3 Modern carbonate environments

3.1 INTRODUCTION

Carbonate sediments are being deposited in many low-latitude areas at the present time, but three locations warrant an extended discussion since they possess a wide range of environments and lithofacies and provide good analogues to many ancient carbonate facies. These are the Bahama Platform, the South Florida Shelf and the Trucial Coast of the Arabian Gulf. These three carbonate platforms also provide modern examples of an isolated carbonate platform, rimmed shelf and ramp (see Section 2.4).

3.2 THE BAHAMA PLATFORM

The Bahama Platform is a classic area for carbonate sedimentation which has provided many modern analogues for ancient limestones. The size of the platform, 700 km north–south and 300 km east–west, is comparable with the areal distribution of many ancient carbonate sequences. The platform is close to sea-level and is divided into several banks by the deep channels, Exuma Sound, Tongue of the Ocean and Providence Channel (Figs 3.1 and 3.2). It is separated from the USA mainland by the deep Florida Straits and from Cuba by the Old Bahama Channel. These deep channels effectively cut the Bahama Platform off from any siliciclastic material, permitting very pure carbonate sediments to accumulate. The Bahamas is a large isolated carbonate platform (see Sections 2.4, 2.8).

The Bahama Platform has been an area of shallow-water carbonate sedimentation since the Jurassic, and around 5 km of limestone and dolomite, with some evaporite, rest on continental basement. The platform is immediately underlain by Pleistocene limestone and outcrops of this give rise to islands and cays, mostly located close to the platform margin. The distribution of the Pleistocene outcrops and the local topography on the platform are partly the result of karstic weathering during the pre-Holocene sea-level low when the platform was subaerially exposed (Purdy, 1974a).

Shallow waters cover much of the Bahama Platform and have an average depth of around 7 m; these give way very quickly to deeper waters at the platform margin where the seafloor descends rapidly to depths of several hundred metres in only a few kilometres, with local slopes reaching 40°.

The Bahama Platform lies within the Trade Wind Belt so that during the summer, March through August, persistent winds blow from the east and southeast, and during the winter, winds are more from the northeast (see Fig. 3.3). Waves generated by these winds are particularly strong along east-facing platform margins. The tidal range is low, around 0.8 m at the platform margin, decreasing on to the platform. Tidal currents are only significant in channels between islands, reefs and sand shoals, and where the platform margin configuration enhances the tidal effects, such as at the heads of the deep Tongue of the Ocean and Exuma Sound. Flood tidal currents are generally stronger than the ebb currents. Storms and hurricanes produce extreme currents and temporary sea-level rises. Winter storms are mostly from the northwest; hurricanes on the other hand are not so predictable, although many originate in the Atlantic. On the Great Bahama Bank, hurricanes may blow water off the platform from east to west, or pile up water against the Pleistocene-founded islands and reefs of the eastern margin. The great lateral extent of the platform, together with the marginal rocky shoals, considerably reduce the exchange of on-platform water with the surrounding open ocean. This protection has led to the development of an extensive region of relatively sheltered water, only affected by major storms. There are some cross-bank currents, mostly occurring where marginal shoals and islands are absent. The energy regime of the platform is such that the daily normal wind and current activity is capable of moving sand-sized particles only near the platform margin. This is where active bed forms occur. Sand and bed forms on the platform itself are only moved during major storms and hurricanes. Mud-grade carbonate can be moved routinely (Gebelein, 1974a).

Rainfall is heavy over the Bahama Platform at

Fig. 3.1 *Aerial photo mosaic of the Great Bahama Bank (GBB) and Little Bahama Bank (LBB). Features clearly visible include the banks and deep channels (Providence, Tongue of the Ocean and Exuma), the oolite shoals and reefs along the northern margin of LBB, Andros Island with tidal flats on the western side, and tidal oolite bars at the head of Tongue of the Ocean. Compare with Fig. 3.2. After Gebelein (1974a).*

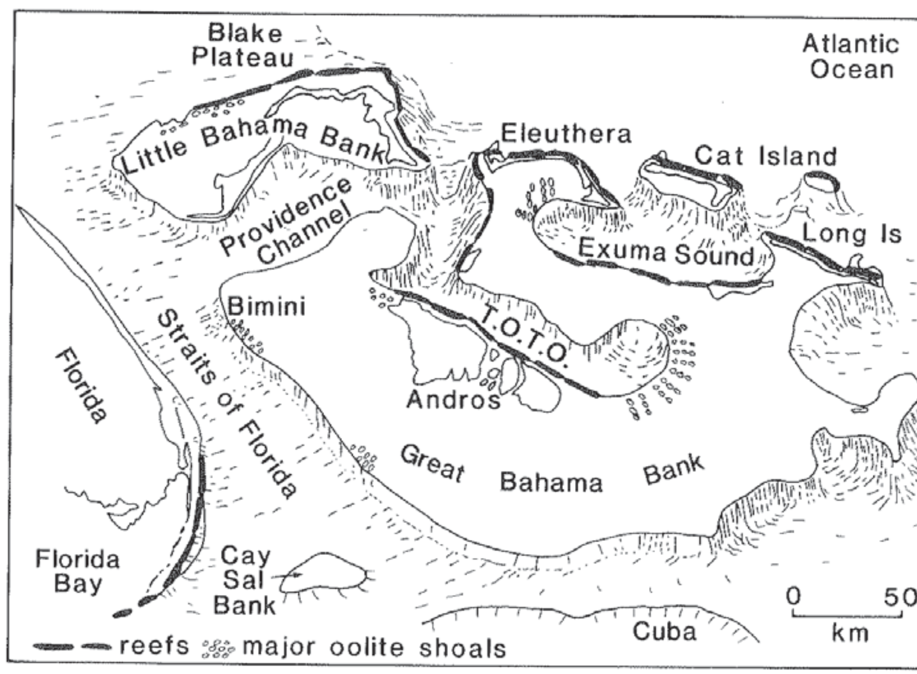


Fig. 3.2 *Banks, channels (Providence, Tongue of the Ocean, Exuma Sound and Straits of Florida), reefs and oolite shoals of the Bahamas and Florida. After Gebelein (1974a).*

