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

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ABSTRACT

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The shelf around Grand Cayman consists of two seaward-sloping terraces separated by a mid-shelf scarp. Except along the exposed windward margin where coral growth is dominant, the upper terrace (0–10 m bsl) largely consists of a barren rocky pavement traversed by erosional furrows. Exposure related trends in the morphology and distribution of these erosional features, and the lack of coral growth, demonstrates that the terrace is the result of contemporary erosion during seasonal storms. The upper terrace is terminated by a mid-shelf scarp (10–20 m bsl) that, in most areas, is partially to completely buried by modern carbonate deposits. Along narrow sections of the leeward shelf, the scarp is commonly exposed and displays an erosional intertidal notch at −18.5 m. The lower terrace (12–40 m bsl) extends from the mid-shelf scarp to the shelf edge. Its surface is a modern reef-and-sediment wedge that thickens toward the shelf edge, reaching up to 40 m in thickness. These deposits are underlain by a seaward-sloping bedrock terrace (20–40 m bsl). This buried terrace and the mid-shelf scarp, which are geomorphic equivalents of the upper terrace and coastal cliff, represent an earlier episode of marine planation when sea-level was stabilized at a lower position.

The contemporary erosional features of the upper-shelf terrace and the presence of identical terraces around recently uplifted islands demonstrates that the terraces on Grand Cayman were sculptured by marine erosion during the last deglacial sea-level rise. The lower terrace and the mid-shelf scarp were eroded during a slow-rise episode from 11–7 ka and were subsequently drowned by an extremely rapid, 5 m rise-event at ~7 ka. Following this catastrophic event, which drowned fast-growing *Acropora* reefs in other areas of the Caribbean, sea-level stabilized and rose slowly to its present position, producing the upper terrace. This pronounced stepped pattern in Holocene sea-level rise remains to be confirmed from outside the Caribbean-Atlantic reef province but is consistent with the stepped nature of pre-Holocene sea-level curves.

The presence of seaward sloping terraces on many shelves around the world suggests that erosional terrace cutting is a common phenomenon during sea-level rise. In contrast, terraces in areas that have undergone relative sea-level fall are constructional in origin, produced entirely by reef accretion. This suggests that there is a genetic relationship between the sea-level cycle and terrace type: erosional terraces form during rise and constructional terraces during fall.

ADDITIONAL INDEX WORDS: *Coral reef, terrace, coastal cliff, marine erosion, sea level, continental shelf, submarine terrace, coastal erosion, spur-and-groove, ridge-and-furrow.*

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 construc-

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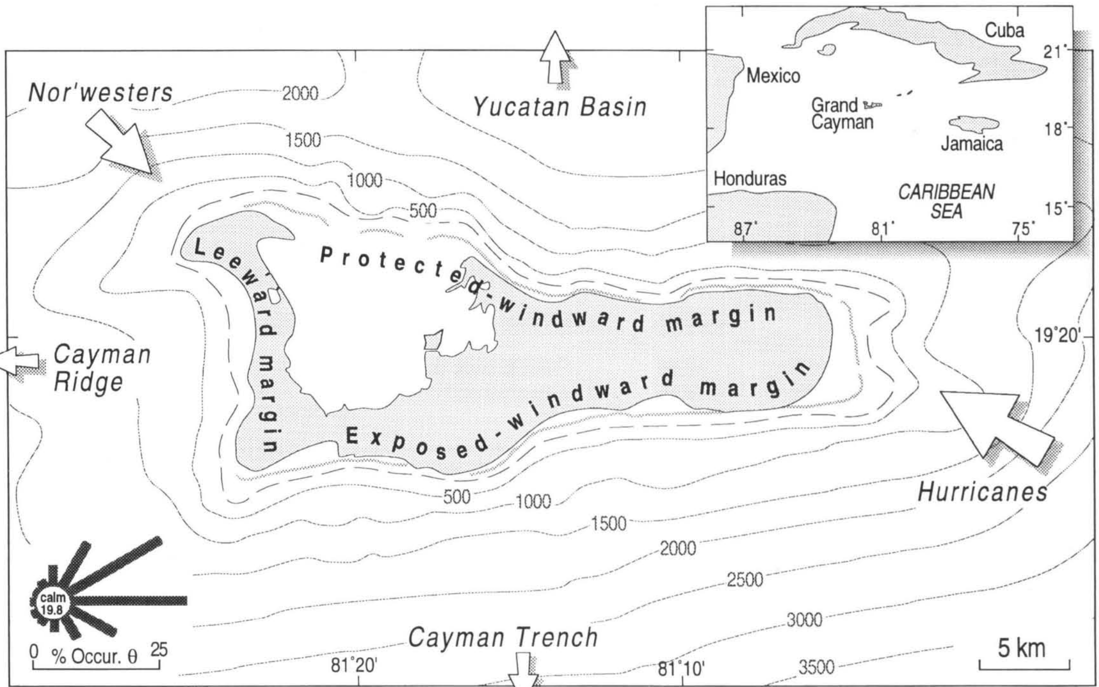

Figure 1. Grand Cayman. Location, bathymetry in metres (after LeBlanc, 1979), extent of shelf (dashed line), position of fringing reefs (shaded lines), and wind/storm directions (from Darbyshire *et al.*, 1976). Shelf divided according to wave regime; exposed-windward margin (south and east sides), protected-windward margin (north side), and leeward margin (west side).

tional 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 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 (Figure 1). By integrating data from submarine drill cores, SCUBA transects, depth-sounding profiles, and seismic profiles, the study describes the bio-geomorphic zonation of shelf-terraces and links this to the marine processes responsible for terrace formation. Then, supported by data from other shelf areas, it constrains the timing of the sea-level changes responsible. Finally, by providing insights into terrace formation around Grand Cayman, it proposes a general model of terrace formation that relates different terrace types to different phases of the sea-level cycle.

SETTING

Climatic

Grand Cayman is a small, low-lying island located in the northern part of the Caribbean Sea (Figure 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 (Figure 1). Due to variations in fetch, 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 (Figure 1). Both types of disturbances regularly inflict serious damage on the island's marine and terrestrial ecosystems.

Bathymetric

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 (GRAMMAR *et al.,* 1989).

Tectonic

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).

Methods

Direct observations and measurements of underwater features were made on 20 SCUBA transects from shore to the edge of the island shelf (Figure 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 (Figure 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 resolu-

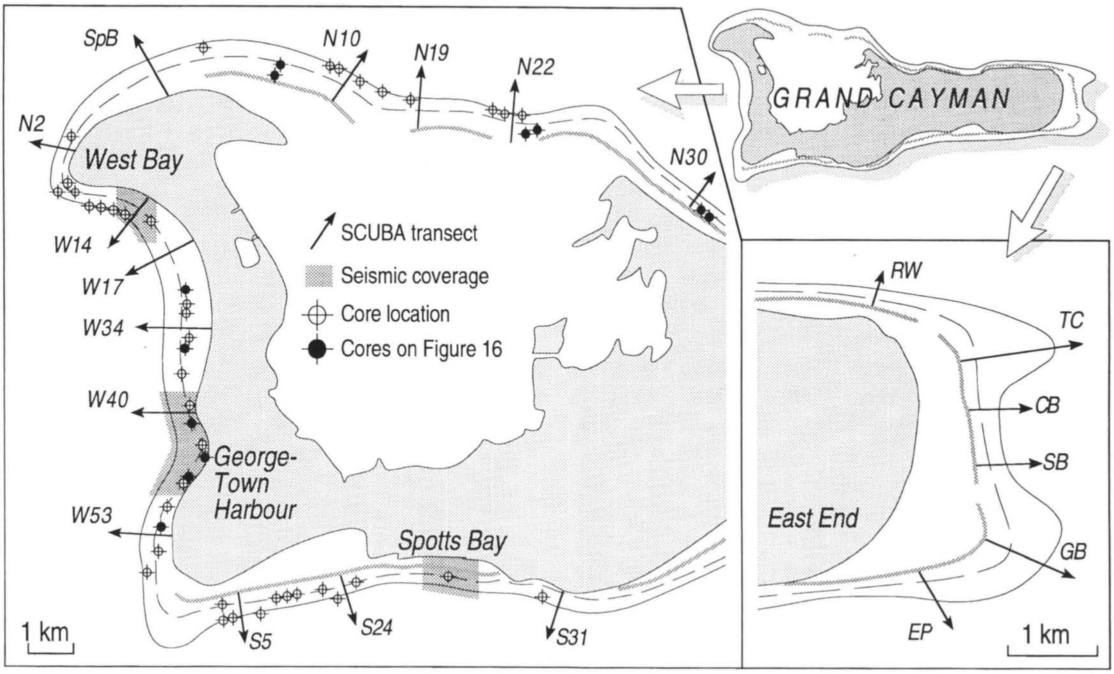

Figure 2. Location of SCUBA transects, areas of seismic coverage, and diamond-drill core locations on the shelf around Grand Cayman. (Dashed line shows position of mid-shelf scarp and shaded lines are fringing reefs).

tion 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 (Figure 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} the average of empirical values that range 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 (Figure 2). A further 40 cores, cut during a dive-site mooring installation program, were donated by the

