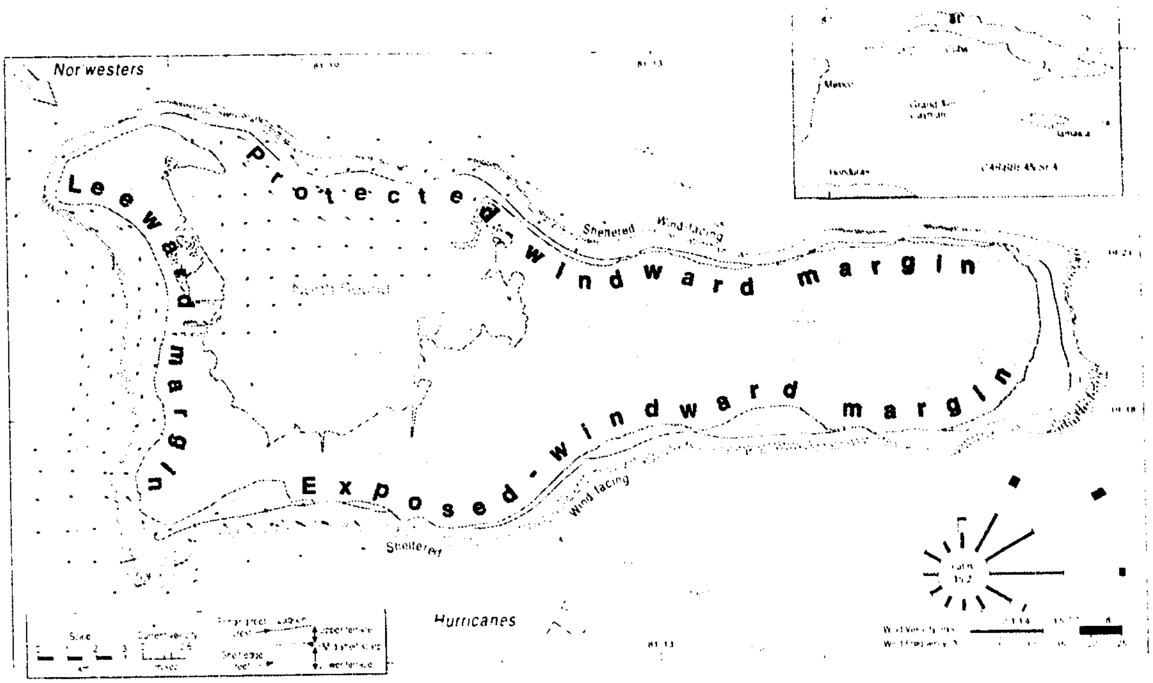

Figure 6.1. Grand Cayman. Location, bathymetry in metres (LeBlanc, 1979), wind and storm directions, surface currents (Darbyshire et al., 1976), and details of shelf, including energy classification of island margins, distribution of shelf-edge reef, and position of mid-shelf scarp and fringing reefs.

level and their architecture was therefore controlled by shallow fairweather waves. During sea-level rise, the reefs were unable to keep up due to their slow growth rates, and became stranded in deeper water where they continued to develop under a lower wave-stress regime (Roberts et al., 1975). Drilling showed, however, that shelf-edge reefs owed much of their relief to accretion of deeper-water corals. And even though some had nucleated upon relict shallow-water reefs (Adey et al., 1977; Hubbard et al., 1986), others were found that apparently had no such foundation (Macintyre et al., 1982; Hubbard, 1989).

To a large extent, early investigations of shelf-edge reefs have been somewhat parochial and localized in scope. There has, as yet, been no systematic analysis of their architectural variation and little consideration of the processes responsible. In this study we focus on shelf-edge reef (SER) development around Grand Cayman (Fig. 6.1). We present the first architectural analysis, describing variation in morphology, coral zonation, and structure around the island. We find that architecture varies in response to the changing intensity of hurricane-generated waves and currents as margin orientation changes. By considering the longer term geological effects of these processes, we propose a preliminary model of shelf-edge reef accretion and suggest that modern shelf-edge reefs are not in the process of catching up with sea level, as previously thought, but are actively accreting laterally in a way that appears similar to many ancient shelf-edge reefs.

Setting

Climate

Grand Cayman, located in the northwest Caribbean Sea between Jamaica and Cuba, is a small (197 km^2), low-lying (max. 17 m above msl), riverless island (Fig. 6.1). It enjoys a subhumid, tropical, and orographically unmodified ocean climate that is dominated by moisture-laden air masses of the North-East Trade Wind System. Like many other Caribbean islands subject to this system, its climate is distinctly seasonal (Burton, 1994). During the wet season (May to November), the island is subject to hot temperatures (averaging ~29°C) frequent showers (averaging 4-8 mm/day), high humidity, and easterly or southeasterly winds (averaging 4-5 m s^{-1}). During the dry season (December to April), temperatures fall slightly (averaging 25°C), showers are less frequent (<3 mm/day), and winds move round to the east and north east (averaging 5-6 m s^{-1}).

Cyclonic disturbances, which provide a large proportion of the annual rainfall, are common during both seasons. Tropical storms and hurricanes track east to northeast during the wet season, and storms associated with continental cold fronts track west to northwest during the dry season. Rainfall from these storms drains centripetally through extensive mangrove swamps towards the low western shores of the island (Giglioli, 1976). There, percolation into the ocean takes place through the permeable bedrock and beach-ridges during periods of lower monthly tides (Giglioli, 1976). This process filters out suspended solids and is responsible in large part for the renowned clarity of the sea surrounding the island.

Marine Hydrology

Its small size and micro-tidal setting, mean that large-scale oceanic currents and waves play the dominant role in fairweather water movement around Grand Cayman. Sheltered from high-latitude storm swells by islands of the Greater Antilles, the island's wave field is a product of the Northeast Trades and storm swells generated in the southwest Caribbean. East- and southeast-facing margins receive the highest and most enduring wave energy (~4 x 10^9 ergs/sec), north and northeast-facing margins receive large to moderate energies (~0.9 x 10^9 ergs/sec), and west-facing margins receive the least energy (~0.08 x 10^9 ergs/sec) (Roberts, 1974). This variation delineates three margin types: a high energy *exposed-windward margin* (east and south coasts), a moderate energy *protected-windward margin*, (north coast), and a low energy *leeward margin* (west coast) (Blanchon and Jones, 1995). These margin types are further divided according to local changes in shelf orientation; east-facing sections are described as *wind-facing* and west-facing sections as *sheltered* (Fig. 6.1).

The main ocean current affecting Grand Cayman is a west flowing tributary of the Caribbean Current that enters through the Windward Passage between Cuba and Haiti. By the time it reaches Grand Cayman this tributary is a moderately strong unidirectional flow that averages 30 cm s^{-1} and can be detected down to ~300 m (Darbyshire et al., 1976). Fixed, continuously-recording current meters deployed on the shelf flanking the southwest coast demonstrate that the current is a tidally-enhanced drift with a coherent but lagged response to wind stress (Roberts et al., 1975). It has a weak but distinct periodicity which corresponds to the tidal cycle, reaching 50 cm s^{-1} in mid cycle and slackening thereafter. Current reversing is generally suppressed by the drift component except during enhanced biweekly lunar tides. As the drift current passes around the island it deflects up onto the shelf and suffers a 70% reduction in velocity by the time it reaches inner parts of the reef dominated shelf (Roberts et al., 1975).

Marine Shelf Zonation

The marine shelf surrounding Grand Cayman is usually less than a kilometer wide, and slopes gradually from shore to the 20 m isobath where it is abruptly terminated by a vertical wall that forms the upper-island slope. The narrow shelf is characterized by two seaward-sloping terraces separated by a small scarp (Rigby and Roberts, 1976; Blanchon and Jones, 1995). The upper terrace, which slopes from 0-10 m below msl, is dominated along the exposed-windward margin by zones associated with the fringing reef (stump-and-boulder and spur-and-groove zones of Blanchon and Jones, 1995). Along all other margins, however, fringing reef development is less extensive and the upper terrace consists of a barren bedrock pavement traversed by erosional furrows (ridge-and-furrow zone of Blanchon and Jones, 1995). The upper terrace terminates at a mid-shelf scarp (10-15 m below msl), which is commonly obscurred by active coral growth and sediment accumulation.

In all areas, the lower terrace, which slopes from ~15-30 m below msl, is a zone of active coral growth and sediment accumulation associated with a shelf-edge reef (spur-and-sand and shelf-edge-reef zones of Blanchon and Jones, 1995). These deposits are underlain by a bedrock terrace that slopes seawards from the base of the mid-shelf scarp to ~40 m below msl. This bedrock terrace and mid-shelf scarp, which are geomorphic equivalents of the upper terrace and coastal cliff, formed during a stillstand in the Holocene sea-level rise at 7.6 ka (Blanchon and Jones, 1995; Blanchon and Shaw, 1995). The lower terrace is terminated by a sub-vertical escarpment that extends from the shelf break at 30-60 m, to upper parts of the island slope at 115-145 m below msl (Rigby and Roberts, 1976; Messing, 1987).

Methods

Grand Cayman is ideal for an analysis of shelf-edge reef architecture because: (1), its terraced shelf and environments are representative of many islands in the Caribbean-Atlantic reef province (Land and Moore, 1977; James and Ginsburg, 1979; Grammer and Ginsburg, 1992); (2), the reef is developed around the full perimeter of the island and, (3), in most areas, it is accessible from shore. This allowed us to assemble physical and biological data from 16 transects on all sides of the island (Fig. 6.2). For each transect, aerial photographs (1:10,000 and 1:6,000) were used to calculate spur frequency, using the method of Roberts (1974), and aerial coverage of the SER across the shelf. Continuous 8-mm-video profiles were recorded vertically over the reef with scuba to a depth of 40 m below msl. The video

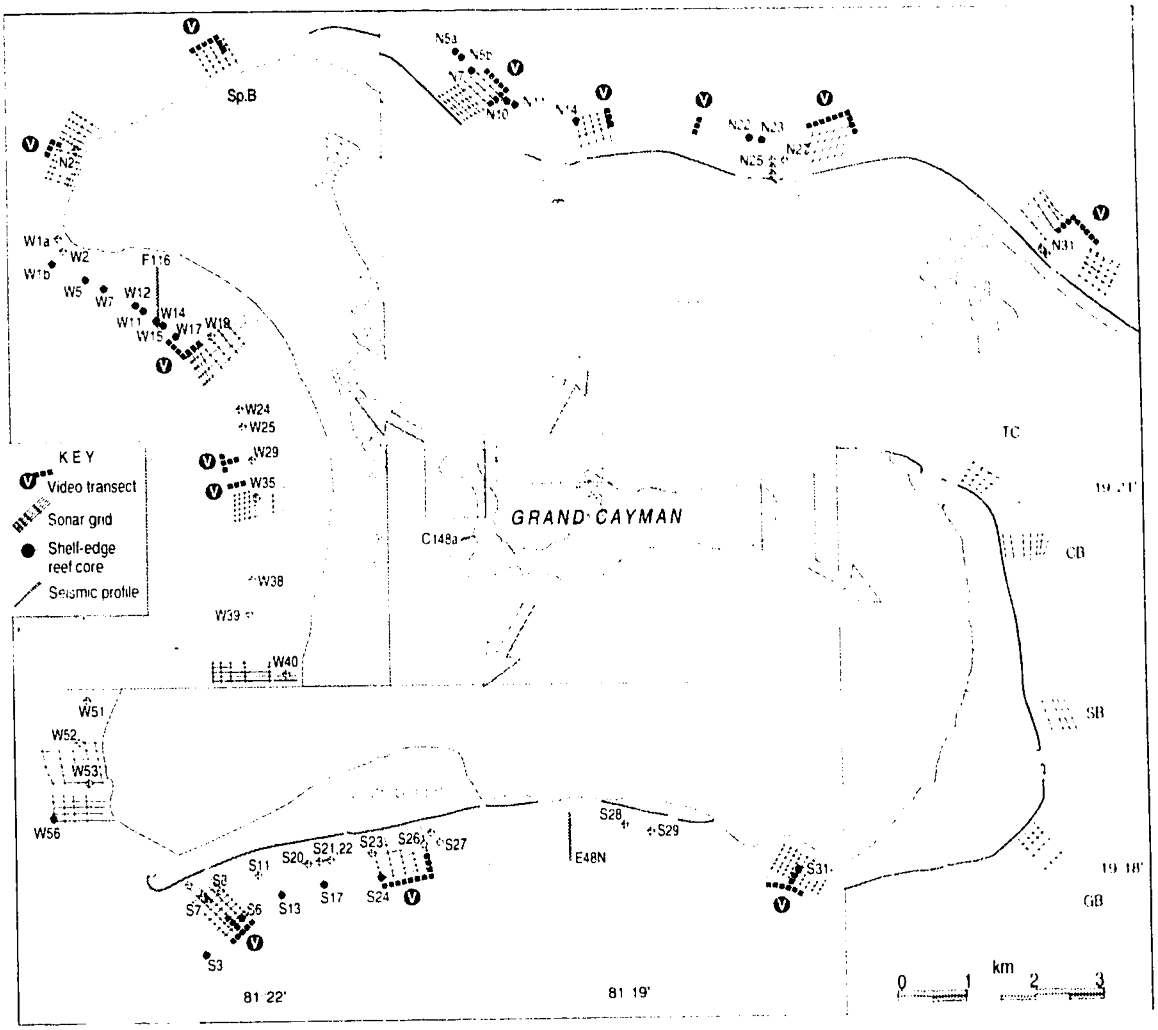

Figure 6.2. Transect map, showing location of video profiles, sonar profile grids, selected seismic lines, and shelf-edge-reef cores. Marine site and cores are numbered after Natural Resouses Unit dive-site mooring intallations (see Apendix A for coordinates). Where moorings are absent, marine sites and cores are named after prominent land marks.

camera, equipped with a wide-angle lens, was held ~2 m from the reef surface in order to give a constant frame area of 2 m^2. Estimates of coral coverage and form abundance were made, to the nearest 5%, from a frame-by-frame analysis of each video transect. Reef morphology was assessed qualitatively from video transects taken horizontally along the seaward side of the reef for a distance of ~750 m, and quantitatively using depth-sounding profile grids obtained over each transect (Blanchon and Jones, 1995). For each grid, depth measurements from profiles (accurate to within ±0.15 m) were averaged for individual SER features; this was more accurate than arbitrarily choosing a 'representative' profile.

The internal structure and foundation of the shelf-edge reef was examined using seismic profiles, a diver-operated sediment drill-probe, and a coring drill (Jones et al., 1992; Blanchon and Jones, 1995). Seismic data acquisition was largely from the west (leeward) and south (exposed-windward) shelf (Fig. 6.2). In all, 25 one-metre cores were obtained from dead coral substrates on the west, north, and south sides of the island in water 16 to 26 m deep (Fig. 6.2). Corals in the cores, which are all aragonite with traces of high Mg calcite, yielded uncorrected radiocarbon ages of <1000 yrs BP with most having ^{14}C activities sufficiently low that only an age of <300 yrs could be assigned.

Architecture of a Shelf-Edge Reef

Morphology

Rimming the shelf around Grand Cayman is an impressive build-up of coral that extends virtually unbroken along the 87-km-long perimeter of the island (Fig. 6.1). This shelf-edge reef (SER) occupies the outer 200 m of the lower terrace and extends over and down the shelf escarpment to a depth of ~75 m. Nowhere does the reef reach the surface and, at its maximum elevation, is still 12 m below msl.

The SER consists of a closely-spaced series of large, coral-armored buttresses dissected by steep-sided, sediment-floored canyons (Fig. 6.3). In plan, the buttresses are ~100 m long and 10 m wide and have a triangular wedge-shape that tapers shoreward. Although their long axes are generally aligned 90° to the shelf edge, they can deviate ±20° from this (Roberts, 1974). In profile, the buttresses have a vertical to steeply sloping front that typically rises from ~40 m on the upper part of the shelf escarpment into waters ~20 m deep (Fig. 6.3B). From there, buttresses flatten out and extend shoreward across the lower-shelf terrace at an average depth of ~25 m.

Each buttress consists of three architectural elements; a wall, a crown, and a spur (Fig. 6.3B). The wall rises steeply from ~40 m forming a near vertical, shield-like escarpment that commonly overhangs in its upper sections (Fig. 6.4A). The sides, and more rarely the backs, of the buttresses are also bounded by steeply sloping walls. Being shaded for most of

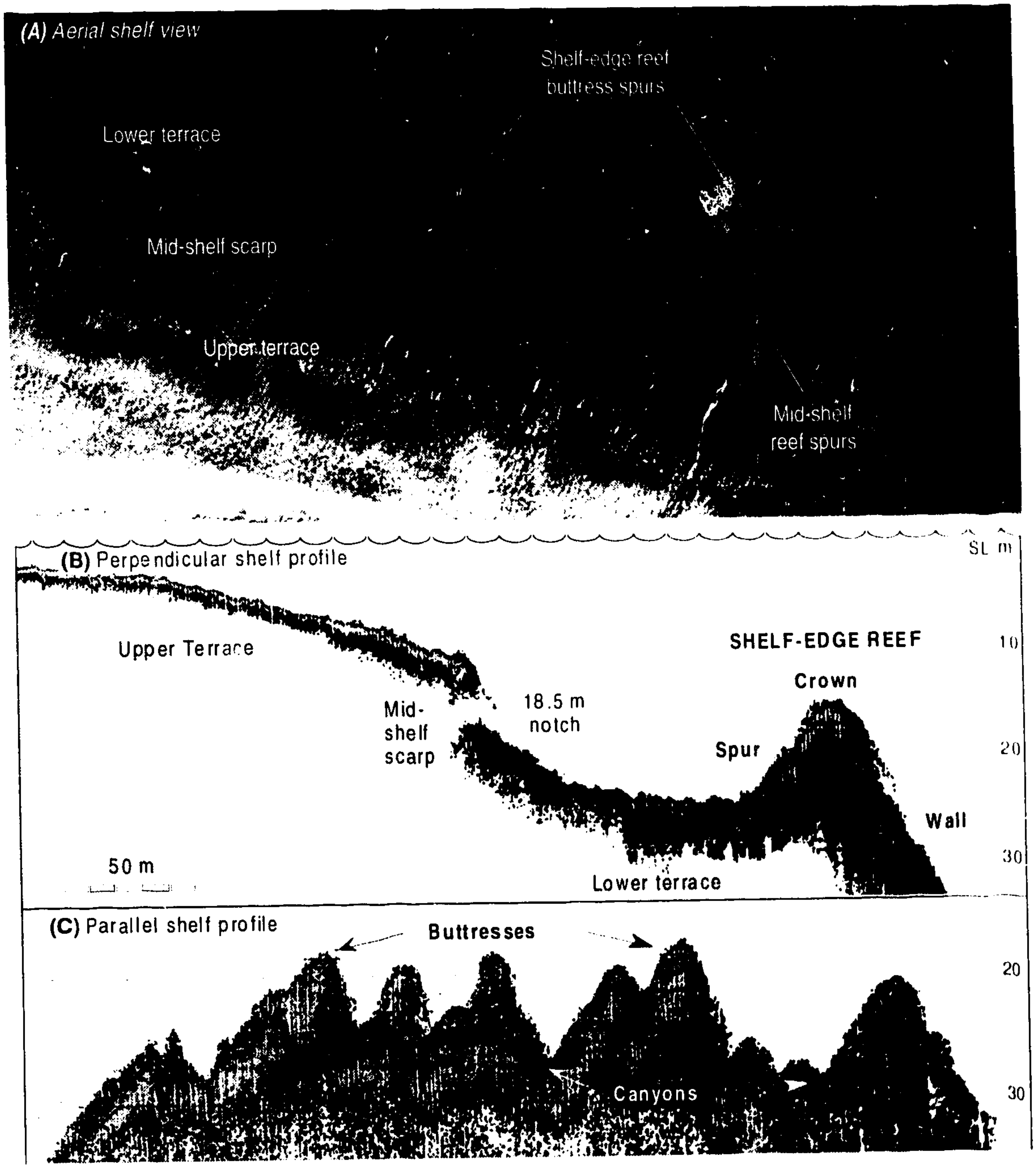

Figure 6.3. General shelf-edge-reef architecture. (A) Aerial photograph showing terraced nature of shelf and extent of shelf-edge reef (site Sp.B). (B & C) Sonar profiles showing morphology of terraces, architectural elements of buttresses, and buttress-canyon relations (site N2).

the day, these substrates generally receive low light. Turbulence is also reduced at these depths with the only water movement being related to the westward drift or more local ebb-tidal currents. In both cases currents rarely exceed 30 cm/sec and are usually < 10 cm/sec (Roberts et al., 1975; Darbyshire et al., 1976). The crown, which forms the upper part of the

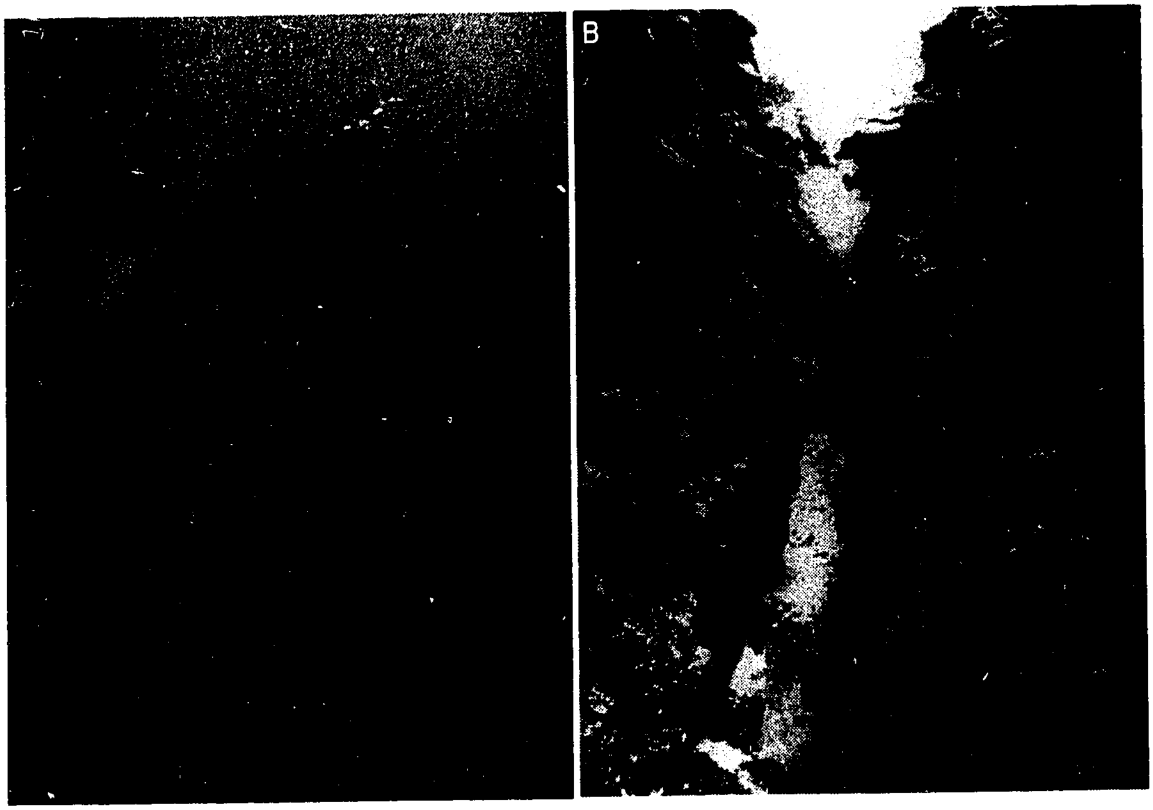

Figure 6.4. General views of shelf-edge reef. (A) View of buttress-canyon array looking shoreward (buttress crown at 18 m below msl, site N18 ~100 m west of N22). (B) View looking shoreward up a steep, narrow, canyon (site GB in 36 m of water; *M. cavernosa* in foreground is 50 cm tall).

buttress, is a broad, gently rounded dome that rises into waters 15-20 m deep. It has a slightly asymmetrical profile toward its seaward side, and is delineated from the wall by an abrupt break in slope. This shallow, well illuminated part of the buttress is affected by weak oscillatory currents generated by fair-weather wave orbitals. Like the wall, it is also affected by unidirectional currents induced by the westward drift and local ebb-tidal currents. The spur, which starts where the crown flattens and begins to slope shoreward, tapers back onto the lower terrace (Fig. 6.3A). Spur length ranges from a few metres to the width of the terrace (150-300 m) and, in general, the longer the spur the lower its amplitude. Light levels and turbulence are similar to the crown.