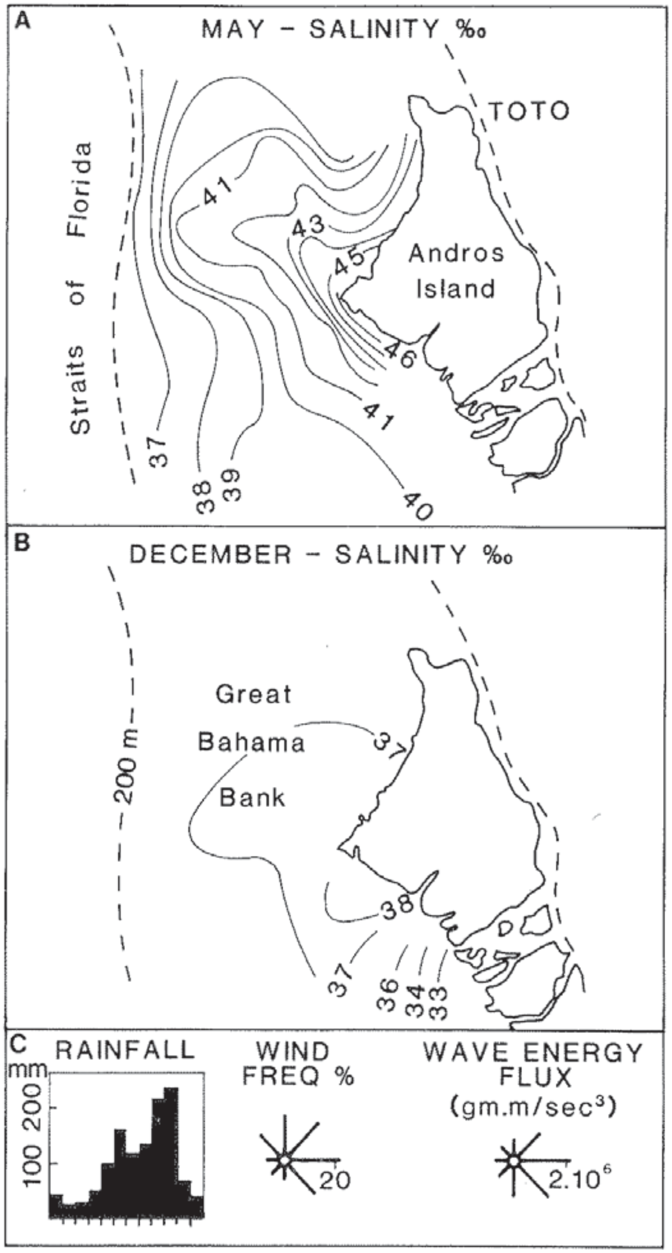


Fig. 3.3 *Salinity, rainfall, wind frequency and wave-energy flux for the Great Bahama Bank. After Gebelein (1974a).*

1200 mm yr^{-1}, but it is seasonal, most falling in the summer months, June through October (Fig. 3.3). Water temperatures generally vary from 22 to 31°C, but higher values are reached locally during the summer. Salinity has a normal oceanic value of 36‰ near the margins but it does reach 46‰ on the platform at times of intense evaporation as a result of the relatively poor circulation (Fig. 3.3).

The Bahama Platform is covered with only a thin veneer of modern carbonate sediments, generally less than 5 m thick, and these have accumulated in the last 4000 years.

3.2.1 Subtidal carbonate sediments of the Bahamas

Research on the sediments of the Bahama Platform has identified four major shallow-marine lithofacies and these reflect, to varying degrees, eight habitats and eight organism communities (Table 3.1; Newell *et al.*, 1959; Purdy, 1963a,b; reviews in Milliman, 1974 and Bathurst, 1975). Important attributes which distinguish the lithofacies from each other are grain composition, size and sorting, and lime mud content. Habitat refers to the environment where a particular group of organisms (a community) lives and of the many factors involved in determining the nature of a particular habitat (topography and depth, salinity, temperature, turbulence, turbidity, substrate, nutrient supply, etc.); the two main factors controlling the Bahamian habitats are topography and substrate. The interrelationships between organisms and their substrates, and other environmental parameters, are discussed in Newell *et al.* (1959). The Bahama–Florida region has also been discussed in terms of its biofacies (Coogan in Multer, 1977), defined as a group of organisms living together in a particular area which can be mapped. In effect, biofacies correspond to the organism communities of Newell *et al.* (1959). Biofacies distributions closely follow those of the lithofacies (Fig. 3.4), since much of the sediment is of biogenic origin; however, there are departures, since some sediment is strongly affected by physical processes and the biofacies respond to these and the several other factors which affect the habitat.

The four major shallow subtidal lithofacies are: (1) coralgal, (2) oolite, (3) oolitic and grapestone, and (4)

Table 3.1 Lithofacies, habitats and communities of the Great Bahama Bank. After Bathurst (1975)

Lithofacies	Habitat	Community
Coralgal	Reef	*Acropora palmata*
	Rock pavement	Plexaurid (sea whips)
	Rocky shore	Littorine
	Rocky ledges and prominences	*Millepora*
	Unstable sand	*Strombus samba*
Oolitic and grapestone	Stable sand	*Strombus costatus*
Oolite	Mobile oolite	*Tivela abaconis*
Mud and pellet mud	Mud and muddy sand	*Didemnum candidum*/ *Cerithidea costata*

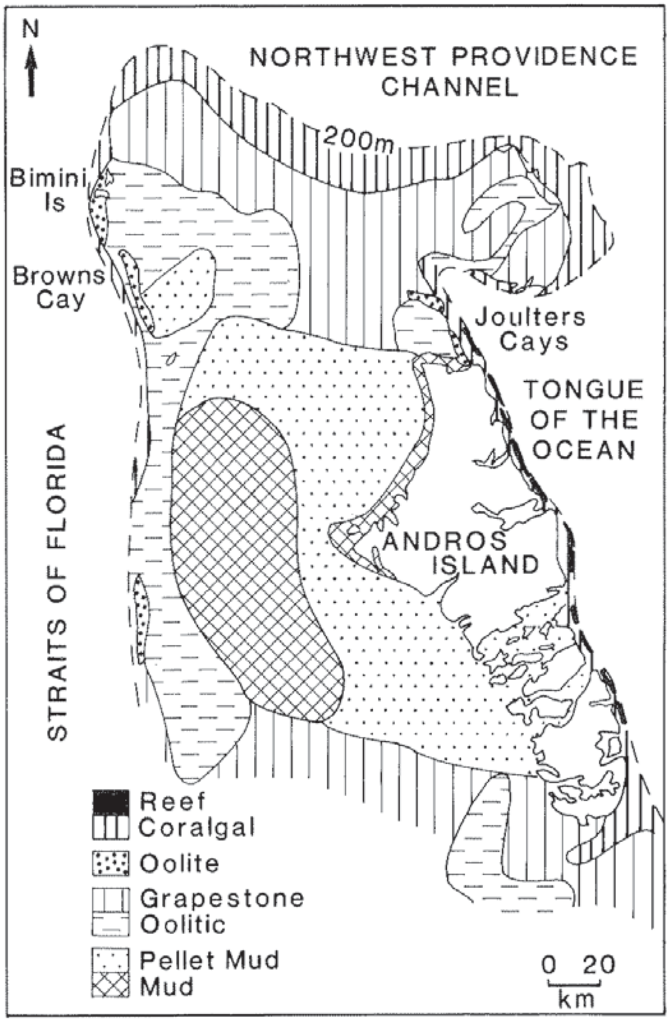


Fig. 3.4 *Lithofacies distribution on the Great Bahama Bank. After Newell* et al. *(1959) and Gebelein (1974a).*

lime mud and pellet mud lithofacies. The first two are principally platform-margin facies, whereas (3) and (4) are platform-interior facies. A fifth category is also described: (5) the deeper-water, periplatform ooze lithofacies.