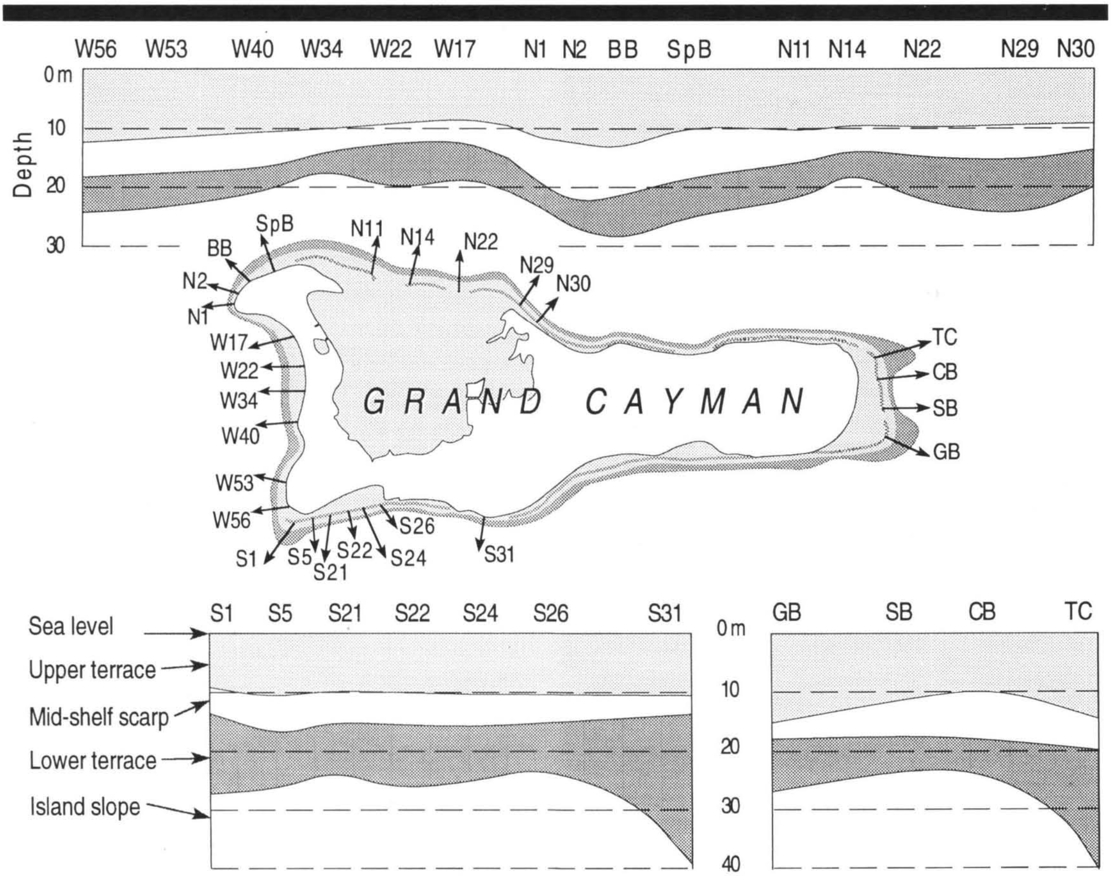

Figure 3. Locations of depth-sounding-profile grids (arrows) and positions of the upper terrace, mid-shelf scarp, and lower terrace at those locations. Position of slope breaks determined from profiles over non-reefal substrates. Note how the lower terrace is flatter on the protected-windward and leeward 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).

Natural Resources Unit of the Cayman Islands Government (Figure 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 standard X-Ray Diffraction. Radiocarbon determinations, made by conventional gas proportional counting and using the Libby half life (5568 yrs. BP), yielded an average analytical uncertainty (1 sd) of ±1150 years for samples older than 17 ka, and ±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).

SHELF MORPHOLOGY AND ZONATION

The shelf around Grand Cayman consists of two distinct terraces separated by a mid-shelf scarp (Figure 4). Although consistently present, their positions vary as a result of sediment deposition and reef growth (Figure 3). For the purpose of discussion, the terraces of the leeward and protected windward margins are considered together

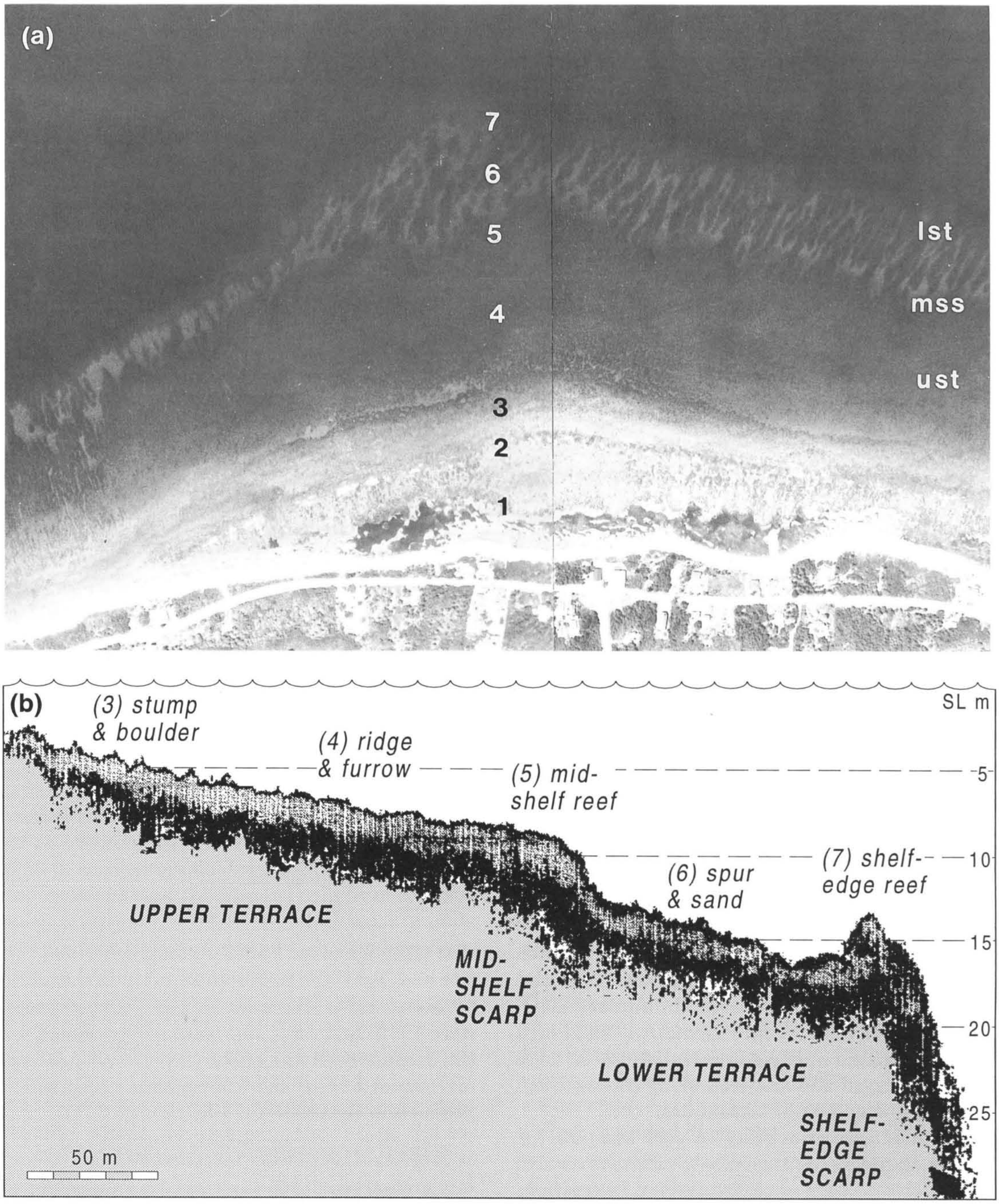

Figure 4. Aerial photograph (a) and depth-sound profile (b) showing terraces and bio-geomorphic zonation along the protected-windward shelf of Grand Cayman (North Side Village, 2 km east of transect N30). Upper-shelf terrace (ust); narrow lagoon (1), fringing-reef crest and flat (2), stump-and-boulder zone (3), ridge-and-furrow zone (4), mid-shelf-reef zone (5). Mid-shelf scarp (mss). Lower-shelf terrace (lst); sand-and-spur zone (6), shelf-edge reef (7). Note how the lower terrace shows a typical exposure trend in the sand-and-spur zone; wind-facing areas have a dense array of spurs whereas wind-sheltered areas have less spurs and are dominated by sand. Note also the concomitant change in spur orientation.

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 Figures 5a and 5b respectively).

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 (Figures 3 and 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 (Figure 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 (Figure 6a). Robust colonies and broken stumps of *in-situ A. palmata* are common but sparsely distributed (Figure 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, 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 (Figure 7a). Along shallow, wind-facing parts of the terrace, ridges and furrows are closely spaced (< 5 m), high-amplitude features (> 0.5 m), (Figure 7b). 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. Increased erosion here commonly accentuates these features, producing blind-ended furrows and pot-holes up to 2 m deep (Figure 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 (Figure 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 (Figure 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 patchy reefal development, and the mid-shelf scarp is generally better exposed (Figure 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 the protected upper terrace and range down to the 10 m isobath. The exposed upper terrace is divided into concentric bio-geomorphic zones which consists of 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 (Figure 5b).