Buttresses are separated by narrow canyons that are commonly floored with skeletal sand- and cobble-sized sediment (Figs. 6.3C, 6.4B). These features, which begin as flat sandy areas between spurs, narrow and steepen in gradient towards the shelf edge. Depending on the morphology of the buttresses, canyon walls are sub-vertical to overhanging and commonly converge to produce tunnels. Locally, canyons may be blocked by coral debris that is shed from adjacent buttresses, but most remain open and extend down to the upper parts of the shelf-edge escarpment at 40-50 m. Canyon light-levels, which are controlled by buttress amplitude and spacing, are generally low. Turbulence is also low, but shoreward moving currents (~35 cm s^{-1}) can be funneled into the canyons as a result of the interference between the westward drift and buttresses (Roberts et al., 1977)

Structure

The position of the SER over the shelf break means that it is flanked by the gently-sloping lower terrace on one side and the sub-vertical shelf escarpment on the other. On the lower terrace, loose skeletal sand accumulates behind the SER buttresses and builds up reducing the vertical expression of the reef to only a few metres (Fig. 6.5). This limited relief is misleading, however, because seismic profiles show that if the sediment wedge were removed, exposing the underlying bedrock terrace, the vertical expression of the reef would be 10 to 15 m (Fig. 6.5). Furthermore, the foundation of reef growth on the seaward side is much deeper and seems to be associated with a narrow terrace at 60-70 m (Darbyshire et al., 1976). From this depth, coral cover rapidly increases and buttress walls become distinct at ~40 m in most areas. The vertical expression of the SER on its seaward side is, therefore, as much as 40 m.

Although seismic data reveals little about the internal features of SER buttresses, scuba observations offer valuable insight. In many areas, pinnacles of coral grow up from the slope just in front of the buttress wall (Fig. 6.6). These structures, founded in waters as deep as 50 m and rising up to 25 m, are bounded by vertical to overhanging walls that commonly widen and mushroom out in their upper sections (Fig. 6.6C). Where the upper part of the shelf-edge escarpment is a moderately inclined slope, pinnacles are located up to 30 m seaward of the buttresses. Where the slope is more steeply inclined, however, development

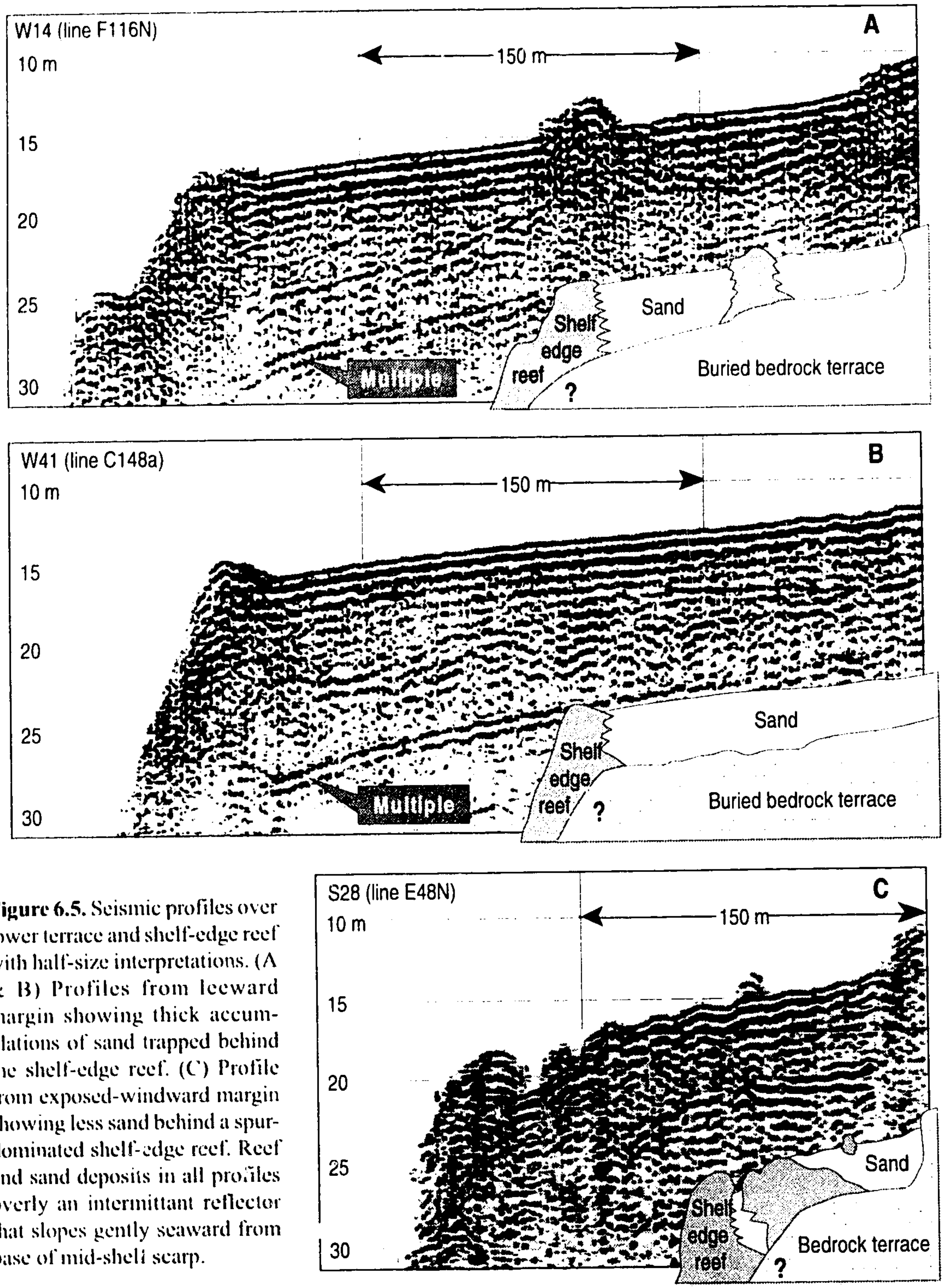

Figure 6.5. Seismic profiles over lower terrace and shelf-edge reef with half-size interpretations. (A & B) Profiles from leeward margin showing thick accumulations of sand trapped behind the shelf-edge reef. (C) Profile from exposed-windward margin showing less sand behind a spur-dominated shelf-edge reef. Reef and sand deposits in all profiles overly an intermittant reflector that slopes gently seaward from base of mid-shelf scarp.

starts closer to the buttress and pinnacles commonly amalgamate with the wall or crown to produce a characteristic arch (Fig. 6.6C). In some areas a series of these pinnacle-arch structures are commonly observed one in front of the other at the base of the buttress.

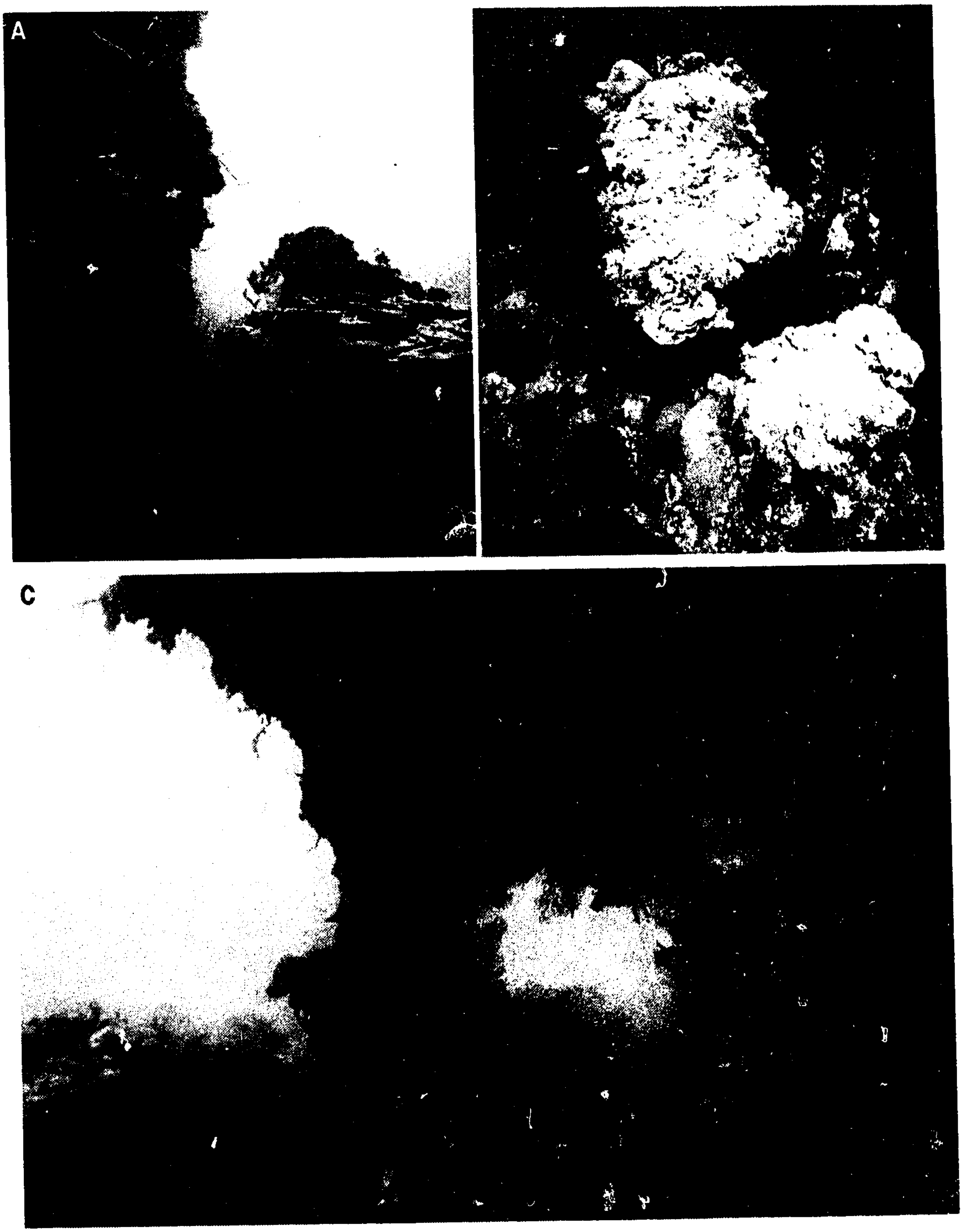

Figure 6.6. Pinnacles developm[illegible]t of in front of shelf-edge reef buttresses. (A) Profile of pinnacle in early stages of development growi[illegible] at [illegible]ase of wall (~30 m depth, site N14). (B) Plan view of pinnacle in late stages of development joining fr[illegible] of [illegible]ttress crown and still showing original gap (~15 m depth, site N33). (C) Profile of buttress base sho[illegible] [illegible]inctive pinnacle-and-arch structure formed when pinnacle 'mushrooms' and amalgamates with wall. N[illegible] that pinnacle toppled into wall at some stage (~30 m depth, site N14).

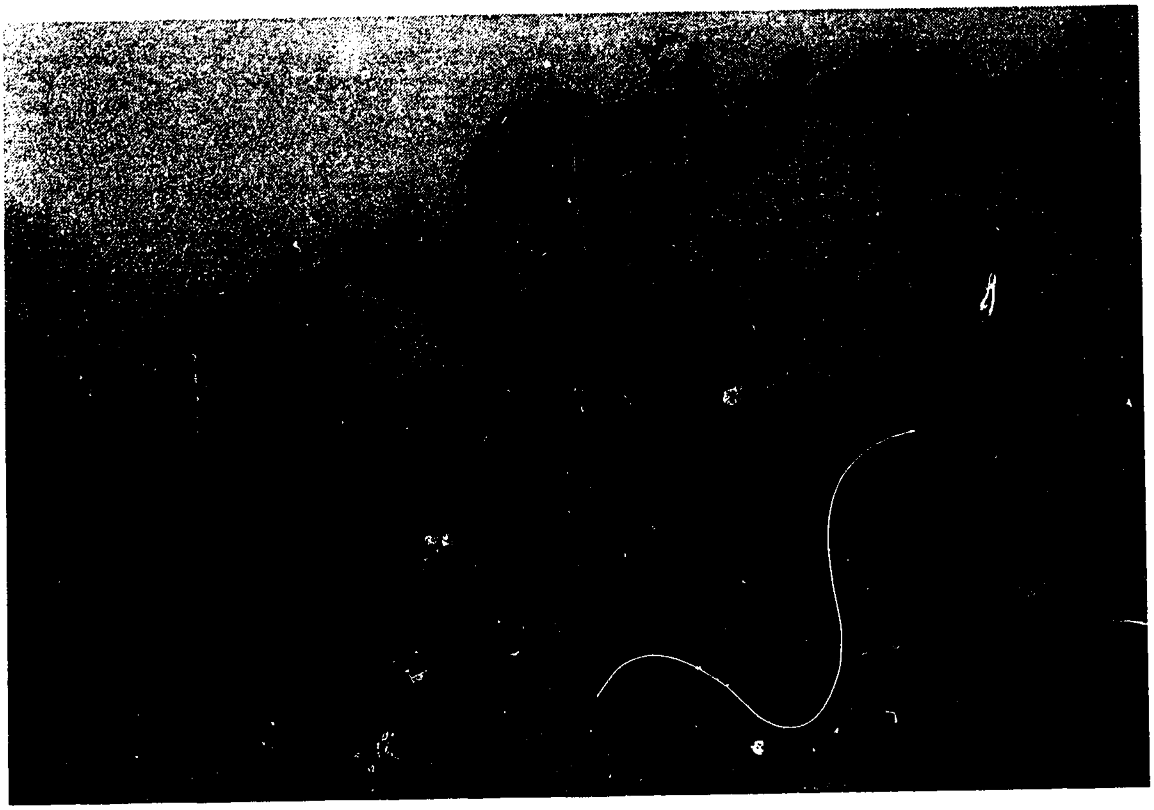

Figure 6.7. General view of shelf-edge reef biota. A diverse benthic community of corals, gorgonians, soft corals, sponges, coralline algae, calcareous algae, and fleshy brown and green algae all compete for space on a buttress. This oblique, seaward-looking view of buttress side-wall shows how severe this competition can be (~16 m depth, site N14).

Biota

The benthic community on the SER, which is the most diverse of any found in the island's marine environments, is largely composed of stony corals, sponges, soft corals, gorgonians, fleshy algae, and calcareous algae (Fig. 6.7). Of these, the stony corals dominate and, of 47 species identified, 33 are found on the SER alone (Hunter, 1994). The SER coral community has been divided into 3 associations (Hunter, 1994): (1), the *Stephanocoenia-Madracis* association, found on the crown and spur and dominated by *Montastrea annularis* and *Agaricia agaricites*; (2), the *Agaricia lamarcki* association, found on the wall to ~45 m and dominated by *Agaricia lamarcki*, *A. agaricites purpurea*, and *M. annularis*; and (3), the *Agaricia undata* association, found on the upper shelf-edge escarpment to ~85 m and dominated by *A. lamarcki*, and *A. undata*.

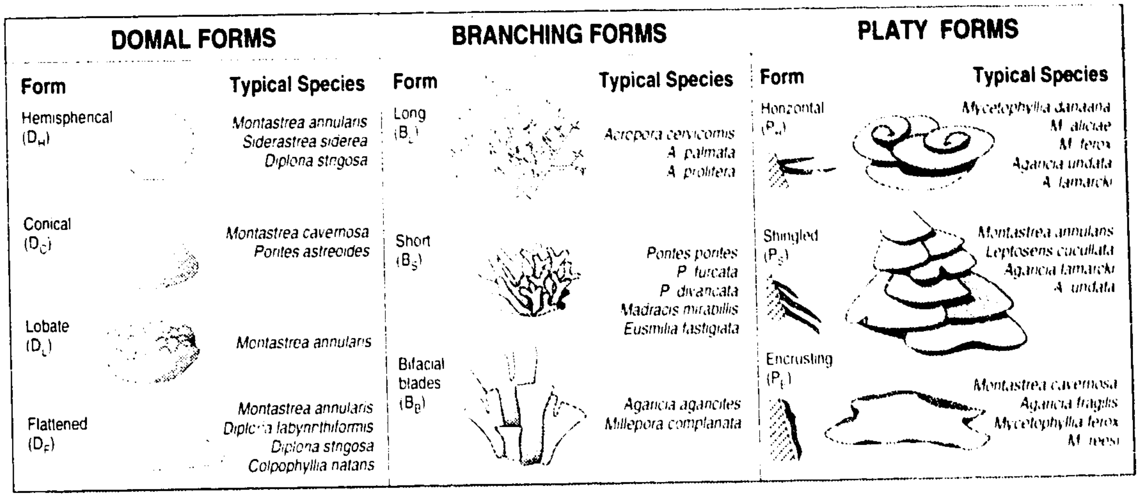

Figure 6.8. Ten common coral forms found on shelf-edge reef with repre:entative species. These forms, which make up over 95% of coral species present, are divided into 3 form groups: domes, branches, and plates. Note that several species, particularly *Montastrea*, have more than one growth form.

Unfortunately, this type of species-zonation approach can hinder comparison between different reefs because no two reefs have the same associations arranged in the same pattern (Fagerstrom, 1987). On Grand Cayman, for example, the dominant coral species in the *Stephanocoenia-Madracis* association, vary significantly around the island: on the west side, the association is dominated by *A. agaricites* and *A. cervicornis* (Rigby and Roberts, 1976), whereas on the south side it is dominated by *Diploria* sp., *Colpophyllia* sp., *M. cavernosa*, and *M. annularis* (Ghiold and Smith, 1990).