3.2.1a Coralgal lithofacies

Coralgal lithofacies sediments consist of material from corals and calcareous red algae, along with skeletal debris from many other organisms. This facies occurs along the platform margins in depths ranging from intertidal (reef crest and beach) to around 50 m at the platform edge (Fig. 3.4). It occurs in areas of high wave energy, strong currents and much turbulence. Within this lithofacies group are included the important platform-margin reefs and lagoonal patch reefs (reef habitat, Table 3.1), back-reef (lagoonal) skeletal sands and muddy sands, and fore-reef platform-margin skeletal sands (both unstable sand habitat). Pleistocene limestone outcrops on the seafloor or along cliffed shorelines form the rocky-shore and rock pavement habitats where coralgal facies sediments also accumulate.

The *platform-margin reefs* are concentrated on windward margins (see Fig. 3.2) where wave action, turbulence and oxygenation are high, turbidity is low and where water temperature and salinity are similar to those of the open ocean. Such reefs are well developed along the eastern side of the Great Bahama Bank, forming a near continuous barrier opposite Andros Island with a back-reef lagoon 0.5–4 km wide between it and Andros (Newell & Rigby, 1957; Newell *et al.*, 1959; Gebelein, 1974a; Multer, 1977). They are also well developed along the northeast side of the Little Bahama Bank and off Eleuthera Island. Reefs are rare along the western side of the platform, since the dominantly easterly winds push warm hypersaline turbid water of the platform interior over the platform margin, and this is not conducive to prolific coral growth.

The reefs are constructed mainly of coral frame-builders with calcareous algae playing a binding and encrusting role. Five subenvironments are recognized in modern shelf-margin reefs (Fig. 2.8; Longman, 1981). From the ocean side to the lagoon these are: the *reef slope* in deep water (30–100 m) varying from steep to gently sloping with patches of bare reef rock (or Pleistocene limestone outcrops) colonized by various hard-substrate organisms (corals, sponges, gorgonians, etc.) and areas of reef talus; the *reef framework* and *reef crest* (may be just emergent at low spring tide), where most coral growth takes place; the *reef flat* just behind the reef and close to low tide level and consisting mostly of dead, *in situ* coral and other skeletons; and the *back-reef coralgal sand belt*, consisting of debris derived from the reef and washed over during storms, and sloping down to the lagoon floor at a depth of 2–6 m.

Some 30 species of coral are common and these frequently show a zoned distribution across the reef. The elkhorn coral (*Acropora palmata*), a strong branching coral, is a major frame-builder; luxuriant growth of this coral occurs in the reef crest area where wave action is intense (Fig. 3.5A,B). *A. palmata* is commonly oriented into or away from the wave surge on the reef crest, whereas on the reef flat it is unoriented, and much of it is dead and encrusted. Thinner-branched *A. palmata* occurs on the upper fore-reef slope. *Acropora cervicornis* (staghorn coral)

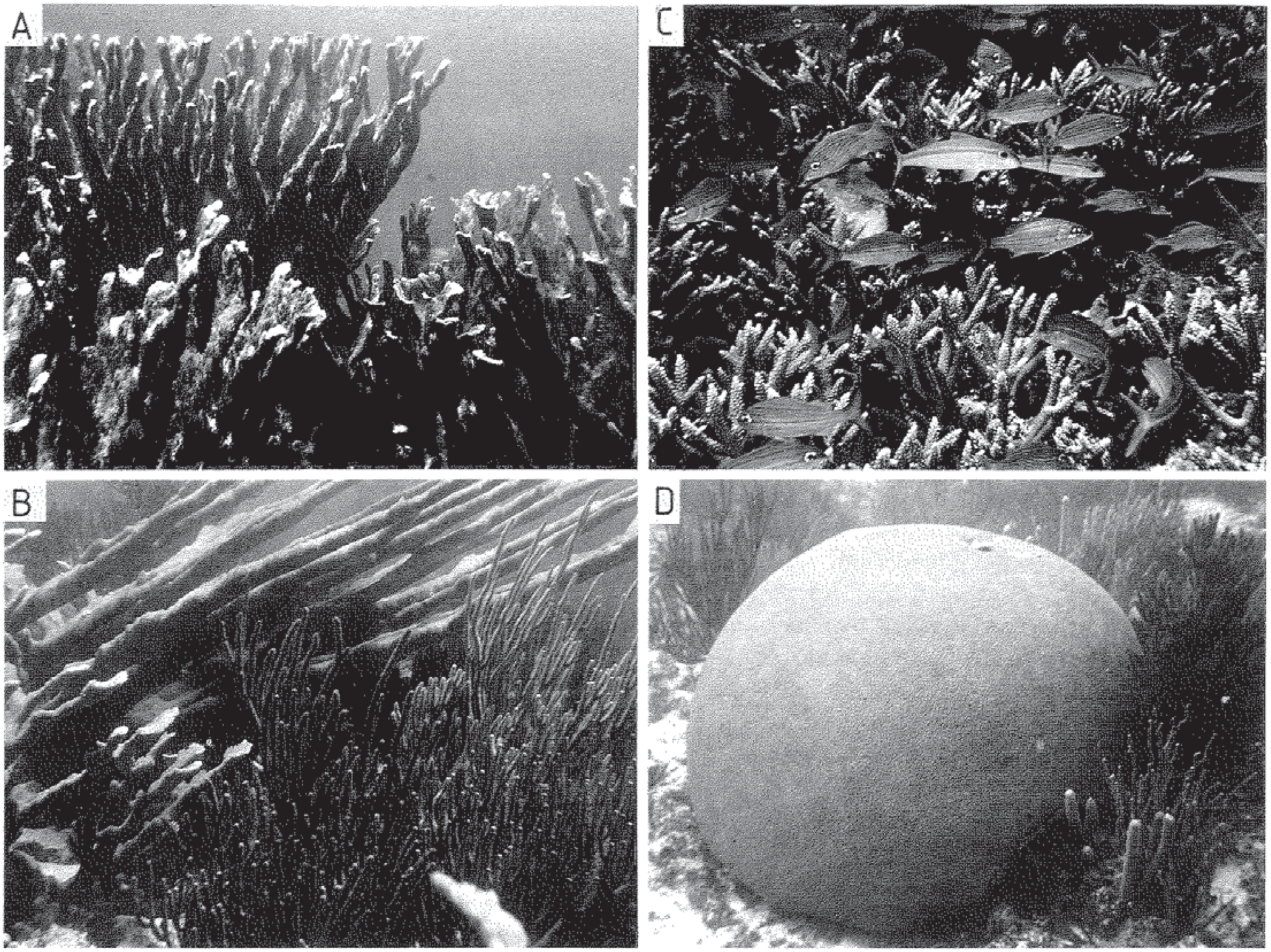


Fig. 3.5 *Bahamian–Caribbean reef corals. (A)* Acropora palmata, *the elkhorn coral. (B)* Acropora palmata *with branches oriented oceanwards, also soft coral (gorgonian) in foreground. (C)* Acropora cervicornis, *the staghorn coral. (D)* Diploria, *brain coral, with soft corals close by. From slides courtesy of R.P. Dunne and Barbara Brown.*