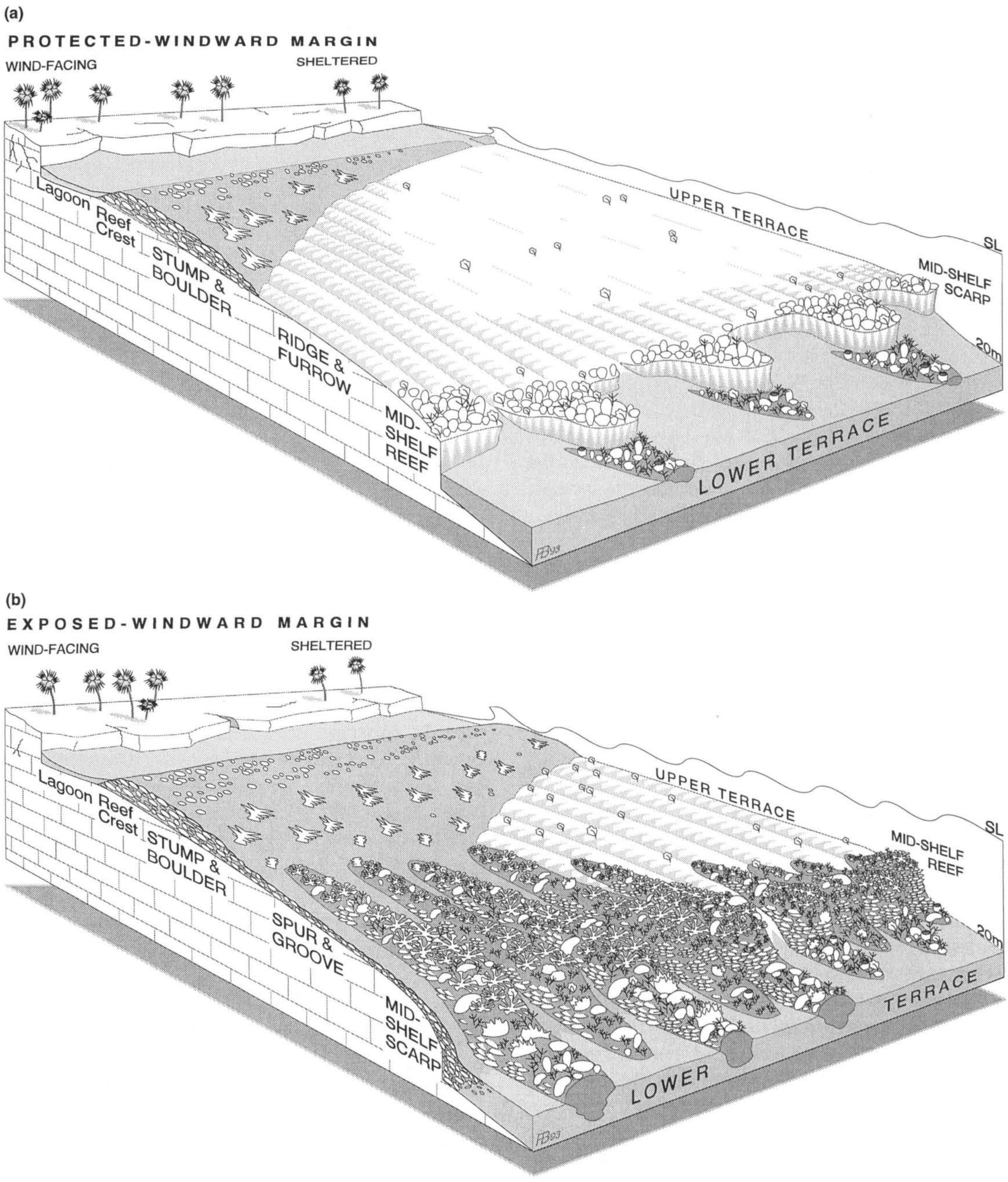

Figure 5. Schematic summary of exposure trends in upper-terrace zonation and substrate composition along (a) protected-windward margin and (b) exposed-windward margin. Note how enhanced surf action along wind-facing parts of the shelf produces wider stump-and-boulder and ridge-and-furrow zones than along 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.

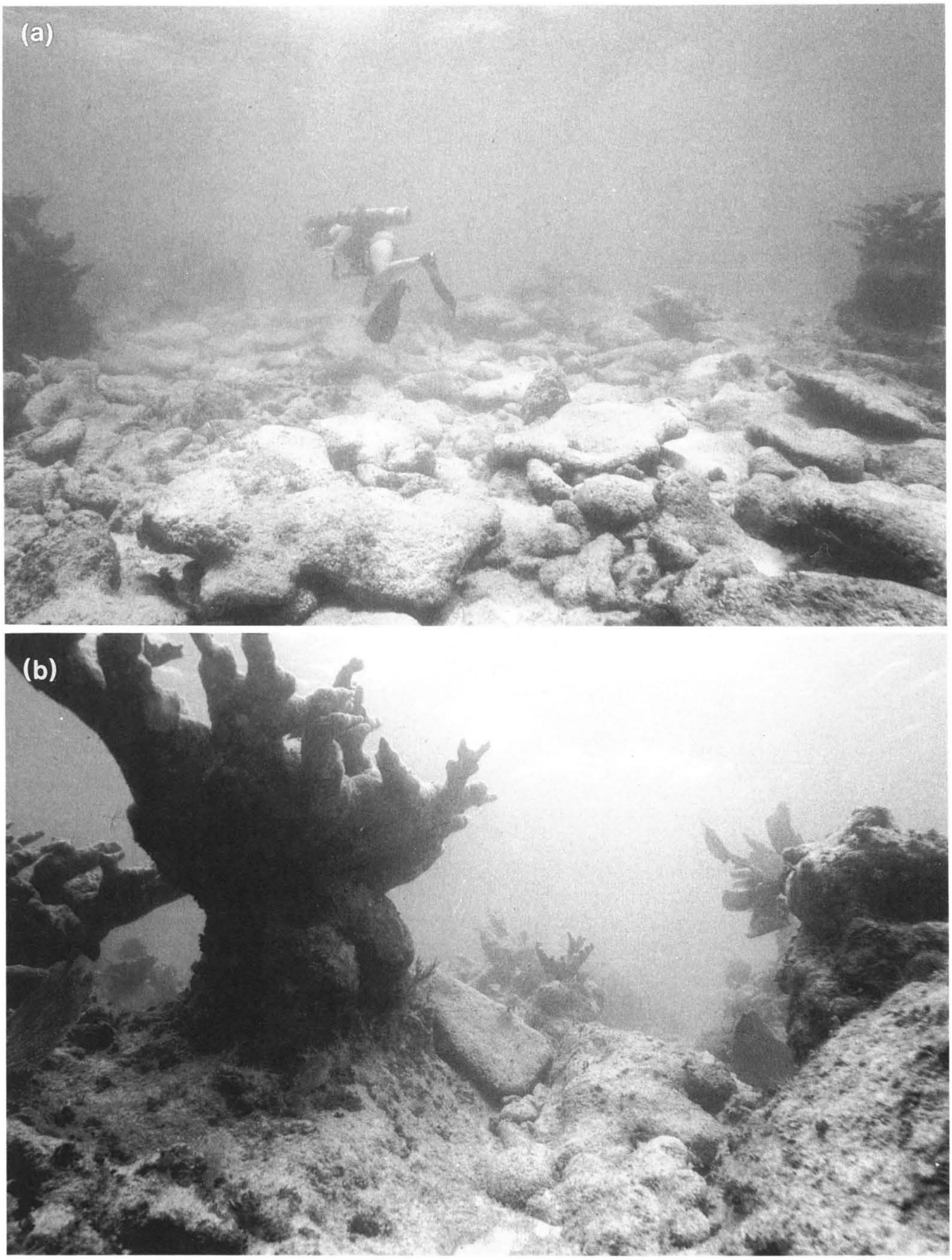

Figure 6. Photographs of the stump-and-boulder zone along the protected-windward margin (transect N30). (a) Well developed, wind-facing, boulder field in 3.5 m of water. Note, boulders are not stabilized. Diver's fins are 60 cm long. (b) Regenerated stumps of *Acropora palmata* (4 m water depth) showing evidence of cyclic growth and destruction (irregular knots in branches; compare with Figure 10(b). Boulder in foreground is 50 cm long.

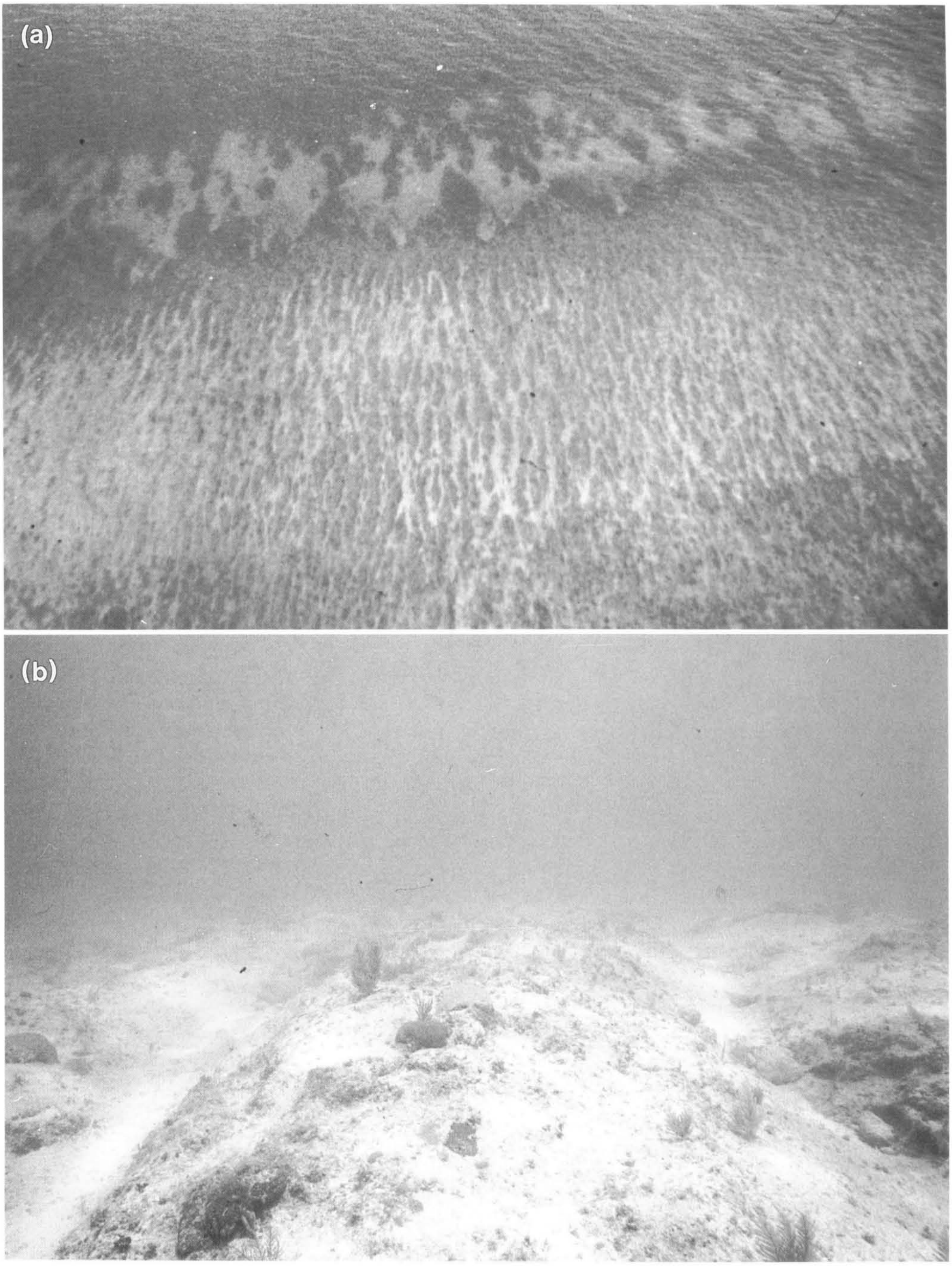

Figure 7. Photographs of the ridge-and-furrow zone along the protected-windward margin (transect N30). (a) Oblique aerial photograph of the ridge-and-furrow zone covering the upper terrace in a wind-facing area. Upper terrace is ~200 m wide. (b) Underwater view of the same area showing a ridge flanked 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 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 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.