An alternative approach is to classify corals into functional/structural groups with similar ecological requirements (Fagerstrom, 1987, 1991). Coral form, for example, allows a comparison of reefs independent of genera or species, and permits comparison of different parts of the same reef in terms of physical processes. This approach is adopted here, and corals on the SER are classified into 10 specific forms which fall into 3 general form groups (Fig. 6. 8). Estimates of the percentage cover of each form type, as well as total percentage of coral cover, show systematic changes over each buttress (Fig. 6.9). Although relations are difficult to descern when all forms are plotted (Fig. 6.9A), the three form groups are systematically distributed (Fig. 6.9B), with walls dominated by platy-forms, crowns by domal forms, and spurs by branching-forms (Figs. 6.10). Total coral coverage for the wall and spur

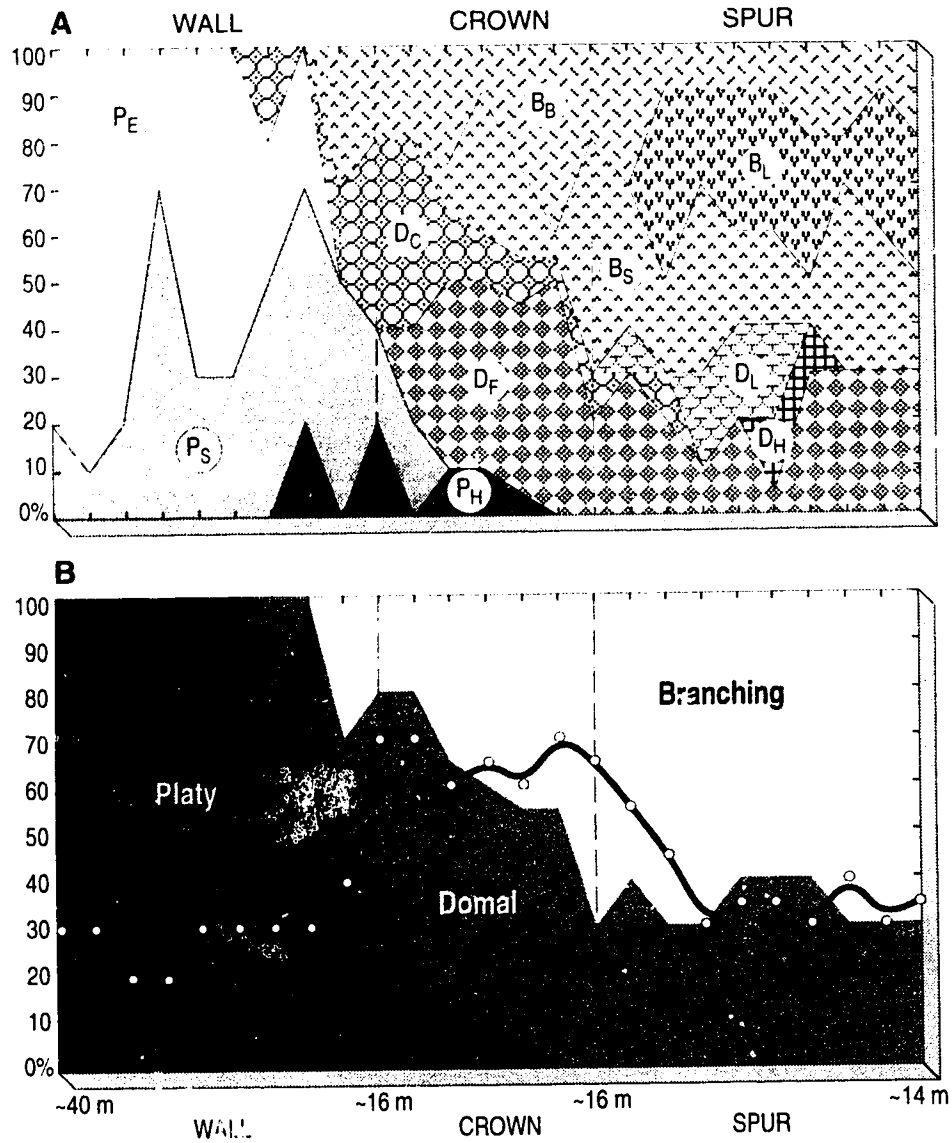

Figure 6.9. Coral form zonation and coral cover across a typical shelf-edge reef buttress (site N10). (A) Normalized abundance of forms from wall to spur (form abbreviations defined in figure 8); few relations are apparent. (B) Same plot showing only form groups. Note distinct zonation with platy forms dominating wall, domes dominating crown, and branches dominating spur. Line shows variation in coral cover, with typical high cover over crown, decreasing down to base of wall and end of spur.

zones typically 30-40%, whereas crown coverage is 60-70% (Fig. 6.9B). Also, coverage decreases towards the end of the spurs and to the base of the wall where it is as low as 20%.

Framework

Cores used to investigate the framework character of the SER come mostly from buttress crowns. Although the short length of cores limits full assessment of framework, two types

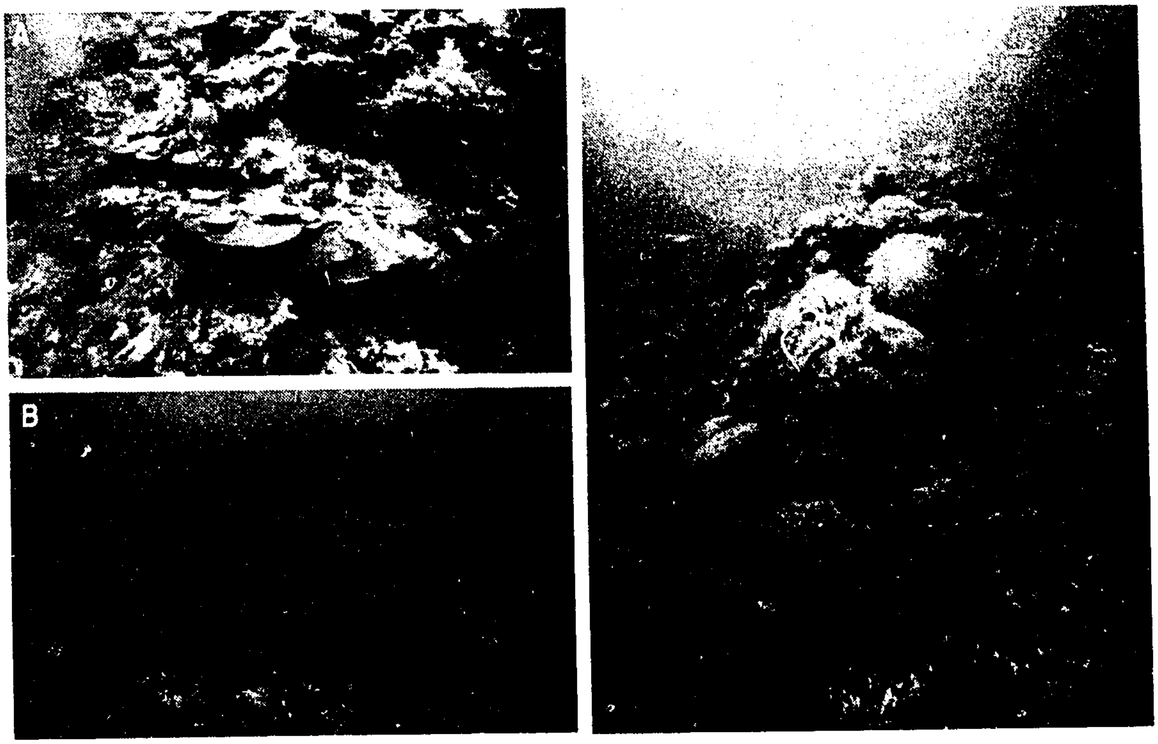

Figure 6.10. General views of buttress zones showing form zonation of corals. (A) Shingled and encrusting platy forms dominate wall (30 m depth, site S31). (B) Long and short branching forms dominate spur (15 m depth, site N18). (C) Low-domal forms dominate crown. Corals up to 1 m in diameter (20 m depth, site N31).

are evident (Fig. 6.11). In the first type, domal forms dominate and consist mostly of *M. annularis* with subordinate *M. cavernosa* and *S. siderea*. Many of the under surfaces of these corals are colonized by a diverse community of epi- and endobionts. The epibionts show two distinct associations; one consisting of sclerosponges (dominated by *Ceratoporella* sp.), cemented brachiopods, and encrusting foraminifera (*Carpentaria* sp.), and the other of coralline algae and encrusting foraminifera (*Homotrema rubrum*). Similar associations are found in cryptic and semi-cryptic reef cavities in shallower reefs elsewhere on the island (Logan, 1981). The second framework type consists of *A. cervicornis* branches in a matrix of skeletal sand and gravel. Although the cavities in this low density framework are smaller and partly filled by sediment, they still harbor abundant epibionts (Fig. 6.11).

Collectively, the core evidence suggests that the crowns of the SER buttresses have an open framework with large, cryptic shelter cavities enclosed by the overgrowth and coalescence of domal and branching corals.

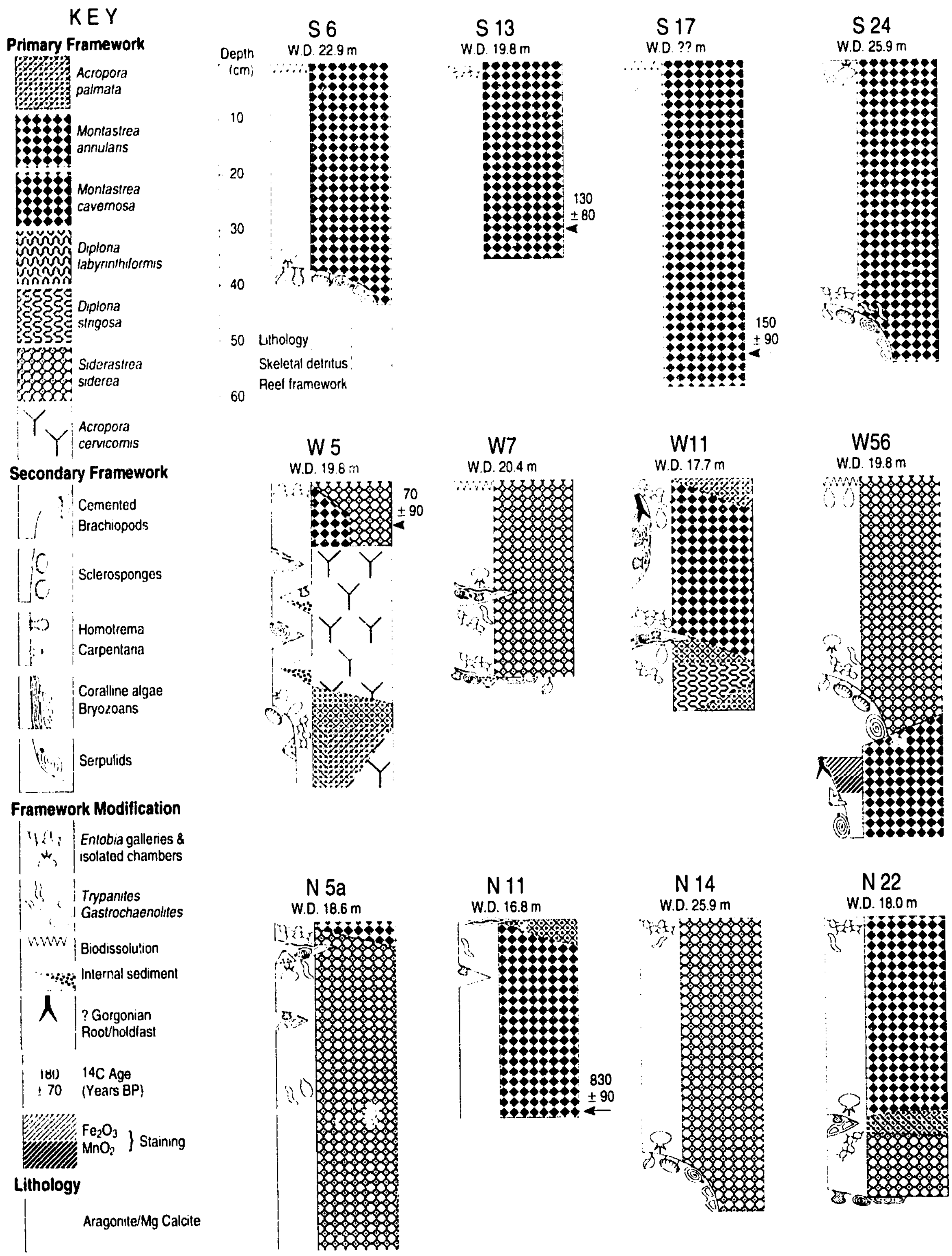

Figure 6.11. Framework character of selected shelf-edge reef cores. Main type shown is dom[illegible]ed framework with large open cavities lined with cryptic associations of epi- and endobionts. Other fra[illegible] types include branch-dominated framework (e.g., core W5). This type is not as well [illegible]sented because mo[illegible] cores are from buttress crowns. Note, cores W11 and W56 contain roots/hold-fasts Identification and dating is presently underway to determine significance of these features.

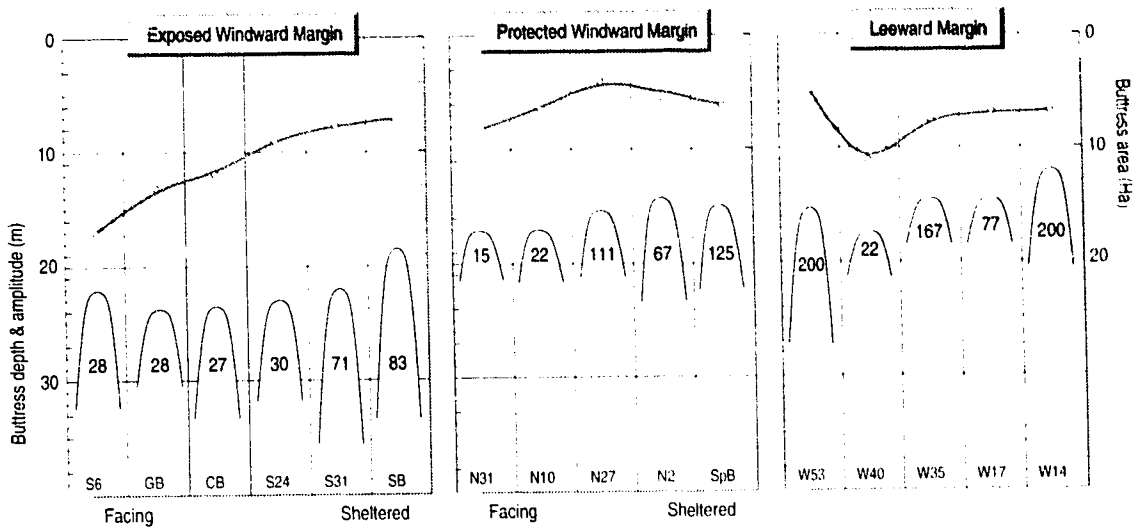

Figure 6.12. Variation in buttress depth (m below msl), amplitude (m), area (ha), and spur frequency (number in buttress icons) measured from sonar profiles and aerial photographs over shelf-edge reef. Trend in buttress depth and amplitude shows distinct architectural styles for each margin (amplitude determined from average crown versus canyon depth on sonar). Exposed-windward buttresses are deeper and have larger amplitudes than either protected-windward or leeward margins. Note relation between local margin orientation (facing/sheltered), buttress area, and spur frequency along both windward margins. Facing sections of reef have large buttress areas and high spur frequencies (low numbers), but both parameters decrease markedly as shelf-edge reef becomes more sheltered.

Architectural Variation

SER architecture shows a systematic variation as margin orientation changes. As a result, distinct reef styles are found on the exposed-windward, protected-windward, and leeward margins. This variation is evident from buttress amplitude and area, spur frequency, and coral coverage and form abundance (Figs. 6.12, 6.13, 6.14).

Exposed-Windward Margin

The SER along the exposed-windward margin has the most extensive aerial coverage, the highest buttress amplitudes and coral cover, and the greatest proportion of large, robust, domal forms of any part of the reef around Grand Cayman (Fig. 6.15). Typically, these buttresses have steep gradients and large amplitudes. They slope from ~15 to ~25 m (spur to crown) and have an average amplitude of 12 m, although buttresses with 25-30-m amplitudes are present. They are bounded by vertical, rarely overhanging, front and side walls. Buttress crowns develop in water depths averaging 22 m (the deepest of any margin) and usually

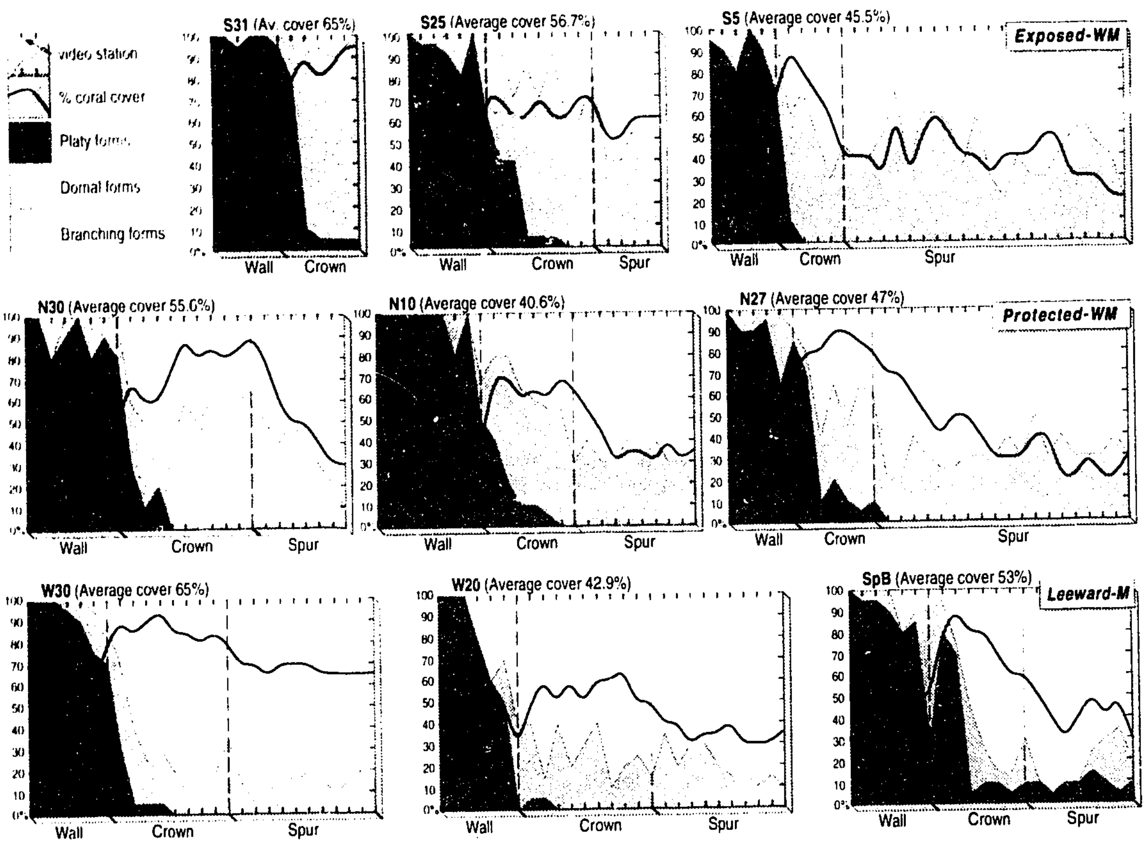

Figure 6.13. Variation in coral form and cover measured from video transects over shelf-edge-reef buttresses. Several relations are apparent: walls are invariably dominated by platy corals; branching corals gradually increase in proportion from leeward, through protected-wind, to exposed-windward margin; crowns on windward margins are dominated by domal forms. Note, site Sp.B is transitional between leeward and protected-windward margins; its coral form zonation is typically leeward, whereas its morphology is protected-windward in character (cf., Fig. 6.14).

have a broad, gently inclined seaward face (Fig. 6.14). In wind-facing areas, buttress spurs have high amplitudes (averaging 4 m), a high frequency (1 every 27 m), and cover more than half of the lower terrace (up to 17 ha), extending across its full width to merge with spurs on the upper-terrace. In sheltered areas, however, spur length, frequency, and coverage values are significantly lower (average spur frequency of 1 every 75 m and coverage of <8 ha).