is a more delicately branching coral (Fig. 3.5C) that is common in less-exposed parts of the reef crest and on the upper fore-reef slope. Massive and domal corals, such as *Montastrea annularis*, *Diploria* (brain coral) (Fig. 3.5D) and *Siderastrea* are abundant on the upper fore-reef slope, with some growing to depths of 60 m. Small head corals, such as *Diploria*, *Siderastrea* and *Porites*, and delicately branching corals, *Porites*, *A. cervicornis and Agaricia* occur on the back-reef slope. Within and close by the reef live a variety of other organisms including molluscs, echinoids, foraminifera, gorgonians and sponges. The hydrocoral *Millepora*, is abundant on the reef crest. The bushy calcareous green (codiacean) alga *Halimeda opuntia* (see Fig. 3.7) is especially common on the reef and on death breaks up to give much lime sand.

Sheet-like calcareous red algae, such as *Lithothamnium* and *Goniolithon*, encrust corals and coral rubble to consolidate the reef framework. Some accessory organisms such as calcareous worms, bryozoans, foraminifera and bivalves also perform an encrusting role.

Many organisms are involved in bioerosion of the reef and this can also produce carbonate detritus. Lithophagid bivalves, clionid sponges, echinoids and endolithic algae bore into corals and other skeletons, making them more susceptible to damage and fracture by waves. Parrotfish eat coral polyps and in so doing rasp the coral skeleton to generate much lime sand and mud.

The platform-margin reefs are not continuous; they are traversed by major tidal channels every few kilometres which connect the back-reef lagoon with the open ocean. Also, along many of the reefs there are zones which are basically dead, and are undergoing erosion (biological and physical), contrasting with zones of active reef growth. In some areas a distinctive spur and groove morphology is developed, with lime sand derived from the reef occupying the grooves and the spurs forming buttresses 5–50 m across, of luxuriant coral growth, which descend down the reef slope into deep water. Studies of spurs and grooves of the Florida Reef Tract (see Section 3.3.1e) show that they are constructional in origin, not related to any antecedent topography in the underlying Pleistocene limestone (Shinn *et al.*, 1981).

Although forming a massive limestone, the reef has an open structure and there is much internal circulation of water. Fine sediment is carried into reef crevices and the pumping action of seawater promotes much precipitation of cement. Internal sediment is lithified, especially by micritic and peloidal high-Mg calcite, and acicular aragonite and bladed high-Mg calcite are precipitated as isopachous linings on cavity walls (see Section 7.4.1a).

Patch reefs occur in lagoons behind barrier reefs (e.g. Fig. 3.6) and are well seen in the windward lagoon adjacent to Andros Island (Gebelein, 1974a). They are several metres to tens of metres across, with 3–4 m of relief above the seafloor. They initiate on a hard substrate, which is mostly a Pleistocene limestone outcrop. Most luxuriant coral growth takes place around the margins of the patch reefs, where currents and turbulence are at a maximum. Sand accumulates in the depression on the reef top generated by marginal growth, and a halo of reef rubble and skeletal sand occurs around the patch reef. A similar organism community to the barrier reefs exists on the patch reefs, but *A. palmata* only occurs on the seaward side of the patch reefs where wave action is most intense. *A. cervicornis*, *Montastrea*, *Diploria*, *Siderastrea* and *Porites* dominate. Low mounds composed largely of calcareous red algae, with few corals, have formed in some back-reef areas.

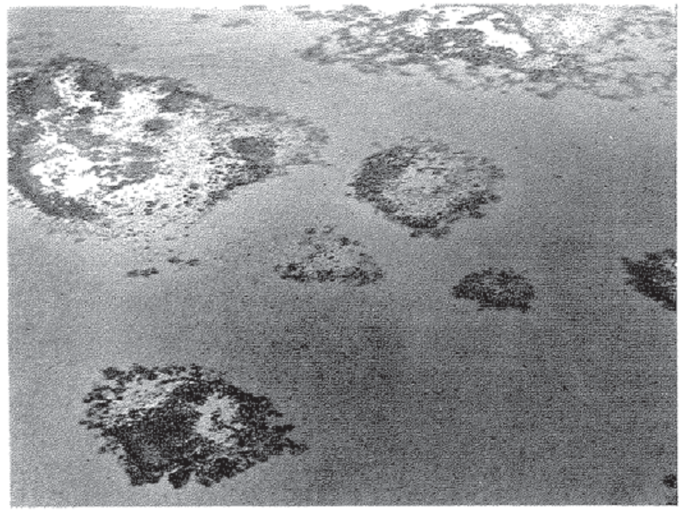


Fig. 3.6 *Patch reefs in lagoon behind shelf-margin reef. Rippled sand occurs between the patch reefs which show some zonation of corals. From slide courtesy of R.P. Dunne and Barbara Brown.*

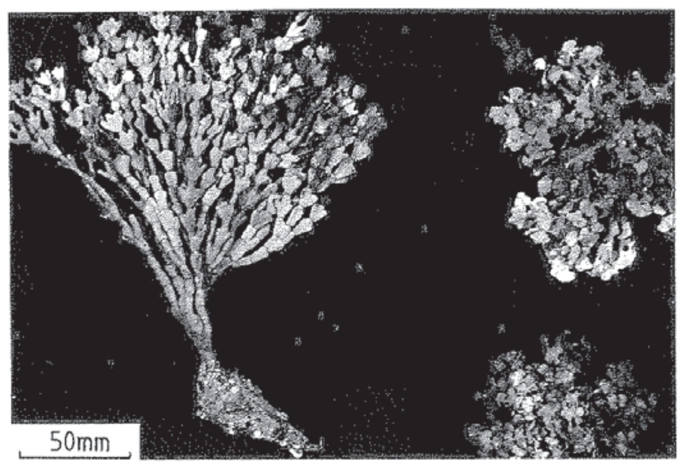


Fig. 3.7 *The green codiacean alga* Halimeda *which on death disintegrates into sand-sized particles of aragonite.* H. incrassata *on the left with conspicuous holdfast is common on the Bahama Platform and Florida Shelf.* H. opuntia *on the right is a more bushy form, typical of reef habitats of the shelf margin.*

The unstable sands of the coralgal facies occur close to the reefs and around much of the Bahama Platform margin as a belt generally 5 km in width, but reaching 40 km where exposed to much wave and storm action (e.g. northern margin of the Great Bahama Bank and south of Andros, see Fig. 3.4). These sand bodies display a hierarchy of bed forms, including sand waves, dunes and ripples. Various types of sand body can be distinguished depending on the relative roles of tidal currents, storms and waves in their formation (Ball, 1967; Hine *et al.*, 1981; Section 4.2.1).