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 (Figure 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 (Figure 10b). In more sheltered areas of the terrace, the zone is narrower and not as well developed; stumps and live colonies are 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 (Figures 5b and 11). 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 col-

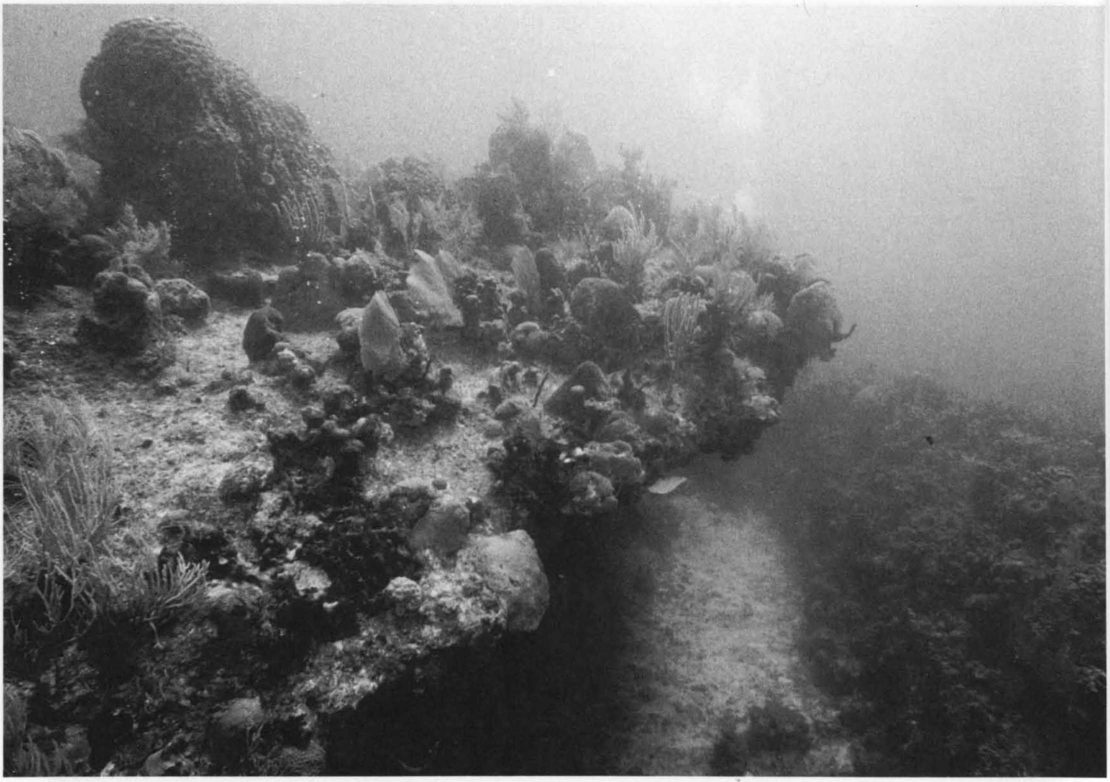

Figure 9. Mid-shelf-reef development along the leeward margin (W40) showing typical sparse coral-coverage concentrated near the shoulder of the mid-shelf scarp (large *M. annularis* coral is 1.5 m tall; water depth is 18 m at base of mid-shelf scarp and 12 m at the shoulder). Contrast with Figure 12.

onized 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 (Figure 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 (Figure 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 (Figure 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 (Figure 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).

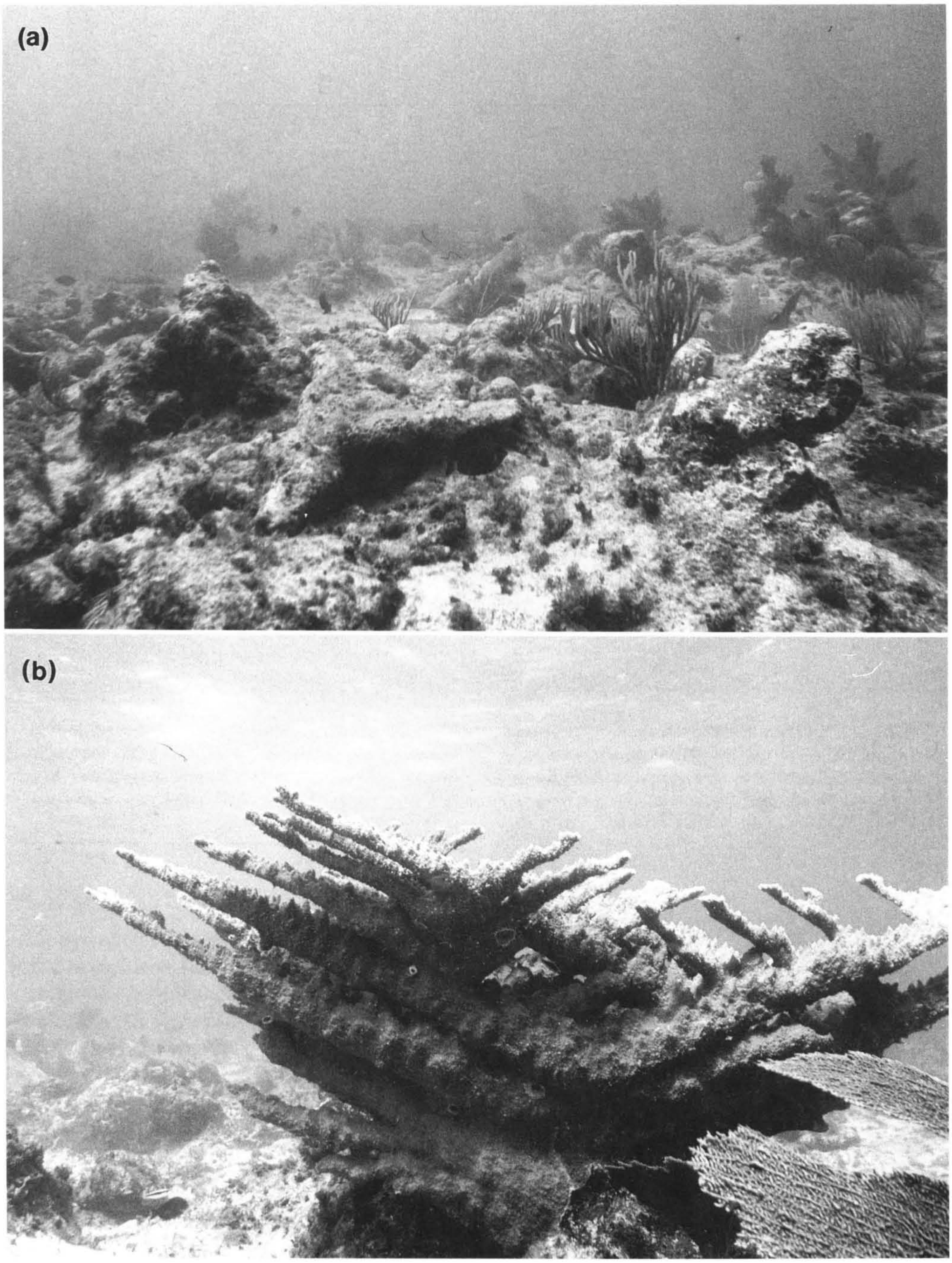

Figure 10. Photographs of the stump-and-boulder zone along the exposed-windward margin (transect S5 and S24). (a) Typical boulder field in 4 m of water along a wind-facing area. Boulders are stabilized by extensive crusts of coralline red algae (those in foreground are 1 m in length). (b) Live colony of *Acropora palmata* showing morphology adapted to strong, fairweather surf-action. Colony is 1 m tall and water depth is 3 m.

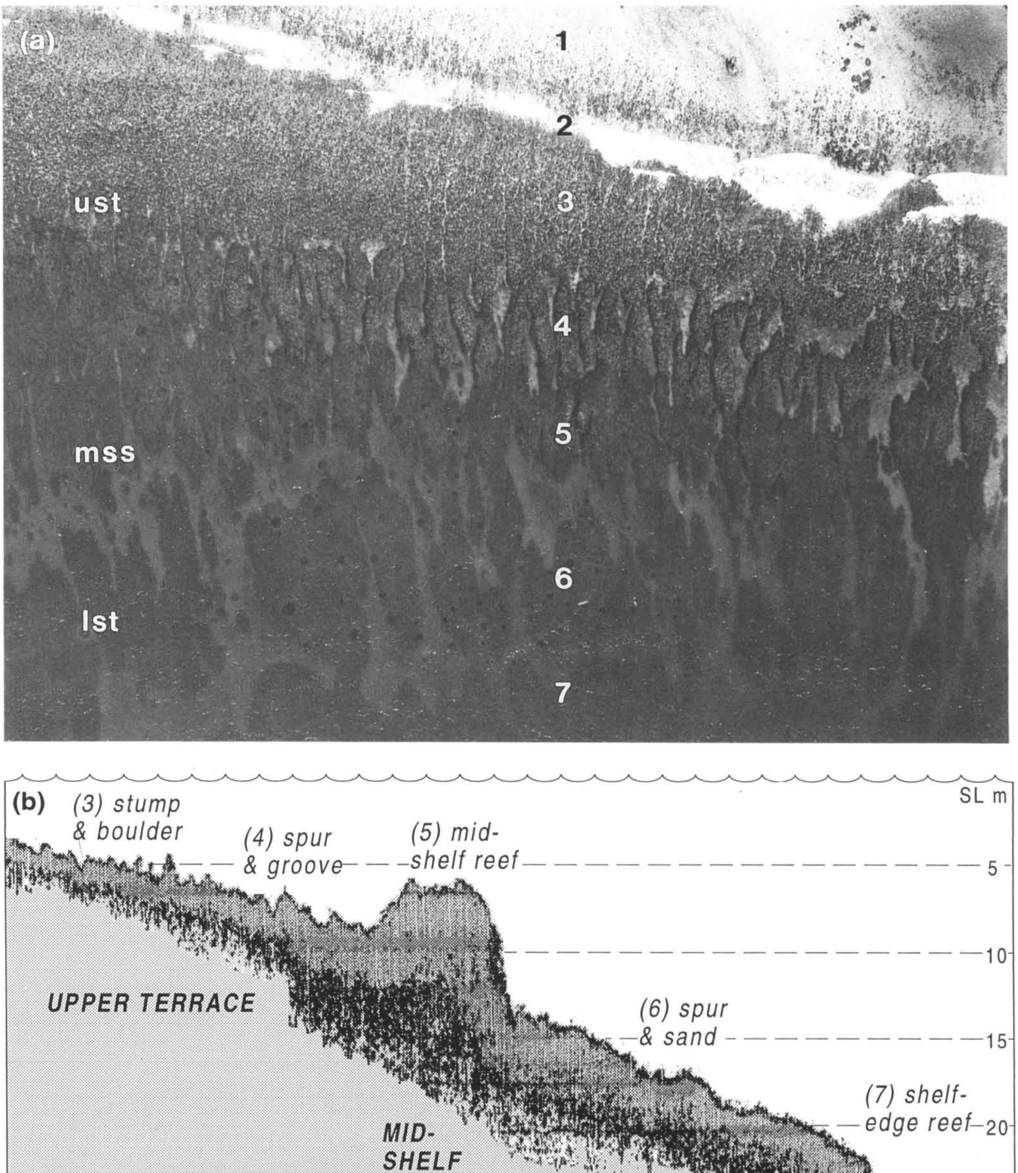

Figure 11. Aerial photograph (a) and depth-sound profile (b) showing terraces and bio-geomorphic zonation along the exposed-windward shelf of Grand Cayman (transect S5). Upper-shelf terrace (ust); lagoon (1), fringing-reef crest and flat (2), stump-and-boulder zone (3), spur-and-groove zone (4), mid-shelf-reef zone (5). Mid-shelf scarp (mss) mostly buried. Lower-shelf terrace (lst); sand-and-spur zone (6), shelf-edge reef (7).