Exposed-windward buttresses are characterized by slightly higher than average coral cover of 56% (Fig. 6.13). Like other margins, buttress walls are dominated by platy forms, but unlike the others they have an unusually high cover (50-60%). Crowns have the highest cover (up to 95%), and domal forms account for 70-95% of the forms present. Spurs also

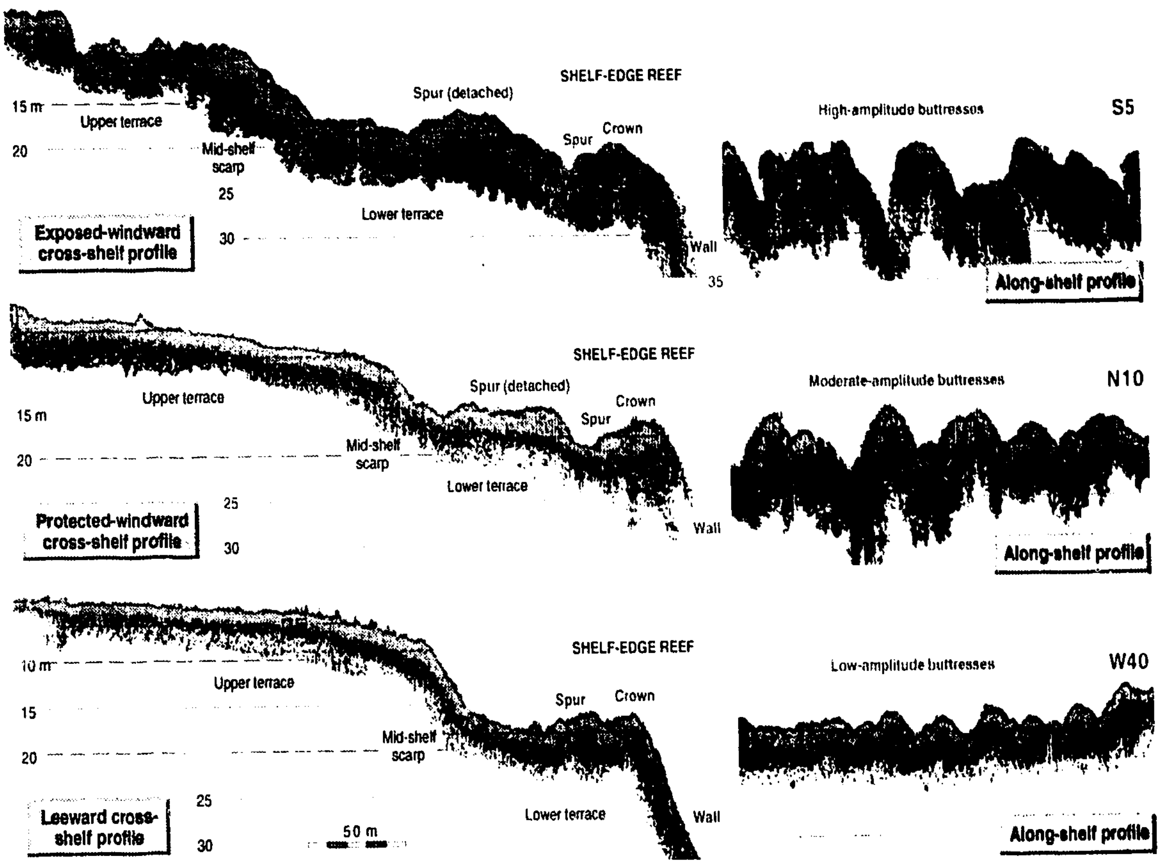

Figure 6.14. Selected sonar profiles over shelf and shelf-edge reef showing variation in buttress depth, amplitude, and morphology. Along-shelf profiles were taken over apex of buttress crowns. Note decrease in buttress amplitude from exposed-windward to leeward margin, and 60-70° sloping wall along leeward margin. Also, note variation in crown depth and morphology: exposed-windward crowns are deeper and have a gently inclined seaward face.

have a high proportion of domal forms and, although cover values are lower than other parts of the buttress, they are higher than spurs from other margins (Fig. 6.13).

The high-amplitude exposed-windward buttresses are separated by correspondingly deep, narrow, and vertically-sided canyons that steepen abruptly from the backs of the buttresses at ~25 m, to ~45 m at the fronts where they open onto the sub-vertical shelf-edge escarpment (Fig. 6.14).

Protected-Windward Margin

The SER buttresses along this margin typically extend into shallower water, have intermediate amplitudes, moderate spur development, and are dominated by robust domal

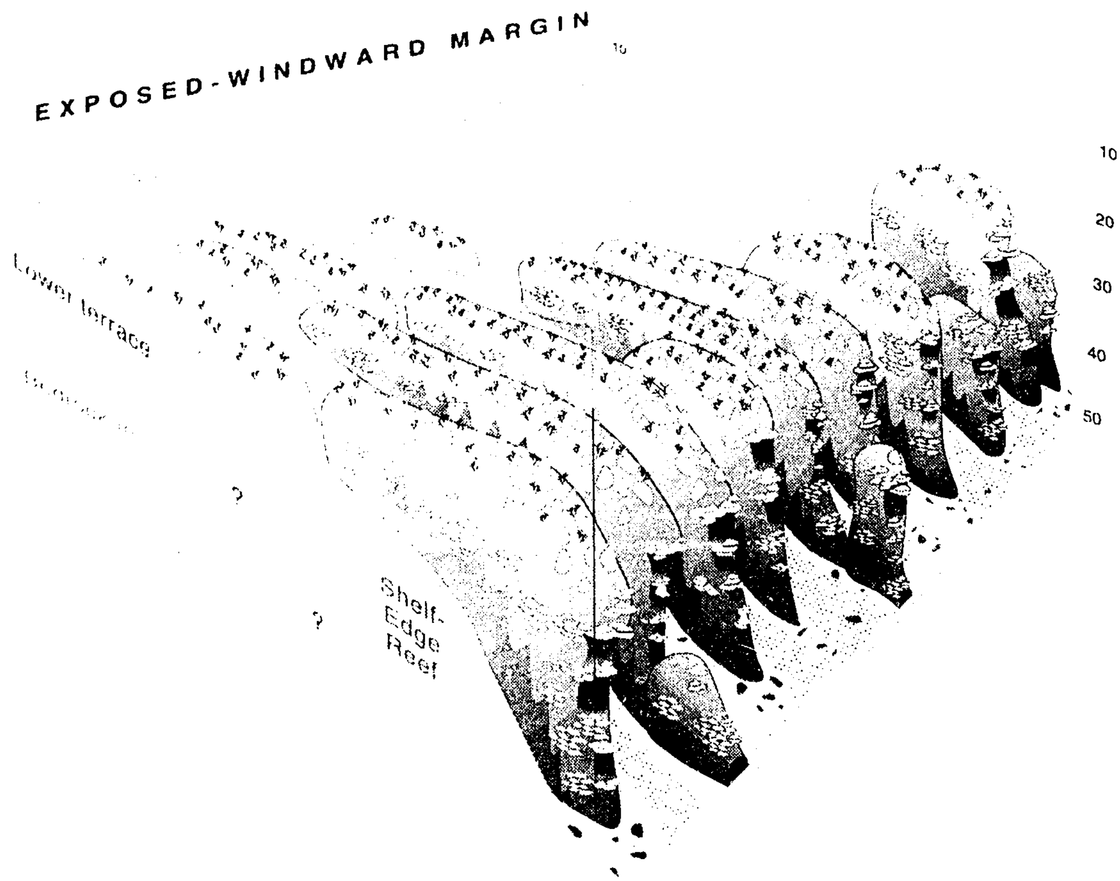

Figure 6.15. Schematic summary of shelf-edge-reef architecture along exposed-windward margin. Buttresses are high amplitude, aerially extensive, and dominated by robust domal corals. Crowns are typically deeper than ~20 m and have gently-sloping seaward face. Note how spur length diminishes as local margin orientation changes from wind-facing to sheltered.

and weak branching corals (Fig. 6.16). In general, they slope gently seawards from ~14 to ~16 m and have amplitudes of 5-10 m (Figs. 6.12, 6.14). These buttresses are distinctive because they overhang in upper parts of the wall and generally have a symmetrical dome-shaped crown (Fig. 6.14). Crown depth, which is much shallower than along exposed-windward margins, is typically 16 m. Along wind-facing areas, spurs have a high frequency (1 every 19 m) and extend across the entire width of the terrace. Unlike the exposed-windward spurs, however, protected-windward spurs have a low amplitude (average 2.5 m) and aerial coverage (<8 ha). Along sheltered areas there is a reduction in spur frequency (1 every 90 m) and coverage (~4 ha) (Fig. 6.12).

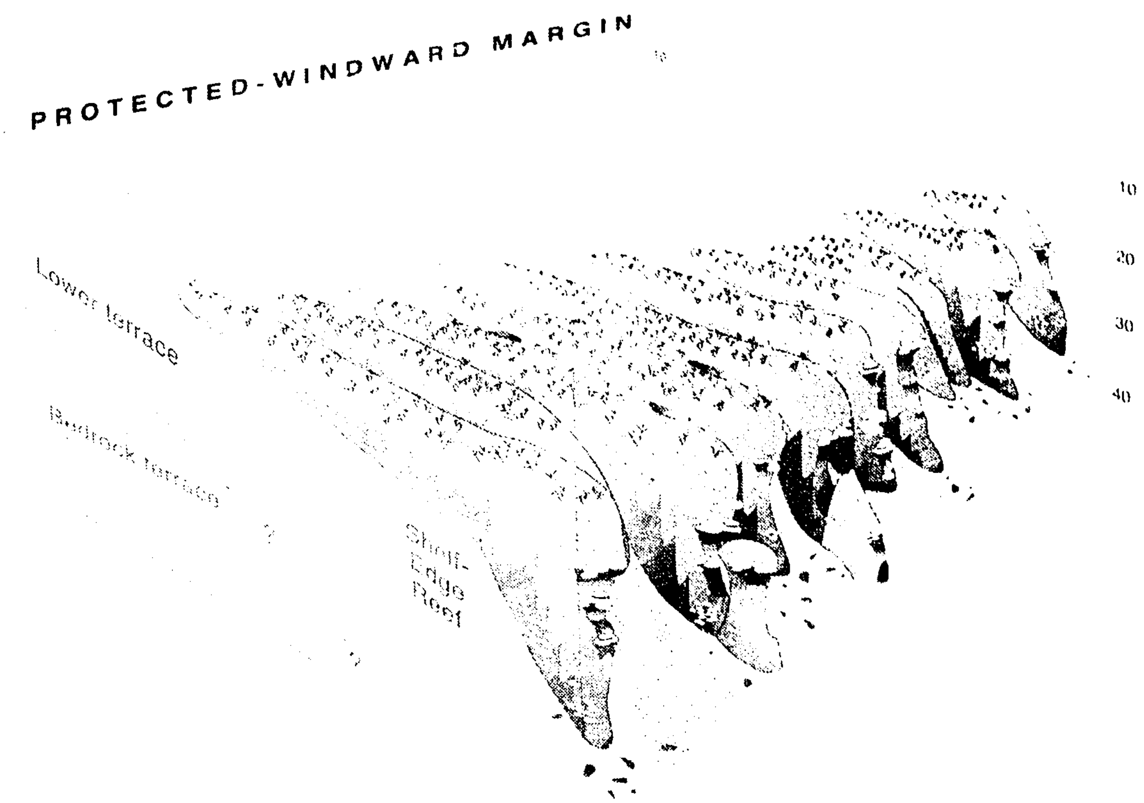

Figure 6.16. Schematic summary of shelf-edge-reef architecture along protected-windward margin. Buttresses have moderate amplitudes, moderate aerial coverage, and are dominated by domal corals on crowns and branching corals on spurs. Crowns are typically shallower than 20 m and form simple mounds. Spur length diminishes as local margin orientation changes from wind-facing to sheltered.

The protected-windward buttresses have a tripartite coral-form zonation: platy corals dominate the walls, domal corals dominate crowns, and branching corals dominate spurs (Fig. 6.13). Although they have a slightly lower than average coral cover (48%), buttress cover commonly reaches ~90% in crown areas but decreases as low as 20% on wall and spur.

Intervening canyons are narrow and typically bounded by overhanging buttress walls. With lower buttress amplitudes, however, canyons are much shallower than those along exposed-windward margin, sloping from ~18 to ~25 m before terminating abruptly at the shelf break (Fig. 6.14). Along wind-facing sections, the narrow canyons are overgrown and commonly roof-over forming tunnels. In more sheltered locations, however, canyons tend to remain open.

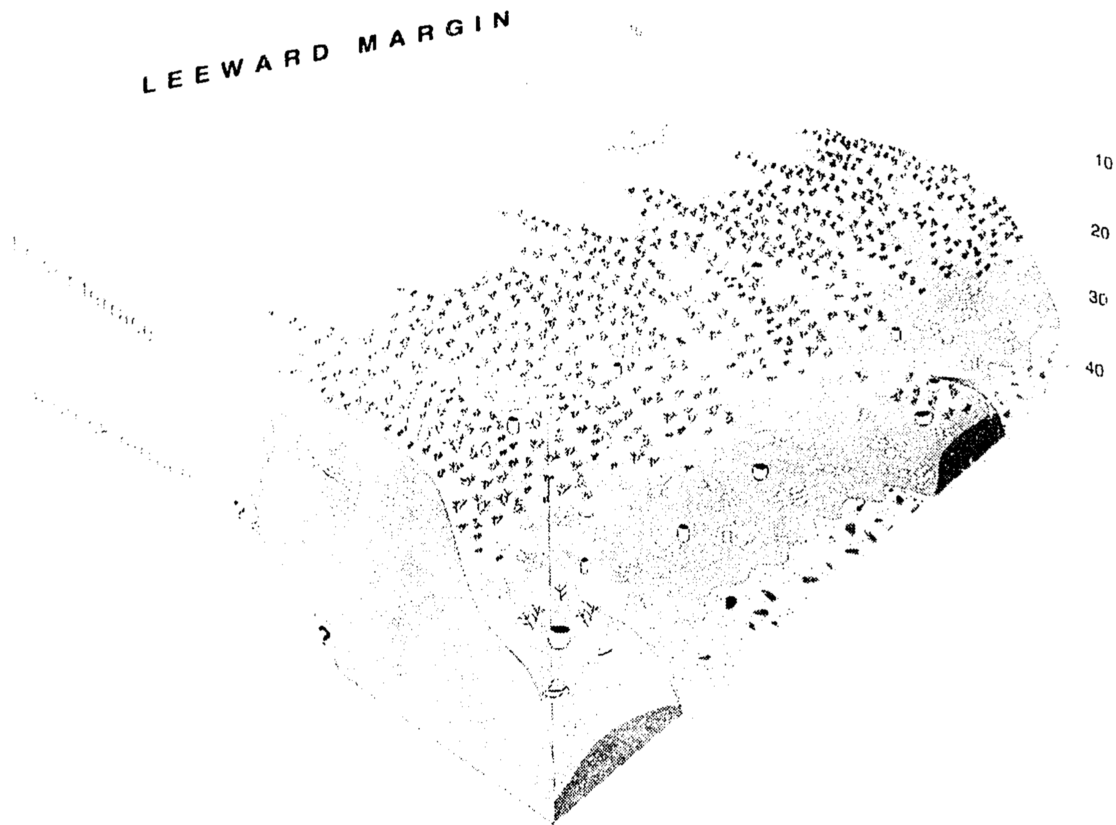

Figure 6.17. Schematic summary of shelf-edge-reef architecture along leeward margin. Buttresses-canyon architecture breaks down and buttresses merge laterally producing a belt of branching coral. Buttresses are still discernible as low-amplitude ridges. Note coral ledges at ~25 m extending seaward from sloping wall; significance of these features is unclear (see also seismic profile F116N in fig. 5).

Leeward Margin

Although largely protected, parts of the leeward margin are exposed to moderate southwest and northwest swells. Consequently, the style of the SER is somewhat variable. In general, reef architecture is characterized by shallow low-amplitude buttresses, little spur development, and a dominance of weak branching corals (Fig. 6.17).

Along embayed and fully protected sections of the leeward margin, the characteristic buttresses-canyon structure of the SER breaks down and buttresses merge into a continuous belt of coral devoid of canyons. Buttresses in these areas form gentle ridges with an average amplitude of 3-4 m and slope gently seaward from ~14 to ~20 m (Fig. 6.14). Unlike other margins, the wall is not vertical but forms a moderate 60-70° slope. In its upper part, this slope gradually flattens to become a diffuse crown at depths of 15-20 m. It then slopes gently upward, back onto the lower terrace where there is an abrupt junction with the sand deposits of the lower terrace. Few distinct spurs develop (frequency as low as 1 spur every 200 m), but where present, they are low amplitude (<1 m) and extend only a limited distance onto the terrace.

The style of the SER along slightly more exposed parts of the leeward margin is similar to parts of the protected-windward margin, and consists of a moderate amplitude buttress-canyon array.

As in other areas, the highest coral cover on leeward buttresses are over the crowns and decreases down the wall and toward the back of the reef (Fig. 6.13). Form zonation, however, differs significantly from other areas. The gentle change in slope from the wall to crown is paralleled by a gradual change in form types. Also, crown and spur zones cannot be differentiated using form because both are dominated by branching corals. Branching forms are common even on the wall (Fig. 6.13).

At rare sites, the SER along the leeward margin is broken by very broad canyons that are aptly named 'sand chutes' or 'rivers-of-sand' by divers. These chutes are deep and slope from ~15 to 40 m. On either side of the chutes, buttresses have significant relief with their side walls rising steeply from 40 to 20 m below msl.

Physical Processes Affecting Shelf-Edge Reefs

Fairweather waves

Architectural variation of the SER around Grand Cayman is related to the physical processes that are acting on it. Roberts (1974), for example, found a positive correlation between wave power and spur frequency and suggested that subtle changes in spur orientation from the lower to upper terrace correlated with fairweather wave refraction patterns. To test the hypothesis, Roberts et al. (1975) obtained continuous measurements of water movement over a two week period across the exposed-windward shelf (south coast). This showed that the influence of fairweather waves was only significant in shallow waters over the upper terrace; lower-terrace SER spurs were dominated instead by currents associated with the westerly drift. In spite of these data, Roberts et al. (1975) still argued that fairweather waves were somehow involved, and speculated that lower-terrace spurs must have been initiated during a lower stand of sea level. As sea level rose, spurs continued to develop under weaker wave-stress regime and increased their spacing by amalgamation (Roberts et al. 1975).

Adding to the spur frequency/orientation data of Roberts (1974), this study shows that SER buttress amplitude, area, and coral form zonation, varies systematically as margin orientation changes. This suggests that SER morphology must be controlled by ambient physical processes rather than being an inheritance of fairweather wave processes associated with a lower sea level. There are two reasons for this: (1) variation in buttress form zonation is consistent with variation in the intensity of an actively operating process and, (2) if the orientation of buttress spurs are controlled by fairweather waves, as Roberts et al. (1975) suggests, then they should have the same orientation as present-day spurs on the upper terrace--which they do not. Furthermore, any argument invoking readjustment to ambient fairweather waves is not supported by the data which shows that fairweather waves have little or no hydrodynamic influence on the lower terrace (Roberts et al., 1975). Nevertheless, SER spur orientation *is* consistent with wave refraction patterns. So, if spurs are not controlled by fairweather refraction patterns, then they must be related to refraction of storm-waves.

Storm waves

Although few data exist for Grand Cayman, the impact of tropical cyclones on other reefs is well documented (Hedley, 1925; Blumenstock 1958; Stoddart, 1962; Goreau, 1964; Glynn et al., 1964; Cooper, 1966; Ball et al., 1967; Perkins and Enos, 1968; Shinn, 1976; Randall and Eldredge, 1977; Ogg and Koslow, 1978; Dollar, 1982; Rogers et al., 1982; Aitp and Stearn, 1986). Few studies, however, have adequately documented impacts in deeper shelf areas (Woodley et al., 1981; Harmelin-Vivien and Laboute, 1986; Kobluk and Lysenko, 1992; Hubbard, 1992; Blair et al., 1994). However, together with hindcast data (Hernandéz-Avila et al., 1977; Kjerfve et al., 1986) and in-situ water column measurements (Snedden et al., 1988; Hubbard, 1992), a coherent picture of the processes that act on SERs during hurricanes is emerging. Upon approaching a shallow shelf, hurricane-generated surface currents cause a build-up of water along the coast, producing a positive storm surge and generating a compensatory return current (Swift et al., 1986). Across narrow shelves, where there is insufficient time for the Coriolis force to cause deflection, the return current flows directly offshore at velocities of up to 1.5 m s^{-1} (Hubbard, 1992). This powerful return flow is augmented by storm-wave oscillation controlled by storm-wave height and frequency (Nummedal, 1991). Together, the return and oscillation currents produce a combined-flow regime characterized by the development of a severe unidirectional pulsing current (Swift et al., 1986) that can attain velocities of up to 4 m s^{-1} at depths of ~30 m (Kjerfve et al., 1986; Hubbard, 1992; Blair et al., 1994).

Hurricane-generated currents have a devastating impact on all open-shelf reefs. Extreme turbulence associated with combined flows, causes extensive fragmentation of structurally weak branching and platy corals to depths of 30 m (Hernandéz-Avila et al., 1977; Randall and Eldredge, 1977, Woodley et al., 1981; Tunnicliffe, 1983; Kjerfve et al., 1986). Stronger domal colonies are also damaged by saltating projectiles to depths of 23 m (Kobluck and Lysenko, 1992) and are abraded by suspended sand to 30 m (Kirby-Smith and Ustach, 1986; Hubbard, 1992). Furthermore, turbulence may induce debris flows from shallower reef zones that damage coral growing as deep as 90 m (Harmelin-Vivien and Laboute, 1986; Woodley et al., 1981). Hurricane-induced combined-flows also cause significant suspension and fluidization of sediments associated with SERs. Several studies, for example, have reported the removal or redistribution of significant quantities of sand from SER canyons and outer-shelf sand plains at depths of at least 30 m (Kirby-Smith and Ustach, 1986; Hubbard, 1992; Blair et al., 1994). Clearly, hurricane-induced currents have the potential to cause significant architectural modifications to SERs.

Discussion

Hurricane-controlled reef architecture

We suggest that architectural variation of the SER around Grand Cayman results from varying intensity of hurricane-generated currents. The relation between buttress area, spur frequency, and local margin orientation (Fig. 6.12) is a result of differences in exposure to hurricane-generated waves. Sections of the SER that face directly into oncoming storm or hurricane-generated waves, are prone to high levels of coral fragmentation. Coral detritus shed from individual buttresses is carried shoreward over the lower terrace by the extreme waves, forming spurs of debris that progressively extends across the terrace following each storm. As a result, not only are spurs more frequent and extensive in wind-facing areas, but their orientations coincide with storm-wave orthogonals (Roberts, 1974). More obliquely-oriented sections of the SER are also pruned by the storm waves but the detritus is washed off- or along-shelf, thereby inhibiting the formation of spurs. Only when storm waves approach from less common directions will spurs be produced in these areas. This buttress-shedding process is also responsible for the formation of pinnacles. Coral detritus shed seaward off the buttress accumulates as talus piles on the slope at the base of the wall. Protected from turbulence and sedimentation, these piles are ideal sites for coral colonization and vertical accretion between storms produces pinnacles.