Common organisms of the unstable sand habitat are molluscs, echinoids and some foraminifera, and in areas not constantly affected by currents, calcareous green algae such as *Halimeda incrassata* (Fig. 3.7), and *Thalassia* (turtle grass) grows in the sand.

In the more protected back-reef lagoon, areas of *Thalassia* are common and many rooted calcareous algae are associated, including *Halimeda incrassata*,

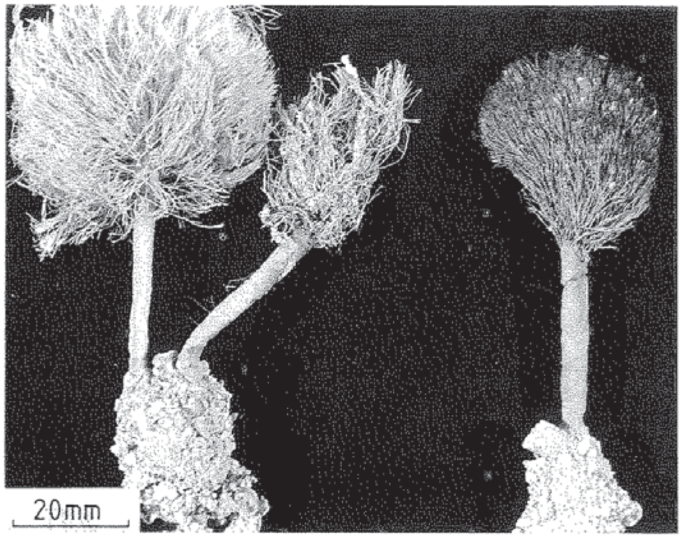


Fig. 3.8 *The green codiacean alga* Penicillus *with sediment attached to holdfast. On death this alga breaks down into micron-sized aragonite needles.*

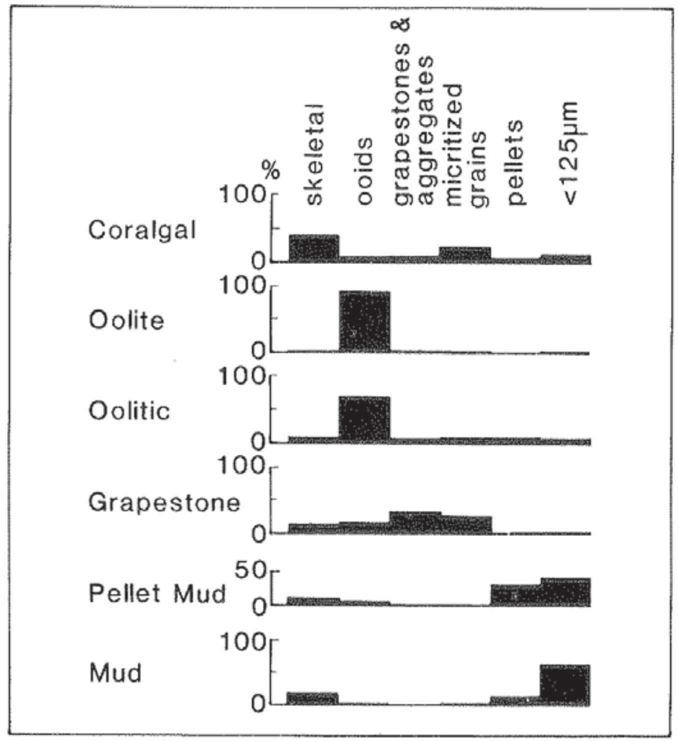


Fig. 3.9 *Compositions of Bahamian lithofacies. After Newell* et al. *(1959).*

Penicillus (Fig. 3.8), *Udotea* and *Rhipocephalus*. Molluscs, echinoids, crustaceans, foraminifera, holothurians and some small corals also occur. Sediments are lime sands and muddy sands, rippled where there is no grass and bioturbated. The effects of boring endolithic algae have commonly micritized grains, making identification difficult or impossible. A typical coralgal sand consists of 40% skeletal grains, 30% micritic grains, many of which will be altered bioclasts, 15% peloids and aggregates, and 15% others (Fig. 3.9).

3.2.1b Oolite lithofacies

The oolite lithofacies is rather localized in its distribution, occurring in some of the highest-energy shallow water (less than 3 m) locations close to the platform margin (Figs 3.2 and 3.4). Examples include Joulter's oolite shoal just north of Andros (Harris, 1979), Cat and Brown Cay south of Bimini, Lily Bank on the northeast side of Little Bahama Bank (Hine, 1977), and oolite ridges at the heads of Tongue of the Ocean and Exuma Sound. In these areas, many grains are in near constant motion and aragonite is being precipitated around them to produce the ooids (Fig. 3.10). Ooid shoals, like coralgal sand bodies, have a variable geometry depending on the importance of tidal currents and waves. Where on-platform wave and storm action is strong, linear, platform-margin parallel sand bodies develop, cut through by tidal channels with spillover lobes on their lagoonward sides (Fig. 3.11; e.g. Hine, 1977). Where tidal currents dominate sand bodies are oriented normal to the platform margins, as linear ridges up to 50 km in length, separated by troughs occupied by *Thalassia* and muddy sand. Ooid shoals may be exposed at low tide and with high rates of ooid formation they may give rise to islands. This has happened at Joulter's, where a narrow zone of