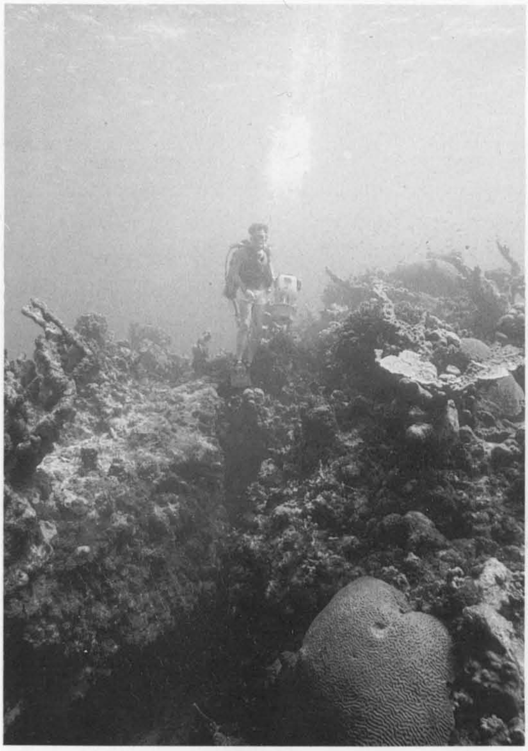

Figure 12. Mid-shelf-reef development along the exposed-windward margin (S24) showing typical dense coral-coverage obscuring mid-shelf scarp. Note roofed-over groove below diver (who is 2 m long in fins; water depth 8 m). Contrast with Figure 9.

Figure 13. Photograph of a drowned intertidal notch at −18.5 m cut into the mid-shelf-scarp (notch is 5 m high; transect N2). Lateral persistance and a drill-core from the notch demonstrates it is an erosional feature cut into the Pleistocene Ironshore Formation. The preservation of this feature in such pristine condition suggests it must have been drowned by a rapid sea-level-rise event (see Figure 19). So far, notches have been identified at the same depth around five other islands (see Figure 20).

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 (Figure 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 (Figure 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 (Figures 5 and 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 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 (Figure 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 (Figure 4

Figure 14. Lower-terrace sand-and-spur zone along a wind-facing section of the exposed-windward margin at a depth of ~20 m looking seaward (transect TC). Shows typical high amplitude (here 3.5 m high) and closely spaced spurs (diver in a 5 m-wide sand channel).

and 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. As a result, there is a significant degree of variation in the position of the terrace surface (Figure 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 (Figure 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 (Figure 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 (Figure 14). Individual spurs have relatively high amplitudes (averaging 3 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 (Figure 4). Spur amplitudes and gradients, however, are much low-

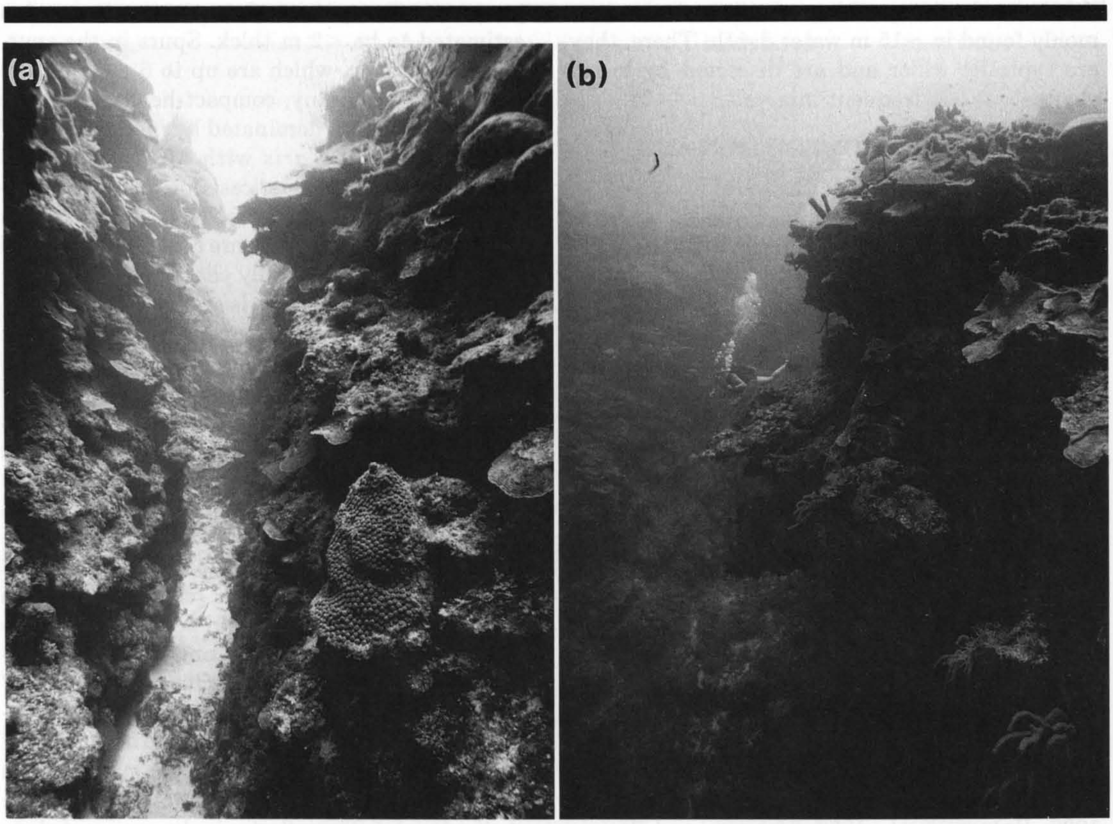

Figure 15. Lower-terrace shelf-edge-reef zone. (a) Steep canyons through the shelf-edge reef along the exposed-windward margin in 36 m of water (transect GB; *M. cavernosa* in foreground is 50 cm tall). (b) Vertical to overhanging shelf-edge-reef front along the protected-windward margin in 35 m of water (transect N30). Diver in background is 2 m in fins.

er than those on the exposed-windward shelf. 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 (Figure 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 *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 (Figure 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 (Figure 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 Figure 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 (Figure 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 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 caused a rejuvenation of the true coral ages, which 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 *Millipora* 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 <300 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 (Figure 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 (Figure 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.

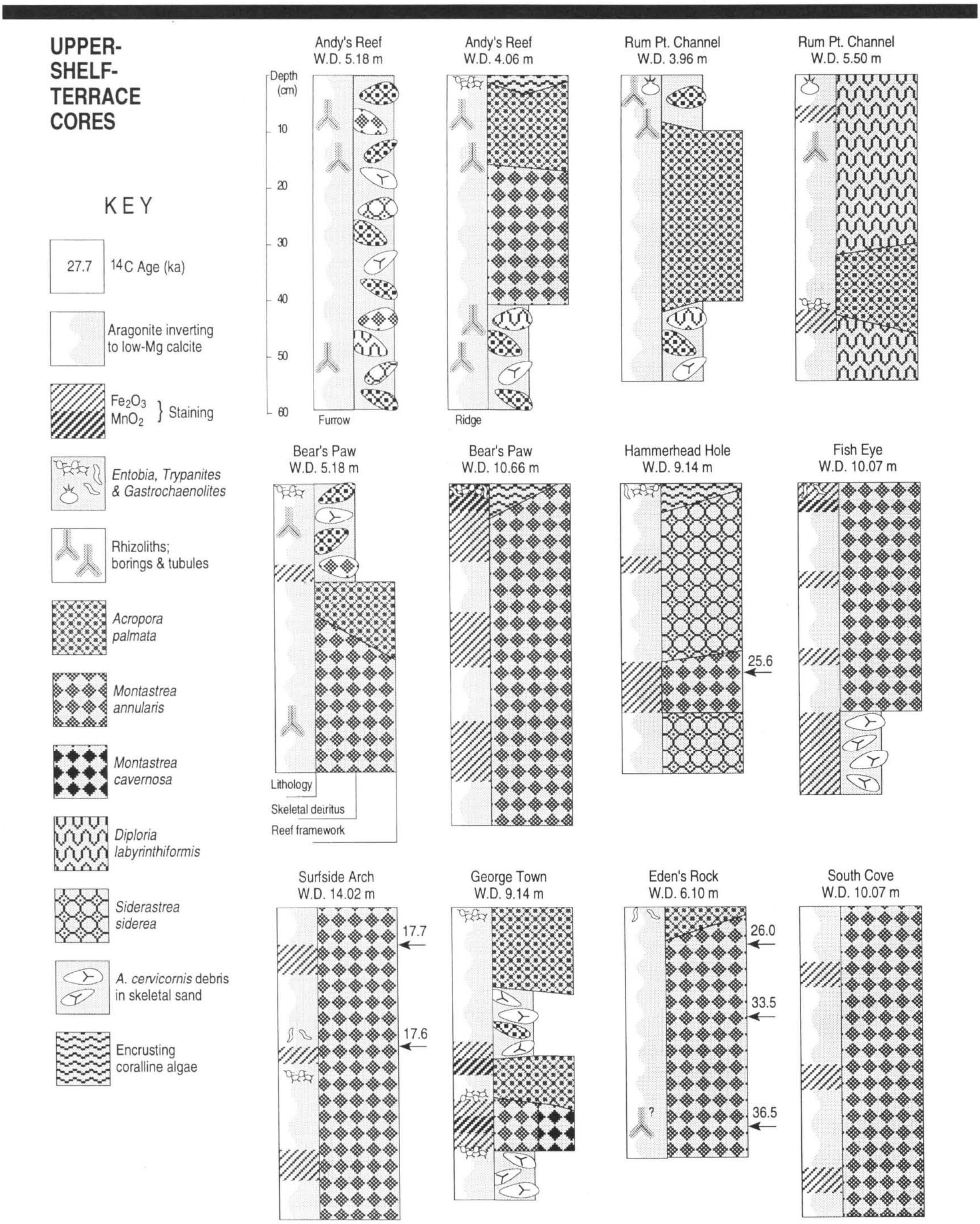

Figure 16. Selected cores from the upper-terrace ridge-and-furrow zone along the north and west sides of Grand Cayman. All cores show evidence of exposure to meteoric conditions including ferric and manganese oxide staining which coats coral framework, skeletal detritus, and meteoric cements. These oxides 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.