The hurricane-induced fragmentation of corals on upper parts of SER buttresses is also consistent with form zonation patterns (Fig. 6.13). The dominance of robust domal corals on the crowns of windward buttresses implies that these areas are severely impacted by storm waves, and weaker branching forms are regularly removed or 'pruned.' Fragments of these corals are carried shoreward across the spur and lodge in crevices and cavities. This pruning and shoreward transport is a vital process because fragmentation of branching corals, such as *A. cervicornis*, facilitates asexual regeneration (e.g., Gilmore and Hall, 1976; Shinn, 1976; Tunnicliffe, 1981; Highsmith, 1982). This pruning process may, therefore, explain the dominance of branching forms on spurs. Alternatively, this dominance might be explained by sand suspension during hurricanes. The large repository of sand on the lower terrace would create high stress levels over the spurs for all but the best adapted branching forms (Hubbard and Pocock, 1972; Tunnicliffe, 1983). Corals on the crowns, walls, and pinnacles, however, would be relatively isolated from such stress because sediment is flushed off shelf through the canyons and effectively by-passes these zones. Buttress crowns and pinnacles are therefore ideal habitats for rare, sediment-intolerant corals such as *Eusmilia fastigiata* and *Meandrina meandrites* (Hubbard and Pocock, 1972).

In addition to explaining local architectural features, the intensity of hurricane currents also explains variation in SER architecture around Grand Cayman. Although at any location wind speed and duration are consistent, the difference in fetch around the island produces significant differences in wave-height. Thus, the exposed-windward margin receives the highest waves and the leeward margin the lowest. As a result, the destructive effect of hurricanes progressively decreases from the exposed-windward margin to the leeward. This is consistent with trends in crown depth and coral-form zonation (Figs. 6.12 and 6.13 respectively). Buttress form-zonation shows a progressive decrease in the proportion of branching-corals from leeward, to protected-windward, to exposed-windward margins. This decrease reflects the increased efficiency of buttress pruning by hurricane waves. Greater crown depths along the exposed-windward margin are also explained by increased efficiency of pruning: larger and more powerful storm waves prune deeper and therefore limit vertical crown accretion.

Larger storm-waves and greater turbulence along the exposed-windward margin are also more effective at removing sand from around the SER (Fig. 6.18). This explains why buttress amplitude is greater, and the level of the lower terrace is deeper, along this margin (Blanchon and Jones, 1995). In contrast, less efficient sand removal along protected-windward and leeward margins means that sand builds up behind the SER, producing smaller-amplitude buttresses and a shallower, flatter lower terrace (Figs. 6.12, 6.14). This build-up of sand also results in decreased buttress area as less substrate is available for coral growth (Fig. 6.12).

Figure 6.18. Base of buttress spur showing erosion of sand by hurricane Gilbert (1988). Arrows show ~1 m difference between pre-storm and post-storm sand level (~20 m depth, site S5; courtesy of Phillipe Bush).

Hurricane-controlled reef accretion

If hurricane-generated waves and currents modify the architecture of the SER, then they might also control reef accretion patterns. We examine this possibility by combining the results of this study with core data from other SERs, and develop a process-response model of SER accretion.

Although many SERs have been reported from the Caribbean (Goreau and Goreau, 1973; Morelock et al., 1977; Hubbard et al., 1976; James and Ginsburg, 1979; Roberts and Murray, 1983; Hine and Steinmetz, 1984; Holmes and Kindinger, 1985; Lidz et al., 1991), only those in St. Croix, Barbados, and Belize have been cored. With its similar oceanographic setting, the SER around St. Croix has an architecture comparable to that of Grand Cayman (cf., Hubbard 1989). Cores drilled on spurs of the exposed-windward buttresses recovered 7 m of domal-coral framework underlain by a drowned 9 ka-old *A. palmata* reef (Adey et al., 1977). Hubbard et al., (1986) found a similar situation in a drowned river canyon along the protected-windward margin on the north side of the island. There, horizontal drilling into the buttress wall at 30 m recovered ~6 m of alternating coral detritus and domal framework, before passing into drowned *A. palmata* framework. The alternating detritus and framework showed a complex lateral accretion pattern with distinct alternations in radiocarbon dates, which suggested that the reef had prograded by slumping (Hubbard et al., 1986). At another protected-windward location, Hubbard (1989) also drilled several vertical holes on SER buttresses at depths of 15-18 m, and found in excess of 4 m of domal framework on the buttress crown (core CB-2) and 2 m on the spur underlain by coral detritus (core CB-7).

The SERs around Barbados and Belize are also similar to Grand Cayman's SER (cf., Ott, 1975; Rutzler and Macintyre, 1982). Cores recovered the southern (exposed-windward) margin of Barbados contained 9 m of domal framework growing over a 7.5 ka-old *A. palmata* reef (Fairbanks, 1989). Whereas a core from the Belize SER recovered 12 m of domal and branching framework and found no underlying *A. palmata* reef (Macintyre et al., 1982).

The cores from St. Croix, Barbados, and Belize show that SERs have undergone at least 12 m of vertical and 6 m of lateral accretion since establishment. Seismic profiles across the leeward SER on Grand Cayman show it is at least 10 m thick and possibly as much as 40 m thick (Fig. 6.5).

Accretion model

Our architectural analysis shows that vertical and lateral accretion of the SER around Grand Cayman is controlled by the efficiency of buttress pruning and shedding during hurricanes. Vertical accretion is controlled by the effective depth to which hurricanes prune fragile and weakened corals from the tops of buttresses. This depth, which is proportional to storm-wave height and therefore varies with margin orientation, is termed the hurricane accretion threshold (HAT). Leeward buttresses have a shallow HAT because they are rarely affected by large waves. Consequently, they have grown to within 12 m of msl. In contrast, exposed-windward buttresses have a deep HAT because they are commonly impacted by fully-developed hurricane waves. These buttresses rarely grow any shallower than 20 m (Fig. 6.12). The consistent depth of buttresses along windward margins suggest that the SER has reached its HAT and is now accreting laterally.

Lateral accretion is controlled by the frequency of buttress shedding. Regular pruning of fragile or unstable corals from buttresses produces a near continuous supply of detritus. Detritus shed off-shelf and deposited in front of the wall is colonized by coral during fairweather, initiating pinnacle development. Over time, pinnacles accrete vertically until

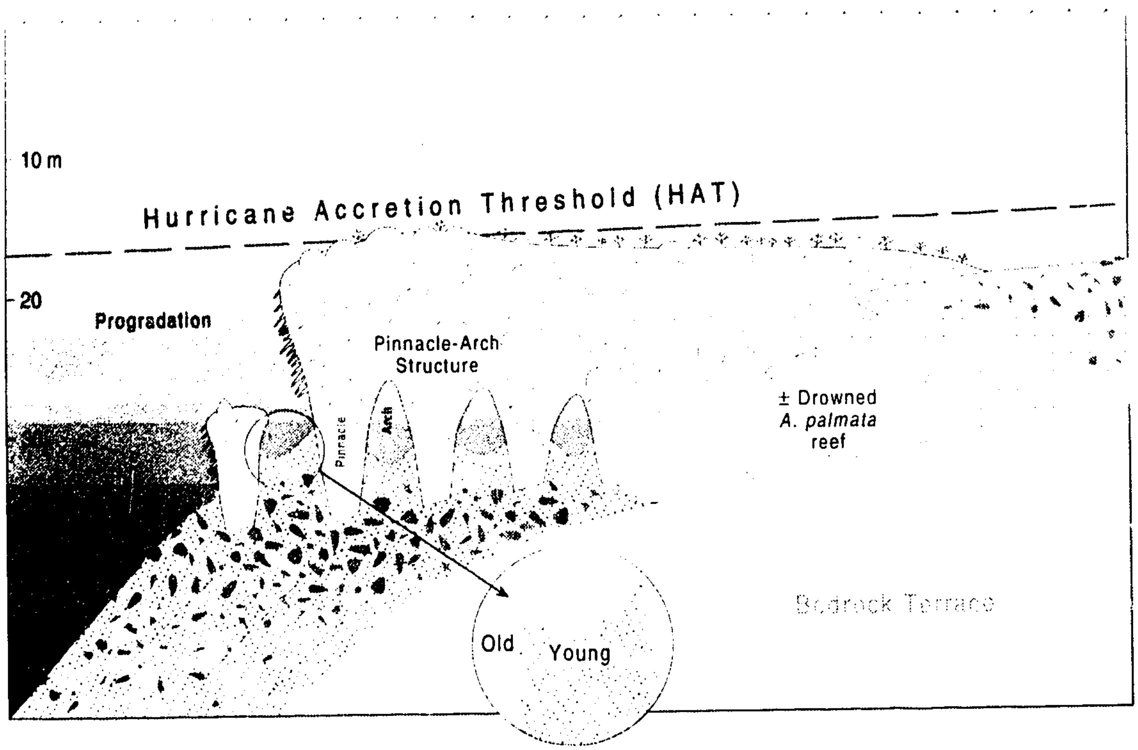

Figure 6.19. Shelf-edge reef accretion model. Reef accretes vertically until hurricane-accretion threshold is reached. At that point hurricanes remove fragile or weakened corals at same rate as they can be replaced by new coral growth. Reef now moves into shedding phase and large amounts of coral detritus are supplied to shelf and proximal slope area. This supply of detritus facilitates lateral accretion by initiating pinnacle growth and providing a substrate over which reef can prograde. Inset shows juxtaposition of old pinnacle base and young sediments shed from buttress above.

either the HAT is reached or the pinnacle amalgamates with the buttress wall (Fig. 6.19). Pinnacles further away from the wall only become linked when the gap is filled by detritus or coral growth. Pinnacle growth, amalgamation, and in-filling produces a characteristic pattern of lateral accretion that is manifested as pinnacle-and-arch structures seen on many buttress walls around Grand Cayman (Fig. 6.6). These features, together with the presence of pinnacles in all stages of development, suggest that the SER around Grand Cayman has undergone a significant amount of lateral accretion (Fig. 6.19).

Pinnacle amalgamation and in-filling also explains the complex patterns of lateral accretion identified in cores from the SER around St. Croix (Hubbard et al., 1986). Horizontal alternations of rubble and coral framework represent the alternation between pinnacles and in-filling detritus. The older coral pinnacles are surrounded by younger detritus that has been regularly shed from the adjacent buttress. Thus, dating of horizontal cores through the reef framework produces alternations of old/young radiocarbon dates (Fig. 6.19).

Modern SERs have been established for only the last 7 ka when sea level began to stabilize near its present position (Blanchon and Shaw, 1995). Yet, in that time, they have undergone a significant amount of hurricane-mediated lateral accretion. This pattern of accretion explains the tendency of carbonate margins to produce steep precipitous slopes: SERs step out over their own debris in a balancing act between lateral growth and slope failure, and in doing so, prograde into the basin. This is reminiscent of the accretion patterns of many ancient shelf-margin reefs. Judging from the delicate nature of the metazoans that constructed these ancient frameworks, it is interesting to note that some of these reefs may also have been deep-shelf structures that did not grow into the surf zone (e.g., see description of Capitan Reef in Babcock and Yurewicz, 1989). Perhaps a reconsideration of their architecture and sedimentology will reveal that they too accreted laterally under the influence of storms and hurricanes.

Conclusions

The shelf-edge reef around Grand Cayman is a distinct and impressive structure adapted to deeper-shelf conditions and varying intensities of hurricane disturbance. Our analysis has shown that variation in the approach and height of hurricane-generated waves controls reef architecture and accretion in four ways:

1) Local coral-form zonation over buttresses reflects differing intensities of hurricane-induced stress. Crowns are influenced by turbulent stress favoring coral associations with robust domal forms, spurs are influenced more by sediment stress favoring associations of branching forms, and walls are permanently shaded favoring platy forms.

2) Local variation in shelf-edge reef architecture is controlled by the angle of hurricane-wave approach. Sections of the reef that face in-coming waves are pruned of weaker corals and the detritus is swept back over the shelf producing large buttresses, high spur

frequencies, and spur orientations that are coincident with storm-wave refraction patterns. Reef sections oblique to in-coming storm waves are also pruned, but the detritus is swept off-shelf producing buttresses with limited spur development.

3) Regional variation in architectural style results from varying intensity of buttress pruning and sand removal by hurricanes. Progressive decrease in the proportion of branching coral and depth of buttress crowns from leeward to windward margins, reflects a progressive increase in the depth and efficiency of buttress pruning by storm waves. Progressive increase in buttress amplitude similarly reflects increased efficiency of sand removal by hurricane-induced currents.

4) Vertical accretion of the shelf-edge reef is limited by the repeated pruning effect of hurricanes. Once the reef reaches this hurricane-accretion threshold it accretes laterally as a result of buttress shedding, pinnacle growth, in-filling, and/or amalgamation. The long term effect of hurricanes, therefore, is to cause lateral progradation. This style of accretion accounts for the common tendency of carbonate margins to produce steep, precipitous slopes.

This predominantly lateral mode of accretion, together with its adaption to deeper-shelf conditions, make the shelf-edge reef a useful analog for those ancient shelf-margin reefs whose biota was too delicate to be surf-resistant.

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7

Synopsis of Reef Development Around Grand Cayman: The Shelf-Gradient Control Theory

Introduction

Probably the most elusive and certainly the most long-standing problem in the study of modern reefs has been identifying the controls on reef configuration (Chapter 1). Darwin (1842) concluded that reef configuration was simply the result of upward coral growth during relative sea-level rise. Fringing reefs first grew close to shore and subsequently developed into more distant barrier reefs and atolls as their foundations subsided. Yet many were convinced it was not that simple (see review by Davis, 1928). Simultaneous upward growth could not explain why, within a single reef system, fringing reefs transformed laterally into barrier reefs (Semper, 1881). Also, if just simple growth were involved, fringing-reef corals should migrate retrogressively upslope maintaining their contact with shore rather than building vertically into a barrier (Steers and Stoddart, 1977). Abandoning the idea of upward growth therefore, later investigators emphasized the role of the underlying antecedent shelf topography in explaining reef configuration. Daly (1916) suggested that the three modern reef types resulted from differential marine erosion of shelves during glacially-lowered sea level. This was later superseded by the suggestions of MacNeil (1954), and Purdy (1974), who argued that it was differential subaerial erosion during lowstands that was responsible for the different reef types. But evidence for these theories was largely circumstantial.

With the advent of shallow underwater drilling during the late 1970's, hard data became available and the notion of antecedent topographic control was put under considerable scrutiny (e.g., Hopley, 1982). Although studies from back-reef lagoons did indeed show that some reefs had developed on residuals of older Pleistocene reefs (Halley et al., 1977; Marshall and Davies, 1984), other studies in open-shelf areas showed that reefs grew on flat, seaward-sloping terraces of unknown origin (summarized by Macintyre, 1988). Furthermore, deeper drilling discovered that reefs as thick as 55 m developed during the last deglacial sea-level rise (Fairbanks, 1989), falsifying the hypothesis that modern reefs were simply veneers over an antecedent substrate.

These discoveries have brought the argument concerning modern reef configuration full circle: if neither simple growth nor antecedent foundations are responsible, then what controls the siting and architecture of modern reefs? By investigating and describing reef development around Grand Cayman this thesis has attempted to identify the processes and agents responsible for reef configuration (Chapters 3-6). This final chapter summarizes these findings and, by placing them in context with work on other modern reefs, proposes a preliminary model of reef development that shows how the interplay between sea-level rise and coastal gradient controlled where, when, and how modern reefs developed.

Reef Development Around Grand Cayman

Like many other Caribbean islands, the terraced shelf around Grand Cayman hosts two major reef types: a wave-dominated fringing reef covered by surf-adapted corals such as *Acropora palmata*, and a submerged shelf-edge reef covered by a diverse deep-water coral association (Fig. 7.1). The fringing reef, which closely parallels the windward coast, consists of several shore-parallel zones that are centered either side of the reef crest. Shoreward of the reef crest, these zones form an unconsolidated wedge of sediment that grades from coral cobbles to coral/foraminiferal sand. Seaward of the reef crest, zones have a variable cover of surf-adapted corals, particularly *A. palmata*, interspersed with broad areas of coral cobbles and boulders stabilized by a crust of coralline algae. Rather than being composed of an interlocking framework of coral, however, the seaward zones are underlain by one or more layers of coral-cobble conglomerate, each capped by a crust of coralline algae. Grain size and compositional trends are consistent with this detrital reef core being deposited during hurricanes. Storm waves destroy the cover of corals and transport their detritus, together with sand entrained from the lower terrace, into the storm breaker zone where it is

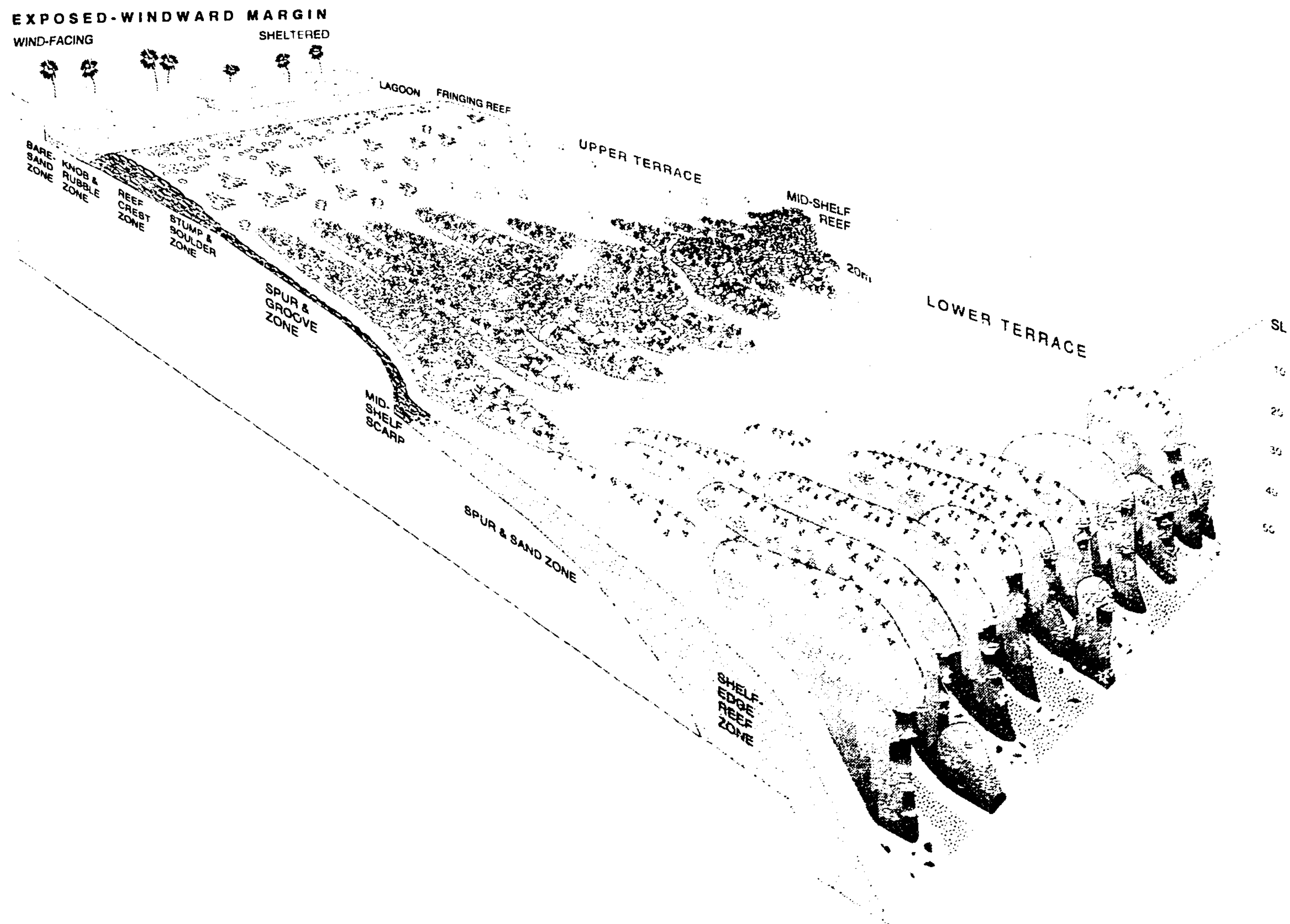

Figure 7.1. Schematic summary of reef development and zonation on the shelf around Grand Cayman. Note architectural variation in both fringing and shelf-edge reefs as margin orientation becomes more sheltered.

washed onto and over the reef crest producing a detrital deposit that thins and fines into the lagoon. By determining how far onto the upper terrace the detritus is carried, storm waves control the location of subsequent fringing-reef development. On wide parts of the terrace that experience large storm waves, the detritus is deposited farther shoreward than in areas that experience fetch-limited storm waves. On narrow parts of the terrace, however, storm waves transport detritus onshore and a fringing reef does not form. Following storms, the rapid growth of new corals ensures that the surface cover will completely regenerate before the next hurricane. This cyclic pattern of growth and destruction consequently produces a fringing reef that is constructed of superimposed layers of hurricane-generated detritus.