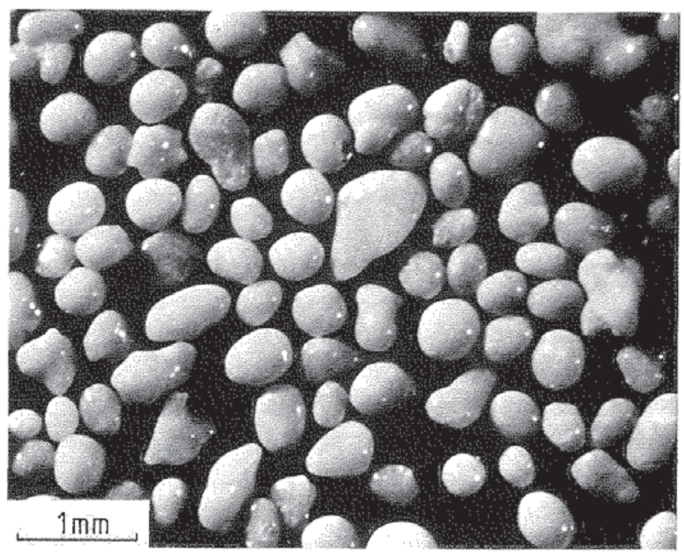


Fig. 3.10 *Bahamian ooids with high polish. Many have probably nucleated upon peloids and are spherical, whereas others probably have less regular bioclasts as nuclei. From slide courtesy of Roger Till.*

Fig. 3.11 *Oolite shoal along Bahamian Platform margin. The open ocean is to the top right; the platform lower left is covered by sea-grass and hence appears dark. Ebb tidal channels cut through the oolite shoal. Sand waves are well developed on the shoal. Lily Bank, view to NW, Little Bahama Bank. From negative courtesy of Albert Hine.*

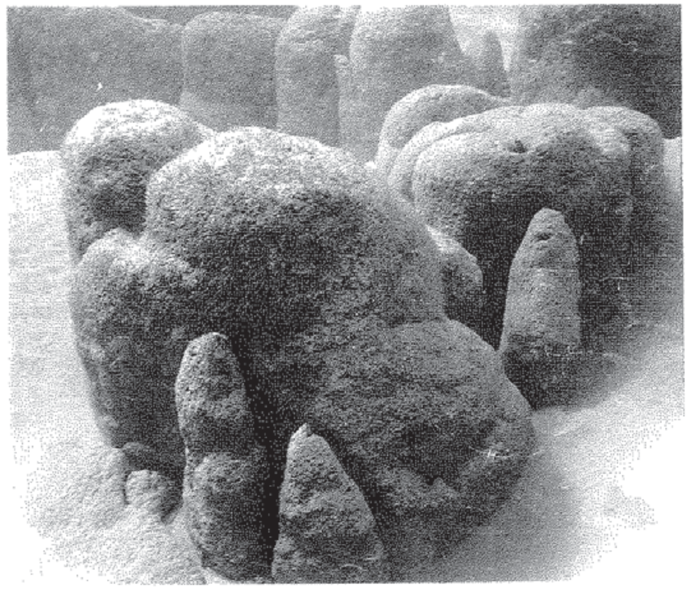


Fig. 3.12 *Large stromatolite columns, rising 1–2 m from the seafloor, in a tidal channel near Exuma Islands. Photo courtesy of Robert Dill, Gene Shinn and* Nature.

active ooid precipitation close to the platform margin led to the on-platform development of extensive sand flats cut by tidal channels, tidal bars and cays (islands), as a 400 km^2 ooid shoal complex (Harris, 1979; see Section 4.2.2).

Local seafloor cementation of ooids by acicular aragonite is giving rise to cemented crusts and hardgrounds (e.g. Dravis, 1979). These are broken up by storms to give intraclasts. The biota is limited in the active ooid sand facies by the unstable substrate. Filter feeders dominate and include the burrowing bivalve *Tivela* sp.

A recent discovery of much interest is the occurrence of giant stromatolites within tidal channels close to the Exuma Islands (Fig. 3.12; Dill *et al.*, 1986). The stromatolites, up to 2 m high, are growing in 7–8 m of water and occur in an area of oolitic–peloidal sand waves, where currents reach 1 m s^{-1}. They are individual columns and large, coalesced bioherms, elongated perpendicular to tidal flow. Many are asymmetric, growing into the incoming tide. They have smooth to pustular surfaces and an internal structure of convex-up laminae which either define small columns (10–40 mm across) or encompass the whole structure. The sediment trapped by the microbial film, which includes blue–green algae and diatoms, is ooid and pelletal sand. The stromatolites are being cemented by acicular aragonite and this has produced a tough rock some 0.3 m below the structure's surface.

3.2.1c Oolitic and grapestone lithofacies

The oolitic and grapestone lithofacies covers large areas of the platform from water depths of around 9 m to just exposed at low tide. Currents and wave action are generally sufficient to prevent much deposition of lime mud, although the lime sand itself is moved only during storms. This is the stable sand habitat since much of the sediment is covered by a thin scummy layer of algae and diatoms (Fig. 3.13) or by *Thalassia* grass (Fig. 3.14A). The surficial algal mat (Bathurst, 1967) reaches 10 mm in thickness and is dominated by the blue–green alga *Schizothrix*. The mat contains within it many minute organisms, such as small molluscs, foraminifera, annelids, nematodes, ostracods and diatoms, as well as trapped sedimentary particles carried on to the mat during storms. The surficial mat, turtle grass and rooted codiacean algae all have a stabilizing effect on the sediment and are able to prevent erosion by quite strong currents (Neumann *et al.*, 1970; Scoffin, 1970). Sea-grass protects the seafloor from tidal currents up to 0.7 m s^{-1} by being bent over and laid flat, but extensive sediment erosion begins when currents exceed 1.5 m s^{-1} (without grass, extensive erosion takes place at velocities of 0.5 m s^{-1}). Erosion takes place more easily in wave-driven, oscillatory current regimes, since the grass blades are then swept back and forth (Scoffin, 1970).

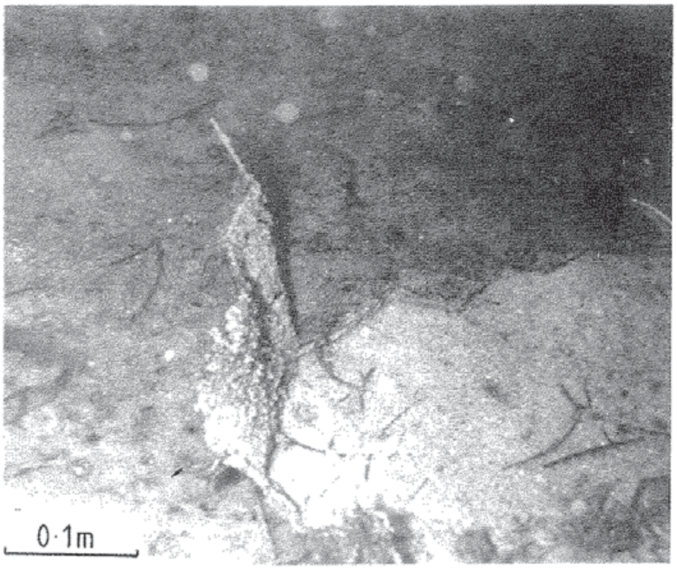


Fig. 3.13 *Surficial algal mat from a lagoonal area in 3 m of water. Part of the mat has lifted off the sediment surface. Microbial mats like this protect the seafloor from erosion during storms.*