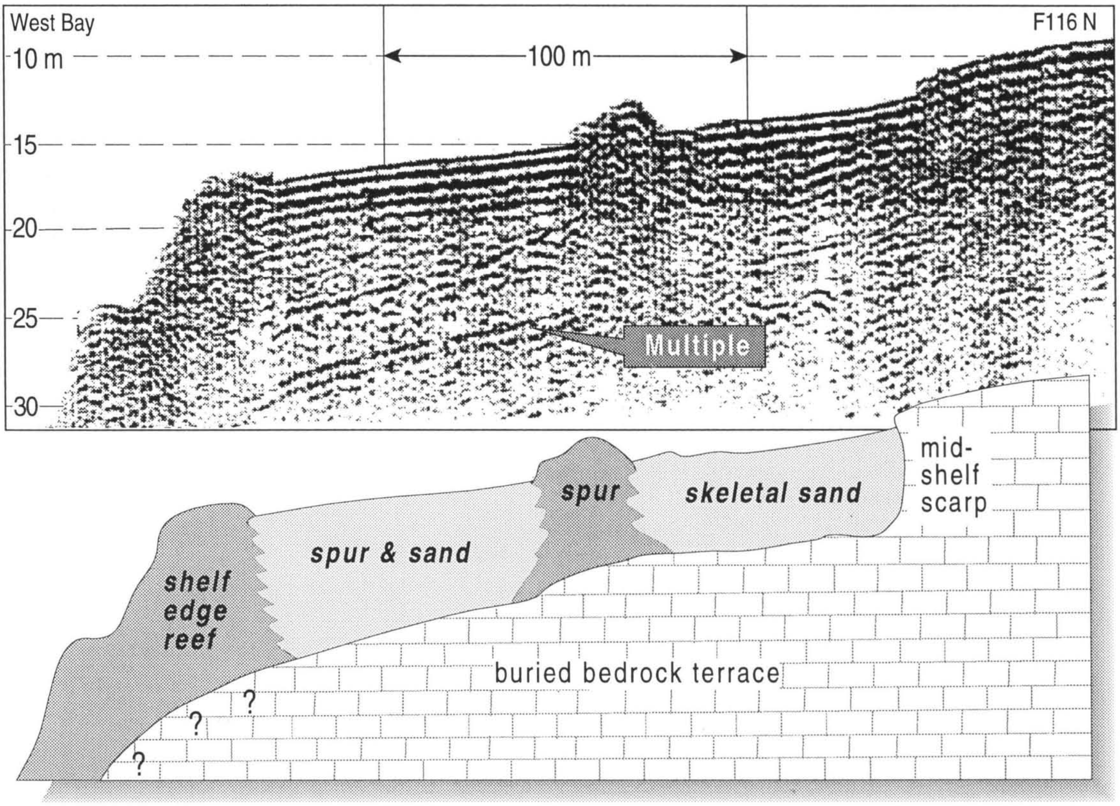

Figure 17. Seismic profile across the lower-shelf terrace on the west side of Grand Cayman (near transect W14). Interpretation below shows the typical architecture of modern deposits on the lower terrace, and the geometry of the bedrock terrace and mid-shelf scarp underlying them.

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); secular variations in storm recurrence intervals are common. Eyewitness accounts of a particularly severe hurricane that affected the Cayman Islands in 1932, provide 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 (Figure 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 (Figures 7 and 8). Third, sand and pebble-sized sediment, in suspension, erodes the terrace surface by sandblasting. The efficacy of these storm-abrasion

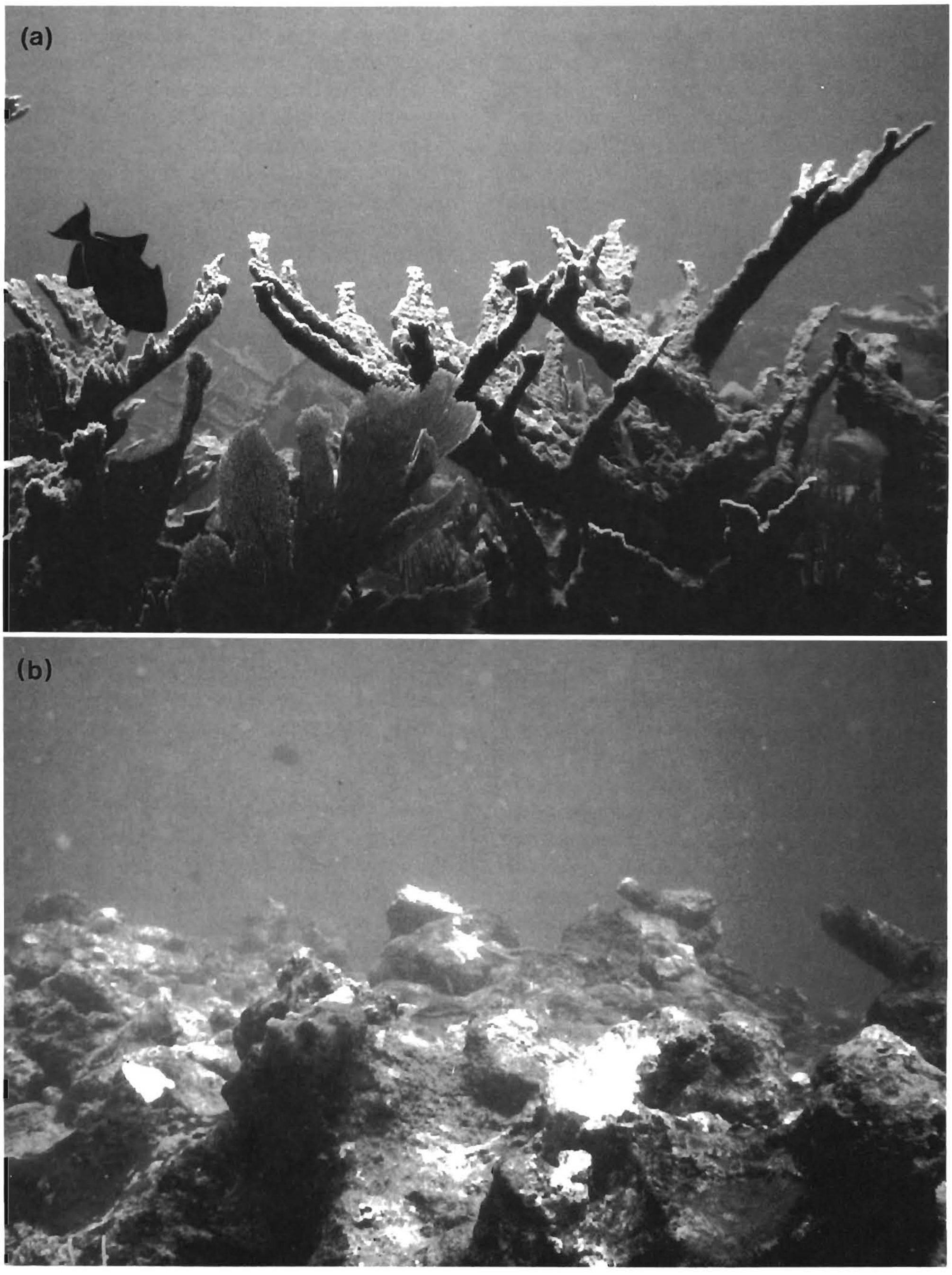

Figure 18. Before (a) and after (b) photographs illustrating the destructive effects of Hurricane Gilbert (1988) on spur-top coral communities (~5 m water depth) along the exposed-windward margin (transect S5). Triggerfish is ~25 cm long. Courtesy of Phillipe Bush.

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 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 (Figure 5a).

Erosion during storms has clearly played an important role in the formation of the upper-shelf terrace 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 (Figure 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 (Figure 13). Because cores from the notch are composed of soft limestone prone to relatively high rates of intertidal erosion, 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 (Figure 19).

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 (Figure 3). More importantly, however, the shelf terraces around Grand Cayman correlate with submerged terraces from islands in the Atlantic, Pacific, and Indian Oceans (Figure 20). This suggests that terraces on these islands were also produced by the same eustatic sea-level changes that affected Grand Cayman.