Rimming the outer shelf of the island is a submerged, 87-km long shelf-edge reef that rises to 15-20 m below msl. It consists of an array of coral buttresses and pinnacles that are aligned perpendicular to shore and separated by narrow, steep-sided, sediment-floored canyons. The coral cover and architecture of these buttresses varies systematically as margin orientation, and exposure to hurricane-generated seas, changes. Along fully exposed windward margins, hurricane waves remove significant quantities of sand from intervening canyons and cause extensive fragmentation of weak branching corals on buttress crowns. This pruning and sediment removal produces high relief, deep buttresses that are dominated by structurally robust domal corals. Along semi-protected margins, smaller hurricane waves cause less fragmentation and remove less sand producing shallower, lower relief buttresses dominated by a combination of branching and domal corals. Along fully protected margins, the normal buttress-canyon architecture breaks down producing a series of shallow, branching-coral-dominated ridges that merge laterally into an unbroken belt of coral. Over time, the repeated pruning by hurricanes limits the vertical accretion of buttresses and detritus is increasingly shed into the sloping fore-reef area. Consequently, once the shelf-edge reef attains its vertical hurricane accretion threshold, it starts to accrete laterally via pinnacle growth and infilling by detritus.

Although studies in other areas have considered fringing reefs and shelf-edge reefs to be a single 'reef complex' (Goreau and Goreau, 1973; Rutzler and Macintyre, 1982; Hubbard, 1988), differences in configuration, architecture, biota, and ambient physical conditions between the reefs around Grand Cayman suggests they are separate structures juxtaposed on a narrow shelf. Their independence is supported by two observations. First, active shelf-edge reefs are present in areas were fringing reefs are absent, such as along the leeward margin of Grand Cayman. And second, shelf-edge reefs are reported from much wider shelves where they are widely separated from inner-shelf fringing-reefs (Donnelly, 1965; Macintyre, 1972; Colin, 1977; Roberts and Murray, 1983; Hine and Steinmetz, 1984; Holmes and Kindinger, 1985; Lidz et al., 1991; Harris and Davies, 1989; Macintyre et al., 1991). In other words, these two reef types are separate structures with different configurations fringing reefs are always located on shallow parts of the shelf whereas shelf-edge reefs are always located around the shelf edge, regardless of shelf width.

The common occurrence of similarly configured fringing and shelf-edge reefs in other parts of the Caribbean implies that reef development has similar controlling factors. There are two possible alternatives: 1) configuration of reefs is controlled by the location of antecedent topographic residuals or, 2) configuration is controlled by contemporary environmental processes. Around Grand Cayman, the fringing reef is underlain by a flat terrace and not by an antecedent ridge (Chapter 5). This is also the case for many fringing reefs investigated in other areas (Goreau and Land, 1974; Adey, 1975; Macintyre and Glynn, 1976; Easton and Olson, 1976; Shinn et al., 1982; Burke et al., 1989). Seismic profiles suggest this is also the case for the Grand Cayman's shelf-edge reef (Chapter 6), but the possibility of an underlying ridge cannot be ruled out until a deep hole is drilled. Shelf-edge reefs in several other areas, however, show no sign of being underlain by topographic ridges (Macintyre et al., 1982; Adey et al., 1977; Hubbard et al., 1986; Fairbanks, 1989). There is, however, one exception: a recent seismic survey over the south Florida shelf revealed the presence of a ridge along the shelf edge (Lidz et al., 1991). Cores demonstrate that this ridge has a variable thickness of modern corals underlain by an older Pleistocene reef (Toscano et al., 1994).

Despite this exception, evidence suggests that most modern reefs, including those around Grand Cayman, are independent structures that owe their configuration and architecture to intrinsic environmental processes and generally not to the location of antecedent topography (Macintyre, 1988). If this is so, then a fundamental question remains: which of these processes or agents controlled reef configuration?

Controls on Fringing-Reef Development

Although no data exist for Grand Cayman, drilling around other Caribbean islands has shown that fringing reefs composed of *A. palmata* were established on the edge of many shelves by the early Holocene (Adey et al., 1977; Lighty et al., 1978; Fairbanks, 1989). These rapidly accreting reefs tracked sea-level rise until ~7.6 ka when they suddenly died off (Chapter 3). In some areas, the dead reefs were immediately colonized by deeper-water corals, and new *A. palmata* reefs began to develop in sites 5 m higher upslope (Chapter 3). Yet in other areas, new reefs did not develop in upslope sites for several thousand years (Adey, 1978; Marshall, 1988). Even within the same reef system, the lag in *A. palmata* re-establishment varied by over 4 ka (Burke et al., 1989).

At the time these relict reefs were discovered, it was widely accepted that modern reef accretion could easily match the average rate of deglacial sea-level rise (Schlager, 1981). Consequently it was thought that the demise of *A. palmata* reefs could not simply be attributed to drowning. Instead it was postulated that early reef development was controlled by the interplay between antecedent shelf topography and sea-level rise (Adey, 1978; Neumann and Macintyre, 1985; Macintyre, 1988). The main idea was that, as sea-level rapidly flooded the flat antecedent shelf, erosion of lagoonal and coastal deposits created turbid and eutrophic waters that suffocated the early reefs (Adey, 1978). Later modifications of this idea included killing the reefs with 'inimical waters' flushed from oversized lagoons (Neumann and Macintyre, 1985; Hallock and Schlager, 1986), or cool waters derived from widespread upwelling (Kinsey and Davies, 1979; Marshall, 1988). None of these explanations, however, accounted for the variable lag in reef re-establishment. Nor did they explain why, if conditions for reef growth had deteriorated, the dead reefs were immediately colonized by deep-water corals.

Critical to these explanations of Holocene reef development was the assumption that deglacial sea-level rose smoothly with only gradual changes in rate. Yet the uniform elevations, similar ages, and immediate colonization by deep-water corals, suggest that these early Holocene reefs were killed by a synchronous widespread event that only affected *A. palmata*. Furthermore, some of the new reefs that backstepped and re-established upslope did so within 100 years of the early reef demise. This selective die-off and rapid backstepping can only have been produced by a sudden increase in the rate of sea-level rise (Chapter 3). This pulse was sufficiently rapid to remove the *A. palmata* reefs from their optimum 5 m growth window and translate reef growth at least 5 m higher upslope to the new surf-zone position (Chapter 3).

Another assumption was that the terraced shelf foundations were antecedent (Adey, 1978). Yet the presence of terraces with similar elevations around other islands, especially those that had undergone uplift during the Pleistocene, suggested this was not the case (Chapter 4, Fig. 3-20). Shelf terraces around Barbados, for example, could not be antecedent because the island underwent uplift at a rate of 25-45 cm ka^{-1} during the Pleistocene (Bender et al., 1979; Fairbanks, 1989). Consequently, if the terraces formed between the last interglacial ~125 ka ago (MIS 5) and the last glacial interstadial ~40 ka ago (MIS 3) they should have experienced at least 10-30 m of uplift. Their lack of uplift, however, demonstrates that they formed instead during the last deglacial sea-level rise. This conclusion is supported when elevations of terraces and slope breaks on stable islands are compared with the pattern of deglacial sea-level rise identified from elevations and ages of drowned reefs (Chapter 3, Fig. 3.2). Terraces on Grand Cayman, for example, coincide precisely with episodes of slow sea-level rise and the positions of slope breaks correspond to the pulses or steps in the rise

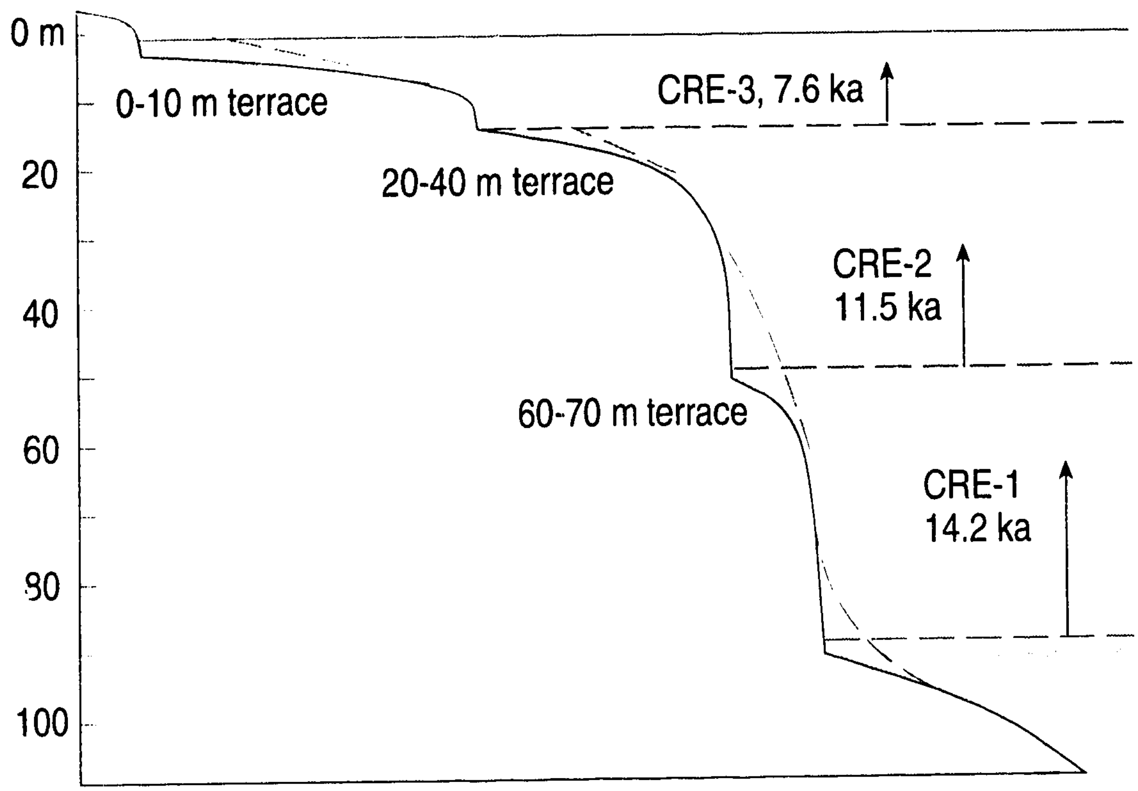

Figure 7.2. Terraced shelf and upper slope off Grand Cayman correlated with catastrophic steps in sea-level rise during the last deglaciation. Positions of slow rises or stillstands correspond with terrace positions and cliffs between terraces correspond with CREs (catastrophic sea-level rises).

(Fig. 7.2). Furthermore, the preservation of an intertidal notch in the slope breaks supports the rapid rate and large magnitude of the pulses (Chapter 4, Fig., 4.19).

Although the pulsed or stepped pattern of deglacial sea-level rise explains the elevation of drowned terraces and reefs, it does not account for reef configuration nor the considerable time lags in reef re-establishment following the sea-level pulses. The key to these problems lies in the interplay between the sea-level rise and coastal gradient. Along coasts with low gradients, marine erosion can rapidly cut wide terraces during slow sea-level-rise episodes (Chapter 4). If the terraces exceed ~300 m, detritus produced during hurricanes can accumulate and a fringing reef can immediately form (Chapter 5). With continued sea-level rise and vertical reef accretion, the width and depth of the reef lagoon gradually increases and the reef transforms from a fringing-type to a barrier-type architecture. Along coasts with higher gradients, a 300 m terrace takes considerably longer to develop and there is a longer time lag before reef establishment. Thus, reefs that established early on low-gradient slopes have a mature barrier-type configuration and those that established late on high-gradient slopes have a more youthful fringing-type configuration (Chapter 5). By mediating terrace width, therefore, local changes in coastal gradient can not only control the timing of fringing reef establishment but also reef configuration and architecture.

Controls On Shelf-Edge Reef Development

Although not identified around Grand Cayman, submerged-reef ridges have been reported on the upper slopes around many other Caribbean islands (for summary see Macintyre, 1972). To date, only ridges around Barbados have been investigated in any detail (Macintyre, 1967; Fairbanks, 1989; Macintyre et al., 1991). Along the leeward side of that island, two continuous ridges parallel the shore for some 20 km. The shallow ridge at ~15 m below msl is the modern shelf-edge reef with biotic and architectural characteristics very similar to the leeward shelf-edge reef around Grand Cayman (cf. Ott, 1975). The deeper ridge at ~70 m below msl is a relict shelf-edge reef that formed during a lower stand of sea level. Like its shallower counterpart, it has steep-sides, up to 15 m of relief, and is flanked on the shoreward side by a sand trough at ~80 m, and on the seaward side by a steep rubble slope down to ~100 m (Macintyre, 1967; Macintyre et al., 1991). Unlike its shallow counterpart, however, the 70 m ridge has a deep-water biota dominated by sponges and ahermatypic corals that overlie a substrate of muddy sand and algal nodules (Macintyre et al., 1991). Although the internal anatomy of this 70 m ridge on the leeward margin is not known, drilling on a slightly less prominent but equivalent 73 m ridge on the windward side of the island shows that it is formed of a 14-m thick framework of *A. cervicornis* and *M. annularis* underlain by a drowned *A. palmata* reef that ceased growing ~14 ka ago (core RGF-9, Fig. 1 in Fairbanks, 1989). This framework sequence suggests that the relict 70 m reef initiated and grew during the first step in deglacial sea-level rise (14.2-11.5 ka), when the sea rose rapidly from ~90 to a stillstand at 55 m (Chapter 3). This scenario is supported by the fact that during the 55 m stillstand, the top of [illegible] ridge would have been submerged by 18 m—roughly the same depth that the modern shelf-edge reef is presently submerged.

Macintyre et al., (1991) suggested that, instead of correlating with the 73 m reef, the 70 m ridge on the leeward margin correlated with a drowned *A. palmata* barrier reef at 43 m below msl (core RGF-12, Fig. 1 in Fairbanks, 1989) and consequently concluded that the ridge must be a drowned barrier reef. Yet the difference in depth between the tops of these reefs is 33 m—almost the limit of significant coral growth on modern reefs. If both were barrier reefs, as Macintyre et al. (1991) claim, they cannot have existed at the same time.

One possible reason for Macintyre et al.'s (1991) poor correlation is the in-grained idea that significant reef growth could not occur in waters deeper than about 15 m and therefore the shelf-edge reef and deeper ridges around Barbados and other islands must be drowned barrier reefs (Jukes-Brown and Harrison, 1891; Macintyre, 1967; Macintyre, 1972). Mounting evidence from underwater drilling shows, however, that shelf-edge reefs are not shallow-water structures but are dominated both internally and externally by a luxuriant deeper-water coral communities that have accreted at prolific rates (James and Ginsburg, 1979; Macintyre et al., 1982; Hubbard et al., 1986; Chapter 6). Such prolific growth and accretion in these supposedly prohibitive deep, outer-shelf areas is difficult to explain. Although attempts to do so have been few, one recurrent idea is that shelf-edge reefs are immature parts of the 'main' shallow-water fringing reef system that are rapidly trying to 'catch up' with sea level (Goreau and Goreau, 1973; Adey and Burke, 1976; Hubbard, 1988). Yet, as already alluded to, the presence of shelf-edge reefs in areas where fringing reefs are absent (leeward margins) and where they are widely separated, precludes this idea and suggests that they are independent structures (Chapter 6). But this still leaves unresolved the question of why a shelf-edge reef would preferentially grow around the edge of a shelf widely separated from shallow-water reefs. In other words, what controls the 'shelf-edge' configuration of modern shelf-edge reefs?

The solution to this problem relates to the deleterious influence of sedimentation on coral growth as documented by many reef studies (Marshall and Orr, 1931; Hubbard and Pocock, 1972; Loya, 1976; Lasker, 1980; Hubbard, 1986; Acevedo et al., 1989). Shelf-edge reefs are generally submerged below fairweather wave base and so develop in a zone of active carbonate sedimentation. Probably the most convincing demonstration of this has been made by Hubbard et al., (1990) who calculated that, of the skeletal carbonate produced by coral growth on the shelf-edge reef around St. Croix, 50 % was converted to sediment by bioerosion; half of this bioeroded sediment was trapped in the interior framework of the reef and the other half accumulated around the reef on the outer part of the shelf. From these calculations it was concluded that locally, around the reef, the rate of sediment accumulation could easily exceed that of reef accretion therefore redistribution or removal by physical processes was required to prevent the reef from being smothered. In a later study this physical removal mechanism was clearly identified when Hurricane Hugo (1989) flushed significant quantities of sediment from areas around the shelf-edge reef (Hubbard, 1992).

Sediment production and storm-induced redistribution clearly make outer shelf areas inhospitable places for young corals aspiring to be reef builders. Why then, do shelf-edge reef initiate in such locations? Two possibilities exist. First, any preexisting topographic highs would become ideal sites for coral settlement and growth. Such highs might include drowned karst-modified landforms preserved in areas where wave energy was insufficient to eradicate them (e.g., Lidz et al., 1991; Toscano et al., 1994), or they might be drowned depositional features such as siliciclastic deltas (e.g., Choi and Ginsburg, 1982) or shallow water *A. palmata* reefs (e.g., Adey et al., 1977). Second, and more important, corals would also preferentially develop in areas where sediment could not build up to any appreciable extent, such as breaks in slope. This has been demonstrated by Porter (1972) on Arcuadargana Island, Panama. Line transects on the shelf between 0-30 m displayed a significant increase in diversity (components, species number, and evenness) as breaks in slope where approached. Porter (1972) tentatively attributed this trend, which he termed the 'edge effect,' to sedimentation on flatter parts of the shelf although he had no supportive evidence. This 'edge effect' can also be seen from coral distributions around Grand Cayman (Fig. 1). Non surf-adapted coral growth preferentially develops over the mid-shelf scarp and the shelf edge escarpment (Chapter 4). However, corals that develop over the mid-shelf scarp are fronted by the sand covered lower terrace. During storms, therefore, these corals are surrounded by a blizzard of suspended sand and will experience significant stress. Coral that develop along the shelf break, however, are isolated from sediment stress because sand suspended during storms will be driven shoreward by wind currents and sand moving seaward in return flows will be channeled between areas of coral growth and consequently bypass the reef. Thus, shelf-edge reefs experience optimal growth conditions and, like a plant towards the light, coral growth is drawn upward and outward into the clear, sediment-free waters of the open ocean. The result—preferential growth of corals on the shelf edge and the development of a shelf-edge reef.

Conclusions

Although much valuable ground-breaking research on reefs has been carried out over the last 150 years, it is fitting and somewhat ironic that the conclusions on shallow-water reef configuration reached in this thesis bring us back full circle to the ideas of Charles Darwin, the father of modern reef study. For just as Darwin (1842) proposed, there is a genetic relation between fringing reefs, barrier reefs, and atolls. In fact I would go as far as to say that these structures are the same type of reef, controlled by the same types of processes, developing in the same way, and only differing in terms of maturity. Darwin's only mistakes were the assumptions he made about the types of processes that controlled reef configuration. Rather than being controlled by purely biological processes and simple coral growth, their position in shallow water meant that these reefs were subject to violent physical processes, such as hurricanes, on a regular basis. On human time scales such disturbances seemed uncommon and were therefore assumed to be relatively unimportant in reef development

(Graus et al., 1984). But on a geological time scale these processes were virtually continuous and therefore constituted a major controlling force (Woodley, 1992). This conclusion is supported by the investigation of the fringing reef around Grand Cayman (Chapter 5). For this study is the first to directly link modern reef development with hurricane induced processes; not only does it show that hurricanes control reef accretion, but it shows that they also determine where the reef will develop. Variations in their configuration and maturity simply reflect the interplay between sea-level rise and shelf gradient. changes in shelf gradient not only account for the temporal variation in reef initiation. but also the spatial variation in reef architecture and configuration.

Shelf-edge reefs, by contrast, have not been as widely studied, or even recognized as separate structures, so an assessment of the factors that control their configuration and architecture has never been made. As far as I know, this thesis is the first to identify shelf-edge reefs as separate entities and consequently the first to proffer any explanation of their architecture or configuration. It finds that, in complete contrast to their physically-dominated shallow-water cousins, the configuration of shelf-edge reef is controlled by the ecological requirements of reef-building coral. These requirements dictate that reef growth initiates only in areas where corals can remain relatively isolated from the build-up of sediment. The most favorable position is the very edge of the shelf. Not only is sediment prevented from building up there, but the sediment-free open ocean waters facing the reef provide optimal conditions for reef development. But this location has one major drawback — it faces the full force of hurricane-generated open-water waves. The destructive impact of these waves can significantly modify the architecture of the reef that develops. Yet, these waves also remove the vast amounts of sediment that threaten to eventually smother the reef. Consequently, the influence of hurricanes on shelf-edge reefs is like a double-edged sword, repeated pruning cuts reef growth back while at the same time ensuring new growth by removing sand.

An understanding of the processes that control the configuration and architecture of modern reefs around Grand Cayman inevitably allows me to make some speculations concerning

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Appendix A

Marine Site Locations Around Grand Cayman

North Side (Protected-Windward Margin)

N1	Hepp's Wall	19'23.15N	081'25.058W
N2	Hepp's Pipeline	19'23.13N	081'25.021W
BB	Boatswains Bay	?	?
SpB	Spanish Bay	?	?
N3	Ghost Mountain	19'24.16N	081'23.12W
N4	Bear's Paw	19'23.66N	081'21.63W
N5a	Blue Pinnacles (2 fingers)	19'23.64N	081'20.73W
N5b	Cliff Hanger	?	?
N6	Hole-in-the-wall	19'23.59N	081'20.69W
N7	Main Street	19'23.44N	081'20.60W
N8	Stingray City West	19'23.02N	081'20.52W
N9	Stingray City East	19'23.08N	081'20.59W
NIO	Tarpon Alley West	19'23.15N	081'20.16W
NII	Tarpon Alley East	19'23.13N	081'20.17W
N12	Princess Penny's Wall	19'23.07N	081'19.96W
N13	Black Forest North	19'23.05N	081'19.72W
N14	Eagle Ray Pass	19'23.03N	081'19.49W
N15	Lemon Wall	19'23.04N	081'19.34W
N16	Lemon Reef	19'22.975N	081'19.471W
N17	Leslie's Curl	19'23.05N	081'19.22W
N18	Hammerhead Hill	19'23.07N	081'18.57W
N19	3 B's Wall	19'23.01N	081'18.47W
N20	Dream Weaver Reef	19'22.91N	081'18.19W
N21	Chinese Wall	19'22.85N	081'18.03W
N22	Robert's Wall	19'22.83N	081'17.80W
N23	Gail's Mountain	19'22.81N	081'17.79W
N24	Haunted House	19'22.81N	081'17.64W
N25	Pinnacle Reef	19'22.77M	081'17.64W
N26	No Name Wall	!9'22.83ON	081'17.422W
N27	Queen's Thrown	19'22.81N	081'17.59W
N28	White Stroke Canyon	19'22.89N	081'17.28W
N29	Penny's Arch	19'22.60N	081'15.93W
N30	Andy's Wall	19'21.89N	081'15.17W
N31	Andy's Reef	19'21.83N	081'15.21W
N32	Della's Delight	19'21.52N	081'14.72W
N33	Grape Tree Wall	?	?