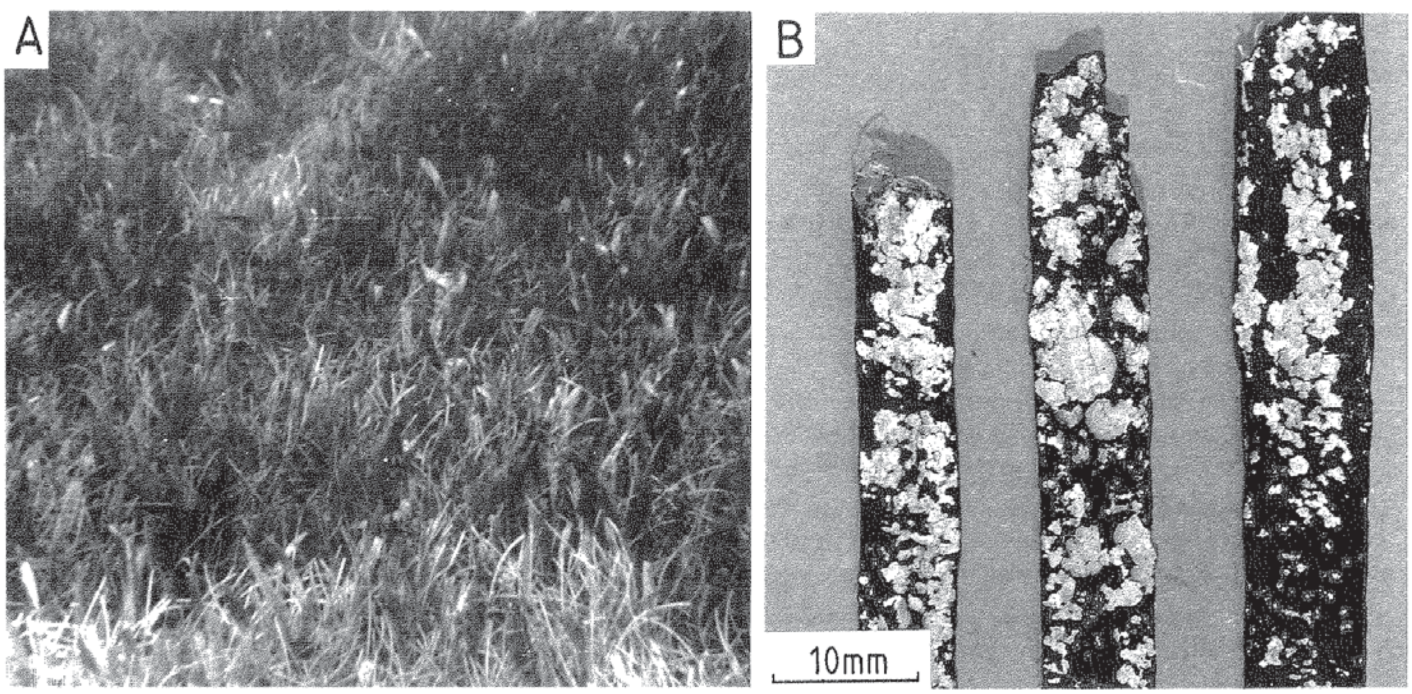
A rich and diverse fauna occurs in areas of this lithofacies, with skeletal grains coming from bivalves, gastropods, echinoids, starfish, foraminifera and small corals. The aragonitic *Halimeda incrassata* (Fig. 3.7) and related forms disintegrate to give sand-sized particles, and *Penicillus* (Fig. 3.8), *Udotea* and *Rhipocephalus* supply lime mud. Many small organisms, such as foraminifera, serpulids and melobesoid algae (epibionts) live upon the *Thalassia* blades (Fig. 3.14B) and contribute carbonate to the sediment. Ooids are washed in from platform-margin ooid shoals during storms. Grapestones are aggregates of grains cemented on the seafloor by micritic aragonite (Fig. 3.15; also see Section 1.2.3, Fig. 1.8). Faecal pellets are common. Much of the sediment consists of sand-sized micritic grains (Fig. 3.9) which have been produced by the boring activities of endolithic algae on skeletal fragments (Bathurst, 1966). An early stage in this micritization is the development of a micrite envelope (see Section 1.3, Fig. 1.11).

Sand waves and ripples occur in some areas of the oolitic and grapestone facies but they are mostly only active during severe storms. Bioturbation is prevalent, especially by crustaceans such as *Callianassa*.

3.2.1d Pellet mud and mud lithofacies

The pellet mud and mud lithofacies occurs in the most protected part of the platform, depths usually less than 4 m, such as to the west of Andros Island and in the Bight of Abaco. It occurs in areas where tidal

Fig. 3.14 *(A)* Thalassia *sea-grass from protected lagoonal area. Calcareous and other algae occur within these grassy areas, which protect the seafloor from erosion during stroms. (B) Blades of* Thalassia *with epibionts, mainly melobesoid (red) calcareous algae.*



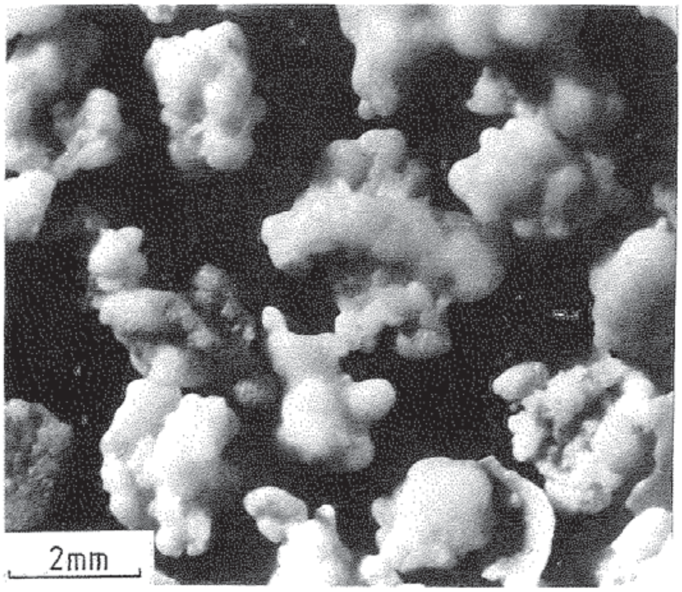


Fig. 3.15 *Grapestones: aggregate grains consisting of bioclasts and peloids cemented by micritic aragonite. From slide courtesy of Roger Till.*

currents and waves are extremely weak, and there are no cross-bank currents, so that sediments are only affected during major storms. The sediment largely consists of aragonite mud and faecal pellets (50–200 μm long) composed of this mud (Fig. 3.16), produced mainly by polychaete worms, but also gastropods. The lime mud itself consists of aragonite needles, a few microns in length. There are few ripples and much of the sediment is bioturbated, especially by the crustacean *Callianassa*, which produces distinctive conical mounds on the seafloor and a simple branching burrow system (Fig. 3.17). Turtle grass is widely distributed but generally sparse, and a surficial algal mat may cover the sediment surface. On the whole, the fauna is sparse and of low diversity, with molluscs the most important group. Echinoids, a soft bottom-living coral (*Manicena*), sponges and tunicates also occur.