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. 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

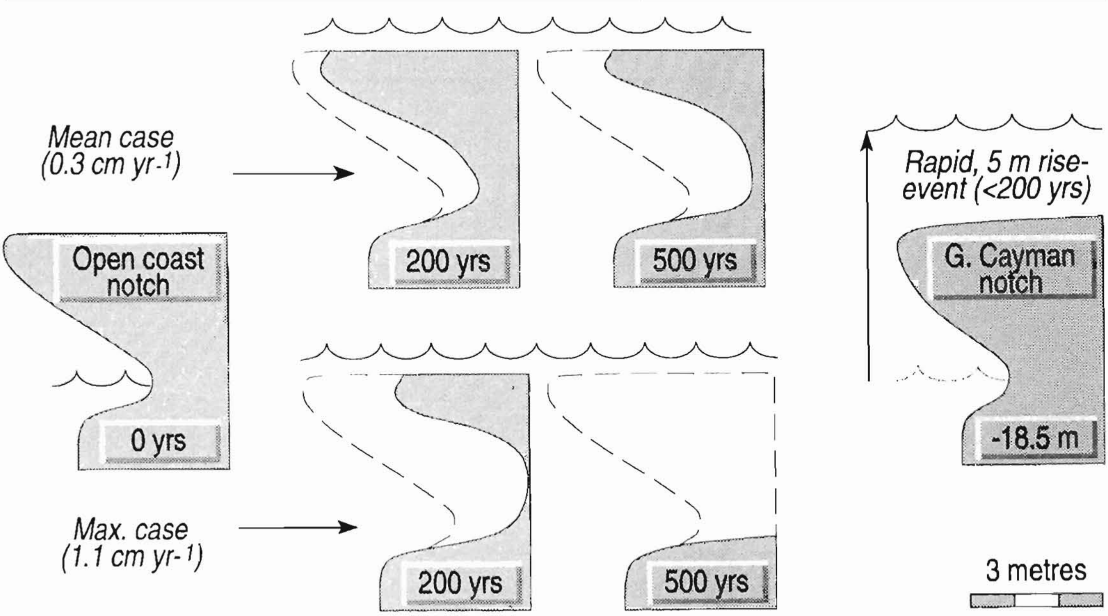

Figure 19. Simulated erosion of an intertidal notch at different rates of sea-level rise, using conservative erosion rates measured from Grand Cayman (SPENCER 1985). The morphology of the starting-point notch (0 yrs) is based on notches from open, microtidal coastal areas with regular wave action (cf. 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 suggests 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.

(ACKER, 1986; ACKER and STEARN, 1990), clearly identify terraces in positions coincident with stable islands such as Grand Cayman (Figure 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). These reefs all started to grow ~11 ka ago at a depth of 30–40 m below present sea level (Figure 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 (Figure 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 (Figure 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 (Figure 21; see Panama, Florida and Barbados).

The dates of *Acropora* reef growth consequently demonstrate that the lower terrace and the mid-

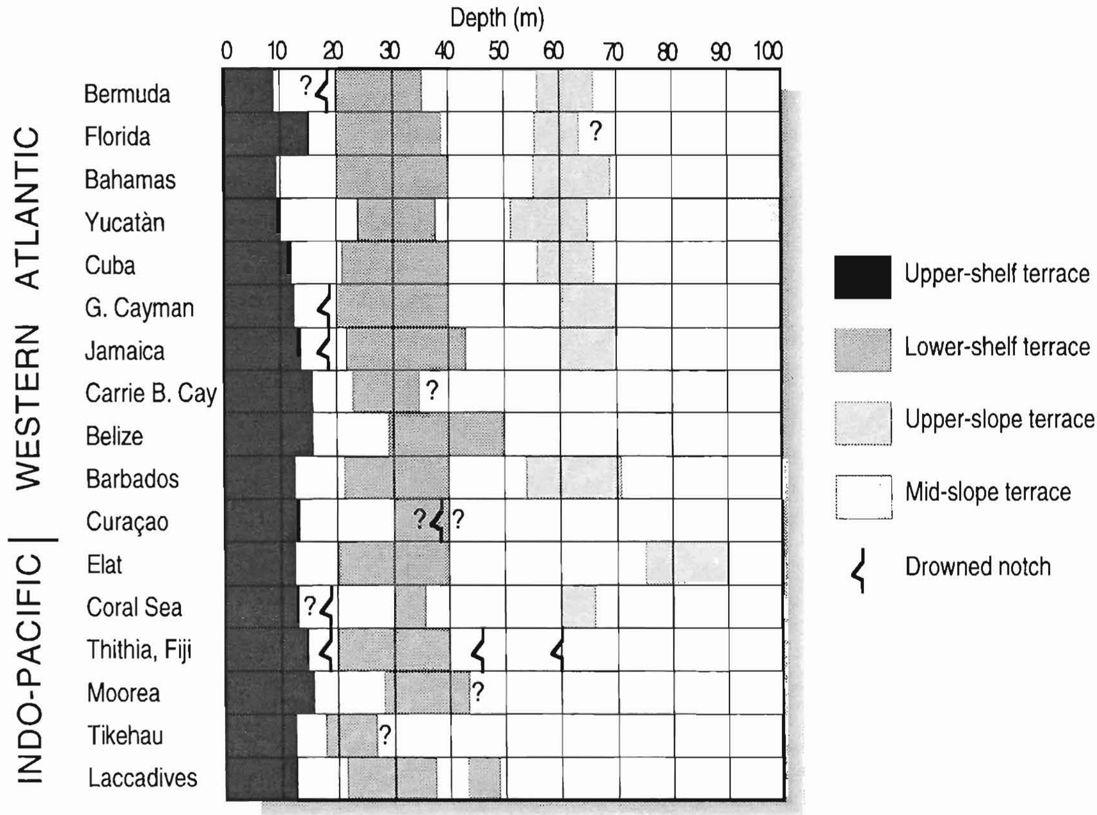

Figure 20. Depth range of terrace surfaces and positions of intertidal notches from islands in the western Atlantic, western and south Pacific, Red Sea and Indian Ocean. Variations in the depth range of terrace surfaces is due to accumulation of modern sediments and reefs. Bermuda—MEISCHNER & MEISCHNER (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).

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-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 (Figure 5). This suggests that such features are produced by, and are in equilibrium with, contemporary erosional

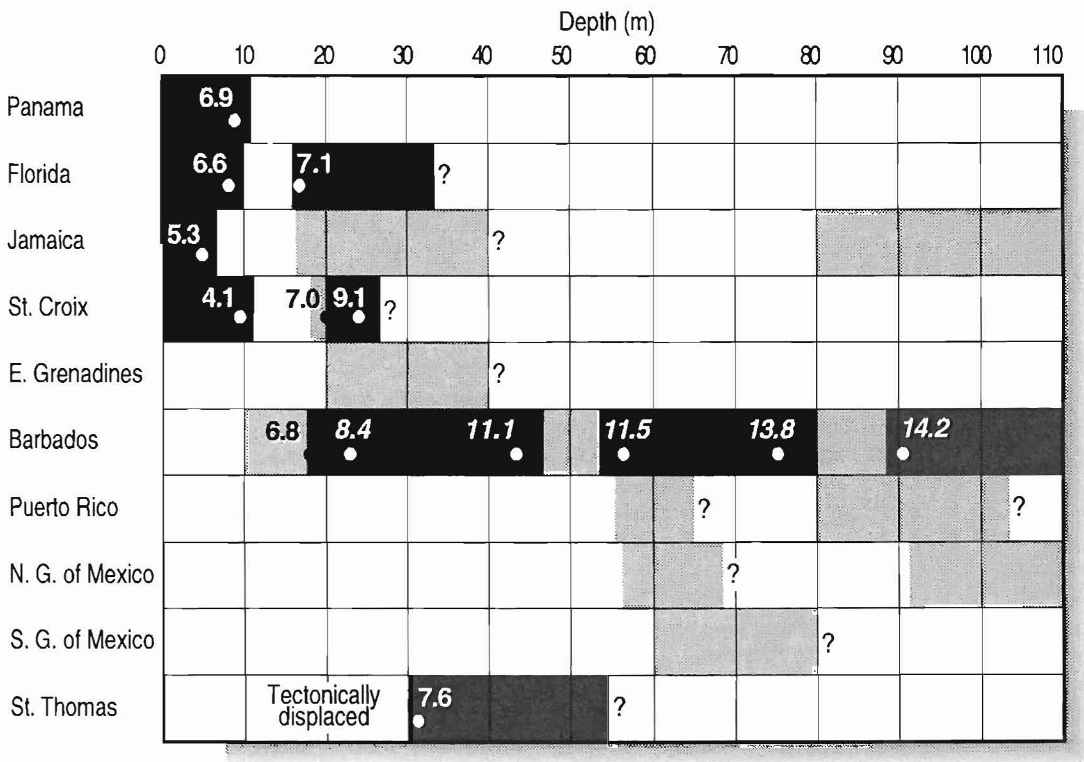

Figure 21. Reported positions and dates (ka) of drowned reefs in the Caribbean-Atlantic reef province. *Acropora palmata* framework shown by dark shade with dates in white; other coral and unknown framework shown by light shade with dates in black. Dates are radiocarbon except those in italics which 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).

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) 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 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 coast 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 could not address seasonal variations in erosion rates (*e.g.*, seasonal storms), let alone longer-term variations due to climate change. By calculating the average widths of the upper and lower terraces around Grand Cayman, it is possible to determine the actual rate of marine erosion over the last 11 ka, when sea-level reached

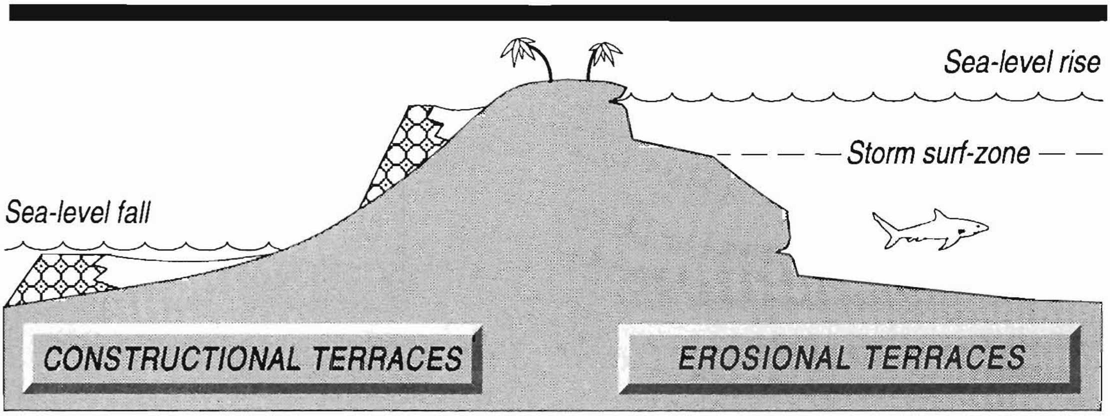

Figure 22. Erosional versus constructional terraces. During slowly rising sea level marine abrasion during storms sculptures seaward-sloping terraces, sea cliffs, and notches into the island bedrock. During sea-level fall, terraces are formed by the growth of fringing reefs and the accumulation of lagoonal sediments. Note, erosional terraces slope gently seawards, whereas constructional terraces are horizontal or slope gently landward.