West Side (Leeward Margin)

W1	North West Point	19'22.13N	081'25.23W
W2	Bonnie's Arch	19'22.20N	081'25.16W
W3	Orange Canyon	19'22.18N	081'25.15W
W4	Sentinel Rock	19'22.06N	081'25.04W
W5	Big Tunnels	19'22.07N	081'24.15W
W6	Easy Street	19'22.07N	081'24.92W
W7	Black Hole (In Between)	19'22.06N	081'24.90W
W8	Dragon's Hole	19'21.99N	081'24.75W
W9	Big Dipper	19'21.96N	081'24.66W
W10	Little Tunnels	19'21.94N	081'24.60W
W11	Round Rock West	19'21.86N	081'24.49W
W12	Round Rock East	19'21.88N	081'24.47W
W13	Trinity Caves	19'21.90N	081'24.44W
W14	Neptunes Wall	19'21.81N	081'24.34W
W15	Slaughterhouse Wall	19'21.75N	081'24.21W
W16	Wall Street	19'21.77N	081'24.18W
W17	Sand Chute	19'21.71N	081'24.10W
W18	Doc Polson Wreck	19'21.67N	081'23.97W
W19	Mitch Miller Reef	19'21.67N	081'23.86W
W20	Marty's Wall	19'21.61N	081'24.OOW
W21	Knife	19'21.54N	081'23.96W
W22	Lost Treasure Reef	19'21.22N	081'23.61W
W23	Spanish Anchor	19'21.17N	081'23.61W
W24	Angelfish Reef	19'20.77N	081'23.51W
W25	Aquarium	19'20.70N	081'23.46W
W26	Great House Wall	19'20.60N	081'23.56W
W27	Killa Puffer	19'20.62N	081'23.45W
W28	Peter's Reef (Governor's)	19'20.56N	081'23.44W
W29	Jax Dax	19'20.47N	081'23.40W
W30	Eagle's Nest	19'20.47N	081'23.53W
W31	Paradise Reef	19'20.28N	081'23.38W
W32	Oro Verde North (Bow)	19'20.32N	081'23.44W
W33	Oro Verde South (Stern)	19'20.31N	081'23.44W
W34	Holiday Inn (Shark Hole)	19'20.27N	081'23.51W
W35	Hammerhead Hole	19'20.16N	081'23.35W
W36	Wildlife Reef	19'19.87N	081'23.39W
W37	Caribb Club Sand Chute	19'19.93N	081'23.57W
W38	Lone Star Reef	19'19.61N	081123.46W
W39	Mesa	19'19.27N	081'23.60W
W40	Royal Palms Ledge	19'19.12N	081'23.62W
W41	Pagent Beach Reef	19'18.39N	081'23.31W
W42	Soto's North	19'18.16N	081'23.13W

W43	Soto's Central	19'18.06N	081'23.14W
W44	Fish Pot Reef	19'18.OON	081'23.[illegible]W
W45	Balboa	?	?
W46	Eden Rock North	19'17.6[illegible]N	081'23.2[illegible]
W47	Eden Rock South	19'17.58N	081'23.24W
W48	Devil's Grotto North	19'17.55N	081'23.24W
W49	Devil's Grotto South	19'17.48N	081'23.30W
W50	Seaview Reef	19'17.48N	081'23.38W
W51	LCM David Nichol (Sunset Reef?)	19'17.41N	081'23.55W
W52	Armchair Reef	19'16.77N	081'23.64W
W53	Smith's Cove	19'16.52N	081'23.60W
W54	Eagle Ray Rock	19'16.43N	081'23.72W
W55	Blackie's Hole	19'16.30N	081'23.76W
W56	Black Forest	19'16.27N	081'23.73W

South Side (Exposed-Windward Margin)

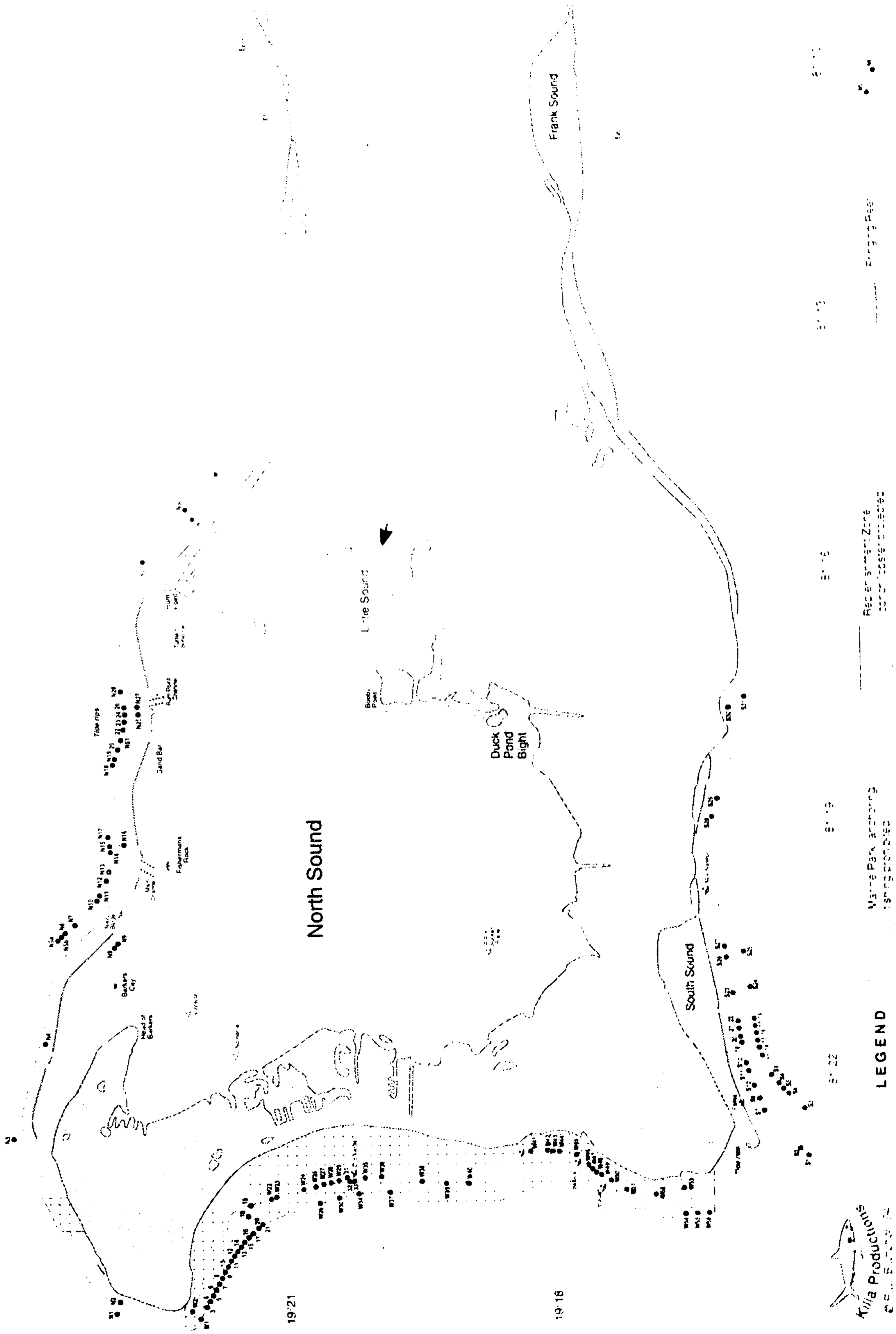
SI	Kent's Caves West	19'15.52N	081'23.18W
S2	Kent's Caves East	19'15.52N	081'23.15W
S3	Ron's Wall	19'15.58N	081'22.61W
S4	Pallas Pinnacle West	19'15.59N	081'22.60W
S5	Pallas Pinnacle Central	19'15.62N	081'22.52W
S6	Pallas Pinnacle East	19'15.66N	081'22.54W
S7	Pallas Reef West	19'15.75N	081'22.69W
S8	Pallas Reef East	19'15.77N	081'22.62W
S9	Christina's Wall	19'15.69N	081'22.44W
SIO	MG Square (Dede's Garden?)	19'15.83N	081'22.46W
SI I	Gary's Reef	19'15.87N	081'22.38W
S12	Pirates' Cove	19'15.86N	081'22.32W
S13	Crossroads	19'15.76N	081'22.30W
S14	Ollen's Office	19'15.75N	081'22.24W
S15	Eye of the Needle	19'15.79N	081'22.17W
S16	Phantom Ledge	19'15.82N	081'22.06W
S17	Gary's Wall	19'15.82N	081'21.99W
S18	Red Bay Caves	19'15.95N	081'21.99W
S19	No Name Wall	19'15.83N	081'21.93W
S20	Lauras	19'15.93N	081'21.75W
S21	Bullwinkle West	19'15.92N	081'21.69W
S22	Bullwinkle East	19'15.94N	081'21.63W
S23	Ned's Tunnels	19'15.968N	081'21.14OW
S24	Dangerous Dans Dropoff	19'16.OON	081'21.05W
S25	Barracuda Ron's Pass	19'15.93N	081'20.94W
S26	Japanese Gardens West	19'16.07N	081'20.86W
S27	Japanese Gardens East	19'16.08N	081'20.74W

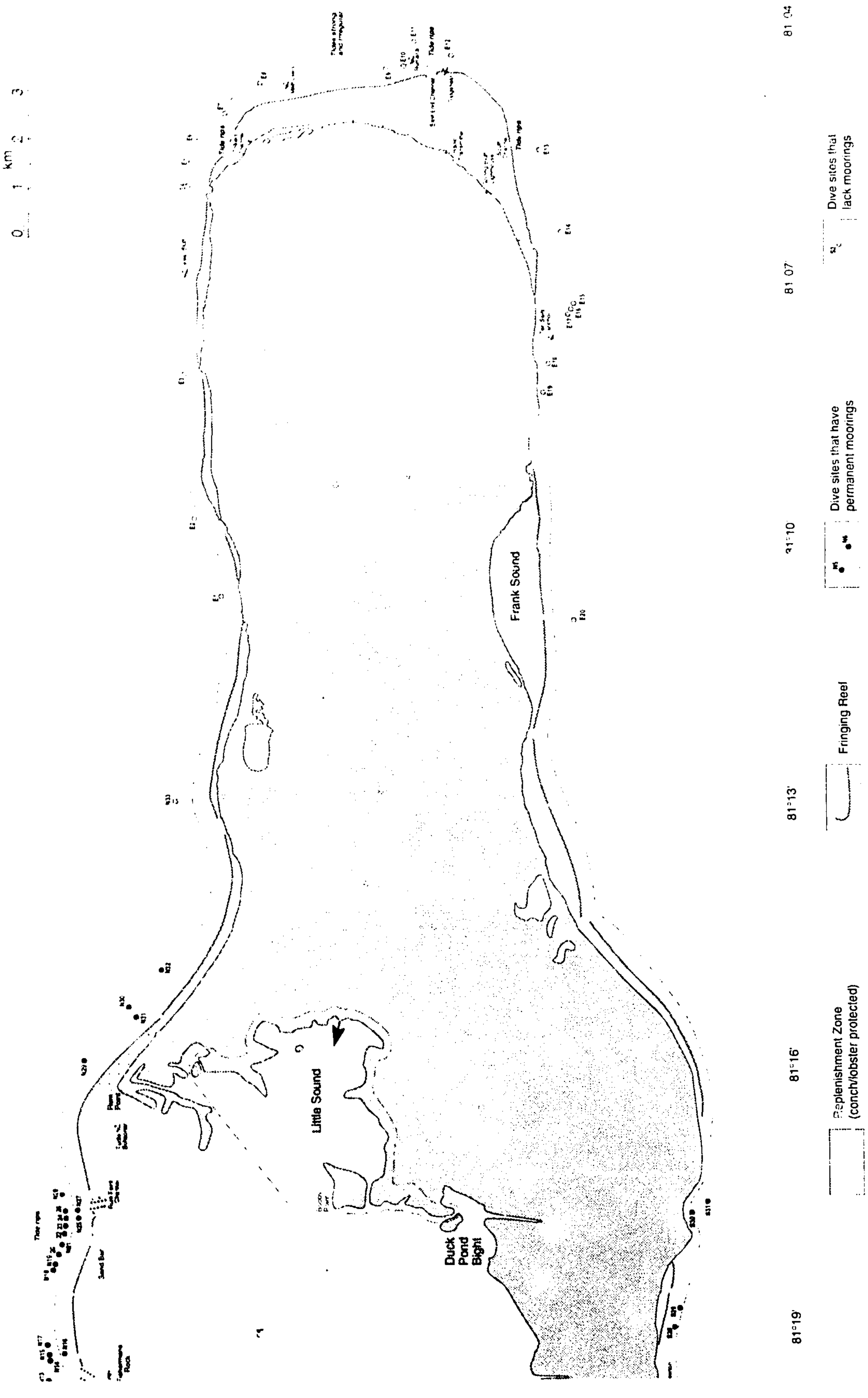
S2?	Spotts Ree[illegible]	19'16.11N	081'18.69W
S29	Bat Cave R	19'16.09N	081'18.59W
S30	[illegible]edro's Reef	19'16.04N	081'17.68W
S31	Ped[illegible] Pinnacle	19'15.80N	081'17.51W

East Side (Exposed-Windward Margin)

E1 Darby's Fantasy
E2 [illegible]le-in-the-Wall
E3 Babylon
E4 Black Rock
E5 Cinderella's Castle
E6 Snapper Hole
E7 Stag Party
E8 Grouper Grotto
E9 Vertigo Wall
E10 Ridgefield Reef (Shark Alley)
E11 The Maze
E12 Baffel-Sponge Slope
E13 Three Sisters
E14 River of Sand
E15 The Chimney (Love Tunnel)
E16 Ironshore Gardens
RW Roger's Wreck
TC Tortuga Club
CB Collier's Bay
SB Sand Bluff
GB Gunn Bluff
EP East Point

East Side sites are at the time of writing are all without moorings


North Sound
Little Sound
Frank Sound
South Sound
Duck Pond Bight
Sand Bar
Fishermans Rock
Barkers Cay
Head of Barkers
Tide rips
Booby Point
LEGEND
19°21
19°18
81°22
81°19
81°16
Killa Productions



0 1 2 3
km
81°04'
81 07'
81°10'
81°13'
81°16'
81°19'
Frank Sound
Little Sound
Duck Pond Bight
Sand Bar
Replenishment Zone (conch/lobster protected)
Fringing Reef
Dive sites that have permanent moorings
Dive sites that lack moorings

Appendix B

HISTORICAL ACCOUNTS OF HURRICANES [illegible] AYMAN

1731 Sea breached over Newlands (ACR).

1735 September: (Doran 1953 in Smith 1981; ACR).

1751 "In a remarkable hurricane on September 1751 the sea made a breach over the island from Little Pedro Point to the North Sound— the dotted line [on the map] shows the bed it left" from a notation on George Gauld's map of Grand Cayman 1773.

1785 August: "so terrific in its force that it tore up all but one tree at S. W. Point. After the recession of the tidal wave one could have walked from SW Point to West bay due to lack of vegetation" (Williams).

1793 population recorded as still struggling to recover in February 1794 (ACR).

1812 reported but no records (Williams).

1826 reported but no records (Williams).

1812 reported by Doran 1953 in Smith 1981 (orig ref not checked).

1835 reported by Doran 1953 in Smith 1981 (orig ref not checked).

1836 "...there must have been a tidal wave if there is any truth in the story that an inhabitant of Boddentown found a Grouper in his water pot the morning after."(Hirst).

1838/7(?) 25th Oct."we, suffering from two violent Hurricanes on the 28th Sept and the 25th October, have now to entreat the sympathy and charitable consideration of your Magesty's Government. By these visitations, St. George Church...has blown down, the other seriously injured; out of the 18 vessels belonging to the island and by which the inhabitants draw their priciple means of support, 13 have been wrecked; every plantation and provision ground utterly destroyed; and unless Chritian sympathy be awakened and Chritian benevolence be extended many of the inhabitants will be involved in the deepest and bitterest distress" From a memorium written to the Queen of England by the Custos, Magistrates and Inhabitants of the Cayman Islands (Hirst Notes p. 177; ACR).

1846 October (Passed South); Stormy NE winds blew in preceeding week. Winds switched rapidly to SE. Low areas on the south coast (Newlands and Prospect) breached and inundated (50 ton schooner could float!), destroying vegetation, roads and depositing rubbish, fish; loose stones piled up between Prospect Point and Spotts Bay. All wells in the centre of the island filled with salt-water, and for a year the only pottable water was in a Boddentown well. South Side suffered most damage. Muddy sea preceeded Hurricane.(Hirst; ACR).

1876 October (Track ?WNW; eye crossed island centre). Heavy squalls from the SE gradually moving around to the SW. All buildings suffered extensive damage but most damage was sustained on the south-side of the island. Foreshore covered with debris and uprooted trees. All vessels (15 schooners) in the different coves and anchorages were either driven ashore or broken up. (Hirst Notes p. 278: ACR).

1895 August: (Track WNW; eye passed 40 km to south) (NCC).

1896 September: (Track NW; eye passed over Georgetown) (NCC).

1899 October: (Track NNE; eye crossed island centre) (NCC).

1903 August: (Track WNW; eye crossed island centre); Winds started from the NE and moved around to the SE. Sea was not rough at all. North side houses suffered most destruction. Georgetown also badly hit. Wind blew with greater force but did not last as long as those in the 1876 Hurricane. In an attempt to aviod damage to island shipping all schooners, except one which broke her moorings and dissappeared, put to sea as the storm grew in intensity. Only two returned after the Hurricane; the remainder were never seen again (Hirst).

1909 September: (Track NW; eye passed 8 km west of Georgetown). "...new roads were blocked by rocks and trees." (Williams).

1910 October (Track WNW; eye passed 40 km to north). Heavy squalls from the SE gradually moving around to the SW. Extensive damage to property, most severe damage occurred at East End and the South Side where the sea rose up to 15' washing away roads and depositing cobble/boulder ridges. Red Bay and Spotts roads washed away and sea encroached 50 yds onshore bringing rocks and boulders depositing them as a breakwater. At Newlands a school of Jack and a Kingfish were washed inland ~500 m. The Norwegian Barque 'Pallas' was wrecked on South Sound reef. Seas far surpassed the 1846 storm. (Hirst).

1912 October: (Track W; eye passed 40 km to north) (NCC).

1912 November: (Track N; eye passed 8 km to east) (NCC).

1914 September: Sea flooded plantations, Spotts road obliterated (ACR).

1915 September: (Track NW; eye crossed island centre) (NCC).

1916 August: (Track WNW; eye 8 km north of East-side) (NCC).

1917 September: East End severely damaged, 2 deaths and 14 schooners washed ashore (ACR).

1932 Cayman Brac November: (200 mph winds) "The sea swept high over the coast [tides 20 feet high], carrying huge rocks on its crest and the wind hurled rocks, some weighing tons, through the air... Everything lay buried beneath a mass of broken coral, boulders, trees and the dismal reckage of houses." (Williams).

1932 This Hurricane also affected south-side of Grand Cayman and sea breached at Prospect (Hirst; ACR).

1933 Sea breached at Prospect.

1935 Cayman Brac (Williams).

1939 November: (Track E; eye crossed island centre) (NCC).

1944 August: (Track WNW; eye passed 40 km to south) (NCC).

1944 October: (Track W; eye passed 8 km to north; NCC) Red Bay and Prospect flooded; 760 mm rain in 4 hours (ACR).

1948 September: (Track NNW; eye passed over Georgetown) (NCC).

1951 October: (Track NNE; eye passed 40 km to west) (NCC)

1955 September: (Track WSW; eye passed 8 km to north) (NCC).

1969 August: (Track WNW; eye passed 40 km to west) (NCC).

1988 September: Gilbert (Saffir-Simpson category 2). Passed close; sea breached at Red Bay and Pedro; coastal dwellings washed out at East End and North West Point; extensive mangrove damage. Sustained winds 31 m s-1 (bp 976.5 mb) 150 cm surge, 3.4 m waves Note Little Pedro point is 6 m above msl. (Track WNW; eye crossed island).

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Appendix C

PRESENTATION ABSTRACTS

TOWARDS AN IMPROVED TEXTURAL CLASSIFICATION OF CARBONATES.

BLANCHON, PAUL, Dept. of Geology, Univ. of Alberta, Edmonton, Canada, T6G 2E3.

Schemes used to classify the texture of carbonates – especially reefal deposits – are subjective, imprecise and incomplete. To address these problems, Dunham's classification has been extended by adding categories that describe the shape and mutual arrangement of reef-building organisms, and modified by using prefixes to describe grain-size variation.

In order to classify deposits composed of in situ colonial organisms using this scheme, colonies are grouped into one of four general shape categories: 1) Headstones, equant or massive colonies; 2) Branchstones, stick-shaped to dendritic colonies; 3) Sheetstones, thin lamellar forms; 4) Platestones, thicker lenticular colonies. Once a reefal deposit is assigned to one of these categories, its name is modified according to the mutual relationship between colonies; the term 'sparse' is used where colonies are matrix supported (e.g. sparse platestone), whereas 'compact' is used for colony-supported frameworks (e.g. compact headstone). Frameworks with more complex or indeterminate colony arrangements are placed in the general shape categories and modifiers are not used. A second order modifying term that describes the matrix texture can also be added to the name. Hence, a sparse headstone with a grainstone matrix becomes a 'grainy sparse-headstone.'

To classify detrital deposits, Dunham's original texture categories (mudstone, wackestone, packstone and grainstone) are used in conjunction with prefixes that describe grain size. The prefix 'mega' is used to describe deposits dominated by pebble to boulder-sized grains (e.g. mega-packstone); 'meso' for deposits dominated by fine sand to granule-sized grains (e.g. meso-grainstone); and 'micro' for deposits dominated by medium silt to very fine sand-sized grains (e.g. micro-wackestone). In poorly-sorted deposits, where more than one grain size dominates, no modifying terms are used.

Abstract 1992; Geological Society of America. Abstracts with Programs, v. 24, #7.

An Improved Textural Classification of Carbonates.

Blanchon, Paul, Dept. of Geology, Univ. of Alberta, Edmonton, Canada, T6G 2E3.

Carbonate classification schemes presently in use are subjective, imprecise and incomplete. These problems are particularly acute in reef/mound deposits because of the textural complexity produced by variations in metazoan shape and arrangement. To address these problems, Dunham's classification scheme is extended by adding categories that describe the shape and mutual arrangement of reef-building organisms, and modified by using prefixes to describe allochem-size variation in detrital deposits.

In order to classify deposits composed of in situ colonial metazoa using this scheme, colonies are grouped into one of four basic shape categories: 1) Headstones are dominated by equant or massive colonies; 2) Branchstones have stick-shaped to dendritic colonies; 3) Sheetstones consist of thin lamellar forms; 4) Platestones contain thicker lenticular colonies. Once a reef/mound deposit is assigned to one of these categories, its name is modified according to the mutual relationship between colonies; the term 'sparse' is used where colonies are widely spaced and clearly matrix supported, (e.g., sparse platestone), whereas 'compact' is used for completely colony-supported frameworks (e.g., compact headstone). Frameworks with spatially variable or indeterminate colony arrangements are placed in the general shape categories and modifiers are not used. A second order modifying term that describes the matrix texture can also be added to the name. Hence, a sparse headstone with a grainstone matrix becomes a 'grainy sparse-headstone'.

To classify detrital deposits, Dunham's original texture categories (mudstone, wackestone, packstone and grainstone) are used in conjunction with prefixes that describe allochem size. The prefix 'mega' is used to describe deposits dominated by pebble to cobble-sized allochems, (e.g., mega-packstone); 'meso' for deposits dominated by fine sand to granule-sized allochems, (e.g., meso-grainstone); and 'micro' for deposits dominated by medium silt to very fine sand-sized allochems (e.g., micro-wackestone). Original Dunham names without modifiers can be used to denote heterogeneous deposits characterized by rapid spatial changes in allochem size.

Abstract, 1993. Geological Association of Canada annual meet, Edmonton, May 17-19.

Active Shelf-Edge Reefs Around Grand Cayman, (British West Indies); A New Analogue for Ancient Reefs.

Blanchon, P, and Jones, B, Dept of Geology, Univ. of Alberta, Canada. T6G 2E3.

Shelf-edge reefs are common features of modern carbonate shelves but are considered to be inactive, drowned reefs that formed during a lower stand of sea level. Consequently they have been regarded as a subordinate part of the inner-shelf fringing reef, (the so called 'fore-reef' zone). Our investigation of an active shelf-edge reef around Grand Cayman demonstrates, however, that this reef is a significant and separate buildup that has developed independently from the fringing reef system.

The shelf-edge reef around Grand Cayman is developed along the perimeter of a submerged shelf-terrace. [illegible] up [illegible]0 meters thick. It completely encircles the island, including the leeward margin, where fringing reefs are absent. The reef is composed of an array of coral-armored buttresses that are aligned perpendicular to shore, and regularly dissected by steep-sided, sediment-floored canyons. Individual buttresses consist of three morphological zones; the reef-front zone, characterized by a fauna of platy corals, is a steeply sloping to overhanging escarpment extending from 70 to 25 meters. At the apex of this escarpment is the reef-top zone, which extends into waters as shallow as 15 meters. This zone is characterized by a diverse fauna of broad, head-shaped corals on storm-facing margins, and branching forms on storm-protected margins. Behind the reef top is the back-reef zone which is characterized by the development of shoreward-projecting coral spurs. These morphologically-zoned buttresses are separated from the fringing reef by an extensive bedrock terrace that, in many areas, is completely devoid of coral growth.

The shelf-edge reef around Grand Cayman is a separate buildup with several morphological characteristics that make it distinct from fringing reefs. First, shelf-edge reefs develop below fair-weather wave base, and consequently their frame builders are not surf/wave resistant. Second, they have a diverse and morphologically-zoned biota. Finally, the reef is a thick, and predominantly vertical buildup. These characteristics are similar to those documented from many ancient reefs that do not fit into the classical surf/wave-resistant reef model, (e.g., the Permian Capitan reef). We propose therefore, that shelf-edge reefs may provide an alternative analog for these types of ancient reefs.

Abstract, 1993. Geological Association of Canada annual meeting, May 17-19.

Catastrophic Sea-level Rise During Deglaciation; Evidence from Submerged Terraces and Glacial Landforms.

Blanchon, Paul. Dept of Geology; Shaw, John. Dept of Geography, University of Alberta, Edmonton, Canada. T6G 2E3.

Sea-level rise during deglaciation is thought to have been either smooth and sigmoidal, or a two-step rise with an intervening phase of slower rise. The data upon which these curves are based, however, lacks the resolution to identify short-duration melting events which were probably common during this time. Although we do not offer a detailed sea-level curve, we present two independent lines of evidence that strongly suggest sea level during deglaciation was characterized by extremely rapid rises (5-10 meters in less than 200 years) and prolonged periods of stillstand.

The first line of evidence is related to the origin of several glacial landforms, including extensive fields of drumlins, giant flutings, tunnel channels and scoured bedrock tracts which are being increasingly recognized in North America and Europe. These landforms were formed either by depositional or erosional processes during regional-scale meltwater mega floods. These catastrophic floods released huge volumes of meltwater stored beneath the ice sheets, and conservative estimates of flood discharges suggest that individual mega-flood events may have caused an almost instantaneous sea-level rise of about 30 cm. Several closely associated events of this magnitude from North American and European ice sheets could have easily raised sea level by several meters in a very short time.

Further evidence for rapid meter-scale sea-level rise, and new evidence for prolonged phases of stillstand, comes from submerged terraces and intertidal notches present on island and continental shelves. Numerous islands, known to have been stable during the late Pleistocene, show a coincident development of submerged terraces; a modern shallow terrace has developed during the present stillstand, and a deeper terrace was produced during an earlier period of stillstand. Separating them is a drowned sea cliff with a prominent and extremely well preserved intertidal notch. The preservation of this feature, which is developed at the same depth on at least two stable islands, suggests a rapid sea-level rise of 5-10 meters. This rise must have been rapid enough to prevent any significant erosion of the notch, and calculations based on coastal erosion rates suggest that the rise occurred in less than 200 years. These rates also suggest that the associated terraces formed over a time span of a few thousand years, during which time sea level essentially remained stable.

Compelling evidence from both glacial landforms and drowned terraces/notches suggests that deglaciation that was characterized by catastrophic sea-level-rise events interspersed with prolonged periods of stillstand. Although many ends remain untied, we believe this is a plausible and realistic model of deglaciation. Its implications are profound. It could explain the mythical accounts of catastrophic floods and drowned continents recurrent in several ancient cultures and, more seriously, it could predict the nature of future sea-level changes in a warming world.

Abstract, 1993. Geological Association of Canada annual meet, May 17-19.

Catastrophic Sea-Level Rise and Glacial Ice-Sheet Collapse: The Megaflood Triggering Hypothesis.

Blanchon, Paul. Dept. of Geology, **Shaw, John.** Dept. of Geography, University of Alberta, Edmonton, Canada. T6G 2E3.

Theories that attempt to unravel deglaciation are unable to explain why it was triggered by northern summer insolation trends. This weakness is compounded by recent discoveries of extremely rapid changes in the ocean/climate system during deglaciation. We propose that these enigmas are explained by the episodic and catastrophic release of huge volumes of meltwater stored beneath the Laurentide ice sheet.

Evidence for this link between the ocean/climate system and meltwater megafloods comes from the positions of drowned *Acropora palmata* reefs and marine-planation terraces in the Caribbean-Atlantic reef province. These demonstrate that deglacial sea-level rise had a distinct staircase geometry, with individual steps characterized by an initial catastrophic-rise event that drowned reefs and terraces, followed by an episode of slower rise, during which terraces and reefs re-established in upslope positions. Using existing U-Th and radiocarbon dates, three catastrophic rise events have been constrained; an 11 m rise event occurred at 13.5 ka, a 7 m rise event at 11.4 ka, and a 7 m rise at 7.2 ka. Evidence from reef growth rates suggest that each rapid rise-event took less than a century to occur and probably lasted only a few months.

Such catastrophic sea-level-rise events were probably caused by the release of meltwater megafloods from beneath the Laurentide Ice Sheet. These produced regional-scale depositional and erosional landforms across much of Canada and northern United States. In our hypothesis, the first megaflood, and the instantaneous sea-level-rise it produced, caused a worldwide ungrounding of marine ice-sheets and triggered glacial downdraw. Disintegration of ice-shelves and the discharge of massive fleets of icebergs into the north Atlantic by ice-streams (Heinrich events), prevented the northward drift of warm equatorial currents and caused the rapid northern hemisphere cooling associated with the Younger Dryas.

This megaflood triggering hypothesis not only explains recent discoveries but solves several outstanding problems: global ice-sheet collapse was synchronous and rapid because of the de-stabilizing effect of megafloods on marine ice-sheets; northern-hemisphere insolation trends drive glacial cycles because only northern ice-sheets were bordered by large land masses where meltwater could be stored; the storage and release of meltwater from subglacial reservoirs exerted a major control on the timing of rapid climatic and oceanic perturbations during deglaciation.

Abstract 1993; Geological Society of America. Abstracts with Programs, v. 25, #6.

PETROGRAPHIC CLASSIFICATION OF BIOTURBATE TEXTURES IN THE MISSISSIPPIAN MIDALE CARBONATES OF THE WILLISTON BASIN, S. SASKATCHEWAN.

A.D. KESWANI, M. BLANCHON AND S.G. PEMBERTON, Ichnology Research Group, Department of Geology, University of Alberta, Edmonton, Alberta T6G 2E3

A complete characterization of carbonate rocks includes both the identification of trace fossils and the classification of bioturbate textures. Although many workers have described general ichnologic aspects of carbonates as "bioturbated," "burrowed," "mottled," or "nodular," they have ignored petrographic characterization. Without this, the influence of biotic processes and the fabrics that are produced are not fully recognized.

To address these problems, we have designed a descriptive, petrographic classification of burrow fabrics on the basis of grain-size selection, and the alignment of skeletal grains. Using these respective criteria, two types of fabrics, distinct and indistinct, are recognized in thin sections: distinct burrow fabrics have well-defined walls with a sharp contrast between the burrow fill and surrounding host sediments. Three types of fill are identified on the basis of grain-size: microfills have a smaller grain-size than the host, mesofills are the same size as the host, and, megafills are larger than the host. Indistinct burrow fabrics are defined on the basis of alignment of arochems and their arrangement. Three alignment types are recognized, tangential, planar, and concentric, and these can have either a dense or a loose spatial arrangement. Once burrow fabrics have been categorized, the mutual arrangements between burrows can be defined. On the basis of burrow abundance and cross-cutting relationships, four types are identified: isolated burrows, clustered non-interpenetrating burrows, clustered interpenetrating burrows, and homogenized burrows.

This descriptive classification can be used to interpret the biotic processes involved in fabric genesis. Processes are largely related to disturbance or selective removal of particular grain-size classes during various biological activities including sediment ingestion, burrow construction, movement through the sediment, or the competitive and coprophagous interaction between benthos.

Abstract, 1995. Geological Association of Canada annual meet, May 17-19

The Beginning of the Holocene

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The beginning of the Holocene marks changing climatic conditions from glacial to interglacial and was characterized by remarkably rapid variation in both climate and ice sheets. Sea level rose in steps and ocean circulation patterns were significantly reorganized at this time. These changes are recorded in preserved floras and faunas (e.g. in pollen records, in foramanifera and their isotopic composition). The effects of rapid sea-level rise are preserved in drowned coral reefs and submerged marine platforms. A pronounced, catastrophic sea-level rise, beginning at about 11 500, has been constrained by drowned *Acropora palmata* reefs. This sea-level rise marks the end of the last widespread cold event (the Younger Dryas) of the Pleistocene glaciations. Its effects, submerged shorelines, abandoned barrier beaches, and drowned reefs, should, therefore, serve as time markers. We propose that this catastrophic sea level rise (CRE) was related to outburst floods from the conti[illegible]tal ice sheets of the Northern Hemisphere. Implications of these floods for sediments [illegible]e ocean bed have yet to be investigated. They may serve as extensive marker horizons of flood events where they extended to the oceans.

The proposed date of 10 000 years B.C. for the beginning of the Holocene may be linked to these events. Evidence for outburst floods and CRE's is expected to be widespread. This evidence is expected to provide excellent reference markers for placing the beginning of the Holocene.

Abstract, 1995. American Geophysical Union Spring Meeting. May 30th to June 2nd.

Catastrophic Sea-level Rise & Reef Drowning 8 ka Ago: An Omen?

Blanchon, Paul. Department of Earth & Atmospheric Sciences, University of Alberta, Canada, T6G 2E3.

Approximately 8 ka ago, fast-growing elkhorn (*Acropora palmata*) reefs across the Caribbean—from Barbados to Florida—mysteriously died off. Yet in as little as 100 years, these reefs had re-established in new positions some 7 m higher up slope. This rapid backstepping demonstrates that the widespread die-off did not result from local deterioration of environmental conditions, but was caused by catastrophic rise in global sea level. This catastrophic rise shifted elkhorn reefs from the surf zone to deeper water so quickly that it prevented elkhorn recovery both during, and after the rise. Subsequent colonization of the drowned reefs by deep-water corals therefore occurred without development of a mixed elkhorn association. Calculations show that for this to occur, the rise rate must have exceeded 45 mm/yr. Elevation differences between the drowned reefs and the new up-slope reefs show that the rise had a magnitude of 6.5 (±2.5) m. Only one process is capable of producing such a rapid and large magnitude rise—ice-sheet collapse.

Similar catastrophic rises drowned elkhorn reefs during earlier stages of deglaciation when ice sheet instability was to be expected. But the rise 8 ka ago was different. By that time, ice sheets had all but disappeared from the Northern Hemisphere and the only significant area of potentially unstable marine-based ice lay over Antarctica. Then, as now, the western half of this ice sheet was probably buttressed by floating ice shelves that were pinned precariously on bedrock pinnacles. With such an unstable configuration and, in the absence of other sources, it seems likely that the collapse of the West Antarctic Ice-Sheet ~8 ka ago produced a catastrophic rise in global sea level that drowned fast-growing reefs in the Caribbean.

This raises the disturbing possibility that the Antarctic ice sheet is inherently unstable and may collapse again—a warning issued by John Mercer almost 18 years ago. Although the 8-ka collapse was probably initiated by the sudden release of meltwater from the last of the glacial lakes, any similar process affecting the underpinnings of the ice shelves could have the same disastrous effects in the future. In this light, dramatic increases in sea-surface temperature recently reported from the southern oceans are of particular concern. While Mercer predicted that such manifestations of global warming might cause collapse he had no evidence that collapse had occurred previously. Now that the significance of drowned elkhorn reefs has been recognized, Mercer's predictions of catastrophic sea-level rise are much more ominous.

Abstract, 1995. 1St SEPM Congress on Sedimentary Geology. August 13-16

Shelf-Edge-Reef Architecture Controlled by Hurricanes

Blanchon, Paul. Department of Earth & Atmospheric Sciences, University of Alberta, Canada, T6G 2E3.

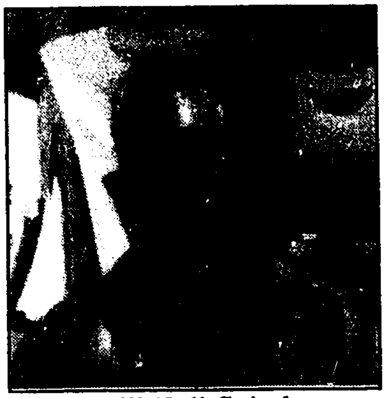
Rimming the outer shelf of Grand Cayman is a submerged, 87 km-long shelf-edge reef (SER) that rises to within 12 m of mean sea level. It consists of an array of coral-armored buttresses aligned perpendicular to shore and separated by steep-sided sediment-floored canyons. Individual buttresses have a diverse biota and consist of 3 architectural elements: a shield-like front wall colonized by platy corals, a dome-shaped crown colonized by robust head corals, and a shoreward-projecting spur covered by branching corals. Buttresses are commonly fronted by coral pinnacles in various stages of development. In some areas these have amalgamated with buttress walls to produce pinnacle-and-arch structures.

As margin orientation changes, SER architecture varies systematically in response to changes in the height and fetch of hurricane waves. Along exposed margins, fully developed storm waves cause extensive fragmentation of weak corals and remove significant quantities of sand from around buttresses. This produces extensive, large-amplitude buttresses with deep crowns and spurs dominated by robust head corals. Along semi-protected margins, reduced storm-wave height cause less fragmentation and sand removal thereby producing a shallower SER with intermediate-amplitude buttresses and a coral fauna co-dominated by branching and head corals. Along fully protected margins storm waves have the least impact and normal buttresses-canyon architecture breaks down producing a series of shallow, low-amplitude, branching-coral-dominated ridges that merge laterally into an unbroken belt of coral. On each margin, local changes in SER architecture are related to the angle of hurricane-wave approach. SER sections facing into storm-waves are pruned of weak branching corals and the fragments are swept back onto the shelf producing extensive spurs. On obliquely oriented sections, storm-waves wash the debris along and off shelf producing little or no spur development. Instead, the debris accumulates in front of the buttress walls and initiates the development of coral pinnacles.

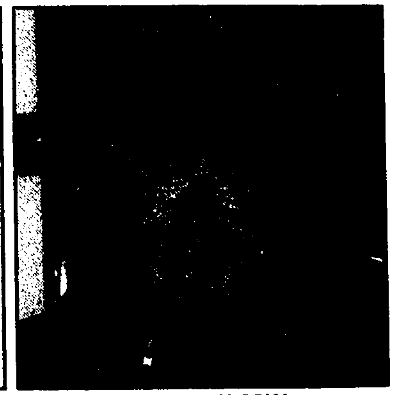
Over time, repeated buttress pruning and shedding during hurricanes not only controls reef architecture but also influences accretion patterns. Vertical accretion is limited by the effective depth of storm-wave fragmentation. Once this hurricane-accretion threshold is reached the reef accretes laterally via pinnacle growth, amalgamation, and infilling. Consequently, the reef rapidly steps-out over its own debris in a kind of balancing act between lateral growth and slope failure—a pattern widely recognized in ancient reefs.

Abstract, 1995. 1St SEPM Congress on Sedimentary Geology. August 13-16

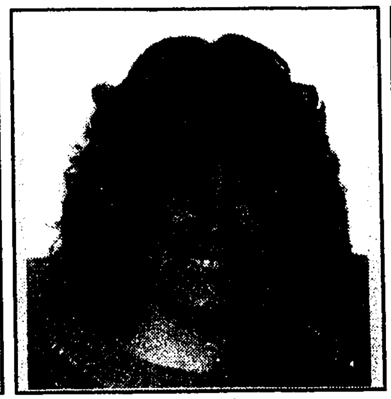
Rogues Gallery


Bill Kalbfleisch


Brent Wignall


Dave 'toast!' Hills


Ian Hunter


Jen Vézina


Leo 'piggoli' Piccoli

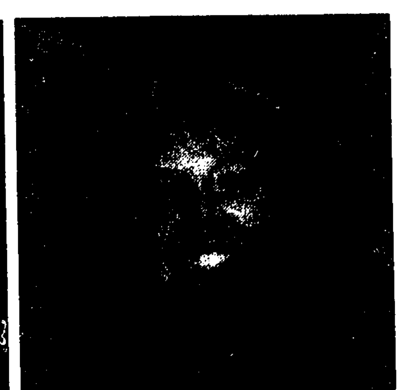

Chun Li


Jason 'Tex' Montpetit


Kevin Brett


Howard Brekke


Mark Hearn


Karen Patey (tanks for the buddies Karen!)