the position of the lower terrace (BARD *et al.*, 1990). Along rocky coasts, 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. These rates 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 (Figure 22). From the depth-ranges shown in Figure 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 (Figure 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 variations in terrace positions reported from the Great Barrier Reef (*cf.* HOPLEY, 1982) and the western Atlantic (Figure 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 positions in other areas 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, suggest that the lower-shelf terrace (20–40 m bsl) 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 suggests 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. Another episode of slow sea-level rise ensued and produced the upper-shelf terrace (0–10 m bsl), 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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□ ZUSAMMENFASSUNG □

Der Schelf um Grand Cayman besteht aus zwei zum offenen Meer abfallenden Terrassen, welche durch eine Böschung voneinander getrennt sind. Die untere Terrasse (0–10 m Tiefe) besteht, außer entlang des Luvhanges, an dem Korallenwachstum dominiert, aus von Erosionsfurchen durchzogenem, nackten, steinigen Untergrund. Auftauchbedingte Trends in der Morphologie und die Verteilung dieser Erosionsmerkmale, sowie das Fehlen von Korallenwachstum, zeigen, daß die Terrasse das Ergebnis junger, jahreszeitlich bedingter Sturmerosion ist. Die obere Terrasse wird von einer Böschung (10–20 m Tiefe) begrenzt, welche, in den meisten Gebieten, teilweise oder vollständig von jungen Karbonatsedimenten bedeckt ist. Jedoch liegt die Böschung entlang schmaler Abschnitte des leeseitigen Schelfes im allgemeinen frei und zeigt dort, in einer Tiefe von 18,5 m, einen erosiven, intertidalen Einschnitt. Die untere Terrasse (12–40 m Tiefe) erstreckt sich von der Böschung bis zum Schelfrand. Die Oberfläche dieser Terrasse wird von einem jungen Riff- und Sedimentkeil gebildet, der zum Schelfrand hin mächtiger wird und bis zu 40 m Mächtigkeit erreicht. Diese Ablagerungen werden von der zum Meer abfallenden Terrasse (20–40 m Tiefe) unterlagert. Die überdeckte Terrasse und die Böschung repräsentieren eine frühere Phase mariner Einebnung als der Meeresspiegel in einer tieferen Position stabilisiert war.

Die heutigen Erosionsmerkmale der oberen Schelfterrasse und das Vorkommen identischer Terrassen auf rezent herausgehobenen Inseln zeigen, daß die Terrassen auf Grand Cayman durch marine Erosion während des letzten, nacheiszeitlichen Meeresspiegelanstieges entstanden. Die untere Terrasse und die Böschung wurden während eines langsamen Meeresspiegelanstieges vor 11.000 bis 7.000 Jahren erodiert und vor ca. 7.000 Jahren von einem extrem schnellen, 5 m Meerespiegelanstieg ertränkt. Auf dieses katastrophale Ereignis, das auch schnell wachsende *Acropora*-Riffe in anderen Gebieten der Karibik ertränkte, folgte eine Stabilisierung des Meeresspiegels und der langsame Anstieg in seine jetzige Position. Der betont stufenweise Anstieg des Meeresspiegels im Holozän muß noch von außerhalb der karibisch-atlantischen Riffprovinz bestätigt werden, stimmt aber mit vor-holozänen Meeresspiegelkurven überein.

Das Vorhandensein zum Meer abfallender Terrassen auf vielen Schelfgebieten zeigt, daß erosives Einschneiden von Terrassen ein häufiges Phänomen bei Meeresspiegelanstiegen ist. Im Gegensatz dazu werden Terrassen bei einem relativen Rückgang des Meeresspiegels durch Riffanlagerung produziert. Dies deutet auf einen genetischen Zusammenhang zwischen Meeresspiegelzyklus und Terrassentyp hin. Erosionsterrassen entstehen bei einem Meeresspiegelansteig, Anlagerungsterrassen bei einem Meeresspiegelrückgang.

□ RESUMEN □

La plataforma alrededor de Gran Caiman está constituida por dos terrazas que buzan hacia el mar separadas por un escarpe. A excepción del margen expuesto de barlovento en el cual predomina el crecimiento de corales, la terraza superior (0–10 m bajo el nivel del mar) esta constuída en gran parte por una superficie rocosa atravesada o cortada por surcos de caracter erosional. Evidencias de exposición en la morfología y la distribución de estos caracteres erosivos, ademas de la ausencia de crecimientos coralinos, demuestran que dicha terraza es el resultado de erosión contemporanea a epocas de tormenta. La terraza superior termina en un escarpe (10–20 m debajo del nivel del mar) que en su mayoría está parcial o totalmente cubierto por depositos carbonáticos modernos. Sin embargo, a lo largo de estrechas secciones en el lado de sotavento de la plataforma a 18.5 m de profundidad, es común que dicho escarpe este expuesto mostrando cortes de carácter erosivo producidos en el área intermareal.

La terraza inferior (12–40 m bajo el nivel del mar) se extiende desde el escarpe hasta el borde de la plataforma. Su superficie actual está cubierta por un arrecife moderno y su respectiva cuña de sedimentos que se engrosa hacia el borde de la plataforma, alcanzando mas de 40 m de espesor. Estos depositos suprayacen un sustrato rocoso que constituye una terraza que buza hacia el mar (20–40 m bajo el nivel del mar). Dicha terraza cubierta por sedimentos, así como el escarpe adyacente a ellas son equivalentes geomorfológicos a la terraza superior y a el acantilado de la costa, estos ultimos representando un episodio mas temprano de peniplanación marina cuando el nivel del mar era mas bajo. Los caracteres de erosión contemporanea en la terraza superior y la presencia de terrazas idénticas alrededor de islas recientemente levantadas, demuestran que las terrazas de Gran Caiman fueron esculpidas por erosión marina durante el último ascenso del nivel del mar en la última deglaciación. La terraza inferior y el escarpe que divide ambas terrazas fué erosionado durante un episodio de subida lenta del nivel del mar entre 11–7 ka, siendo posteriormente sumergida por un rapido ascenso del nivel del mar, en el orden de 5 m, aproximadamente a los 7 ka. Después de este evento catastrófico el cual ahogó el rapido crecimiento de arrecifes constituídos por *Acropora* en otras areas del Caribe, el nivel del mar se estabilizó y ascendió hasta alcanzar su actual posición, formando así la terraza superior. Este patrón de escalonamiento pronunciado durante el ascenso del nivel del mar en el Holoceno falta por ser confirmado en las afueras de la provincia arrecifal Caribe-Atlántico, pero es consistente con la naturaleza escalonada de las curvas del nivel del mar para el pre-Holoceno.

La presencia de terrazas con buzamiento hacia el mar en muchas plataformas alrededor del mundo, sugieren que el fenómeno de incisión por erosión en las terrazas es común durante ascensos del nivel del mar. En contraste, terrazas en areas sometidas a descensos del nivel del mar tienen un origen constructivo, siendo producidas enteramente por acreción arrecifal. Esto sugiere que hay una relación genética entre ciclos de ascenso y descenso del nivel del mar y el tipo de terraza producida, formandose terrazas erosionales durante eventos de ascenso del nivel del mar, y terrazas de caracter constructivo durante descensos del nivel del mar.

□ RÉSUMÉ □

La plate-forme autour de Grand Cayman se compose de deux terrasses inclinées vers la mer et séparées par un escarpement en mi-plate-forme. Sauf le long du bord de la côté exposée au vent où la croissance de corail est dominante, la terrasse supérieure (0–10 m au-dessous du niveau de la mer) se compose en grande partie d'un pavé rocheux et stérile traversé de sillons érosionals. Des tendances dans la morphologie liées à l'exposition et la distribution de ces traits érosionals, et la manque de croissance de corail, démontre que la terrace est le résultat d'érosion contemporaire durant des orages saisonniers. La terrasse supérieure est délimitée d'un escarpement en mi-plate-forme (10–20 m au-dessous du niveau de la mer) qui, dans la majorité des régions, est partiellement à complètement enterrée par des dépôts carbonates modernes. Pourtant au long des sections étroites de la plate-forme de la côté sous le vent l'escarpement est généralement exposé et déploit une encoche entremarée érosionale à −18.5 m. La terrasse inférieure (12–40 m au-dessous du niveau de la mer) s'étend de l'escarpement en mi-plate-forme jusqu'au bord de la plate-forme. Sa surface consiste d'un coin de récifs et de sédiment modernes qui s'épaissit vers le bord de la plate-forme, atteignant des épaisseurs de 40 m. Ces dépôts sont au-dessus d'une terrasse (de roche de fond) inclinée vers la mer (20–40 m au-dessous du niveau de la mer). Cette terrasse enterrée et l'escarpement en mi-plate-forme, qui sont des équivalents géomorphiques de la terrasse supérieure et de la falaise côtière, représentent une épisode plus tôt d'aplanition marine quand le niveau de la mer était stabilizé à une position plus basse.

Les traits érosionals contemporaires de la terrasse à plate-forme supérieure, et la présence de terrasses identiques autours d'îles récemment soulevés, démontrent que les terrasses sur Grand Cayman étaient sculptées par l'érosion marine pendant la dernière montée du niveau de la mer déglacial. La terrasse inférieure et l'escarpement de mi-plate-forme étaient érodés pendant une épisode de montée lente de 11–7 ka et étaient par la suite noyés par un évènement de montée très rapide (5 m) à ~7 ka. Suivant cet évènement catastrophique, qui a noyé des récifs d'Acropora, un type de avrail qui croit bès vite, dans d'autes régions des Antilles, le niveau de la mer a stabilizé et monté lentement à sa position actuelle, produisant la terrasse supérieure. Ce motif échelonné prononcé dans la montée du niveau de la mer holocène attend la confirmation d'au-dehors de la province des récifs des Antilles et de l'Atlantique mais est compatible avec le caractère échelonné des courbes du niveau de la mer pré-holocène.

La présence des terrasses inclinées vers la mer sur plusieurs plates-formes autour du monde suggère que le coupage des terrasses érosionales est un phénomène commun durant une montée du niveau de la mer. En contraste, les terrasses dans des régions qui ont subi une tombée relative dans le niveau de la mer sont d'origine de construction, être produit en entier par l'accroissement des récifs. Ceci suggère qu'il y a un rapport génétique entre le cycle du niveau de la mer et le genre de terrasse, avec les terrasses érosionales développent pendant la montée et les terrasses constructives pendant la tombée.