There has been much discussion over the origin of Bahamian lime mud, with the controversy centring on an inorganic versus algal origin (see Bathurst, 1975 for a review). Cloud (1962) argued from chemical evidence and the low standing crop of calcareous algae in the mud facies area that the aragonite needles were a direct precipitate from seawater as a result of evaporation. However, later studies in the Bight of Abaco by Neumann & Land (1975), and also in south Florida by Stockman *et al.* (1967), were able to show that more than enough aragonite is produced by the disintegration of calcareous codiacean algae such as *Penicillus*, *Rhipocephalus* and *Udotea* to account for lime mud in the lagoons. In fact, overproduction was such that it could explain lime mud occurring on tidal flats and in the periplatform ooze (Section 3.2.1e). Recent work by Loreau (1982) concluded that much of the lime mud of the Great Bahama Bank is a direct precipitate, whereas that of Florida Bay is largely algal in origin. From examination by scanning electron microscope (SEM), Loreau showed that only 25–40% of crystals in codiacean algae are needles, the rest are equant nannograins. The lime mud of the Bahama Platform consists of 90% needles (see Fig. 3.18), whereas that of Florida Bay has much less. There is also a chemical difference between algal aragonite and inorganically-precipitated aragonite, which is shown by Florida Bay and Great Bahama Bank muds: the atomic Sr/Mg ratio of aglae is less than 2, whereas it is more than 4 in inorganic aragonite.

Quite commonly on the Bahama Platform (and in Florida Bay), the sea takes on a milkiness due to suspended material, mostly aragonite needles. These *whitings* appear to be the actual inorganic precipitation of aragonite taking place (Shinn *et al.*, 1989), although stirring up of bottom sediments by shoals of fish can produce the same effect.

3.2.1e Periplatform ooze

Periplatform ooze occurs on the slopes around the Bahama Platform and consists of platform-derived shallow-water mud and sand mixed with pelagic carbonate, mostly planktonic foraminifera and coccoliths, with some pteropods (Schlager & James, 1978; Mullins *et al.*, 1984, 1985). The shallow-water material is mostly taken off the platform during storms, but resedimentation processes of slumping, debris flows and turbidity currents are important, especially in moving sediment from the upper part of the slope into deeper water. The platform component of the ooze is mostly composed of aragonite and high-Mg calcite, contrasting with the dominantly low-Mg calcite mineralogy of pelagic carbonate. Monitoring the composition of the periplatform ooze shows that there is a linear decrease in the contribution from the platform with increasing distance from the shallow-water source (Heath & Mullins, 1984). Periplatform oozes are being lithified on the seafloor and in the shallow subsurface on the slopes around the Bahama Platform (Schlager & James, 1978; Dix & Mullins, 1988), and in some areas this is generating nodular structures (Mullins *et al.*, 1980, 1985). This slope

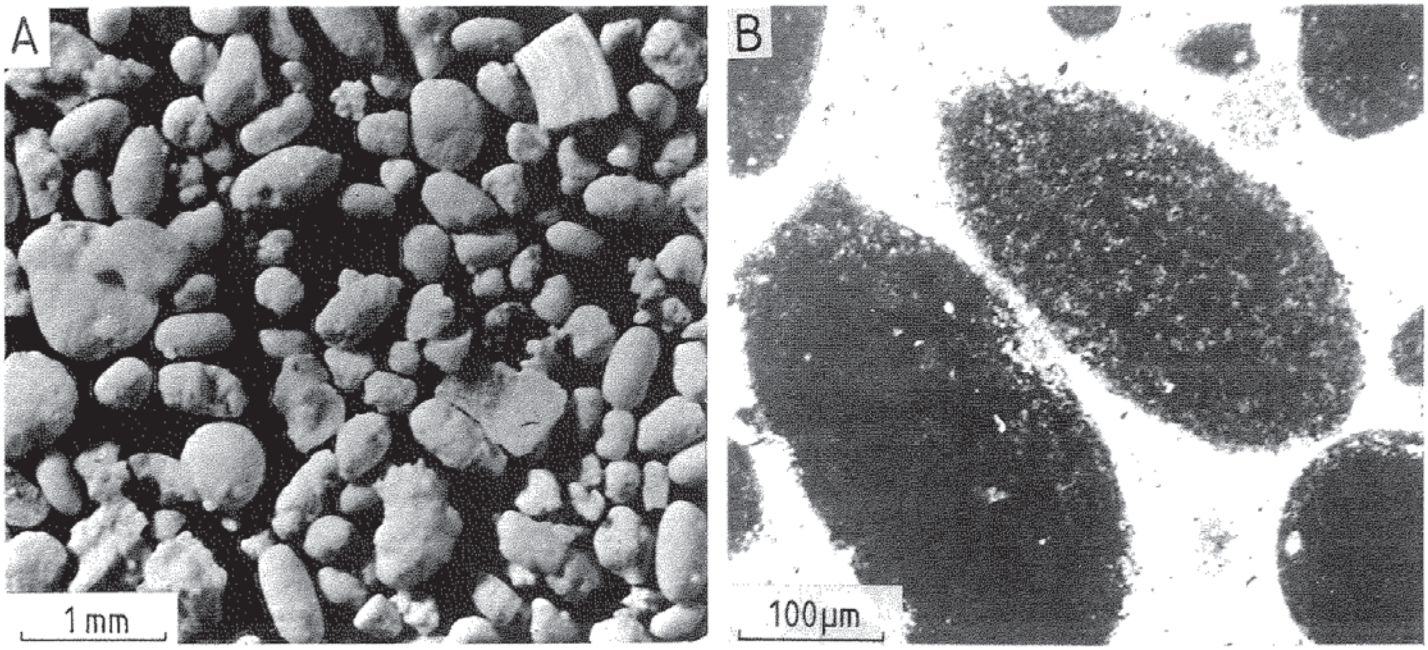


Fig. 3.16 *Peloidal skeletal sand grains washed out of sandy mud. (A) Surface view. (B) Photomicrograph of peloids showing homogeneous micritic nature. From slides courtesy of Roger Till.*

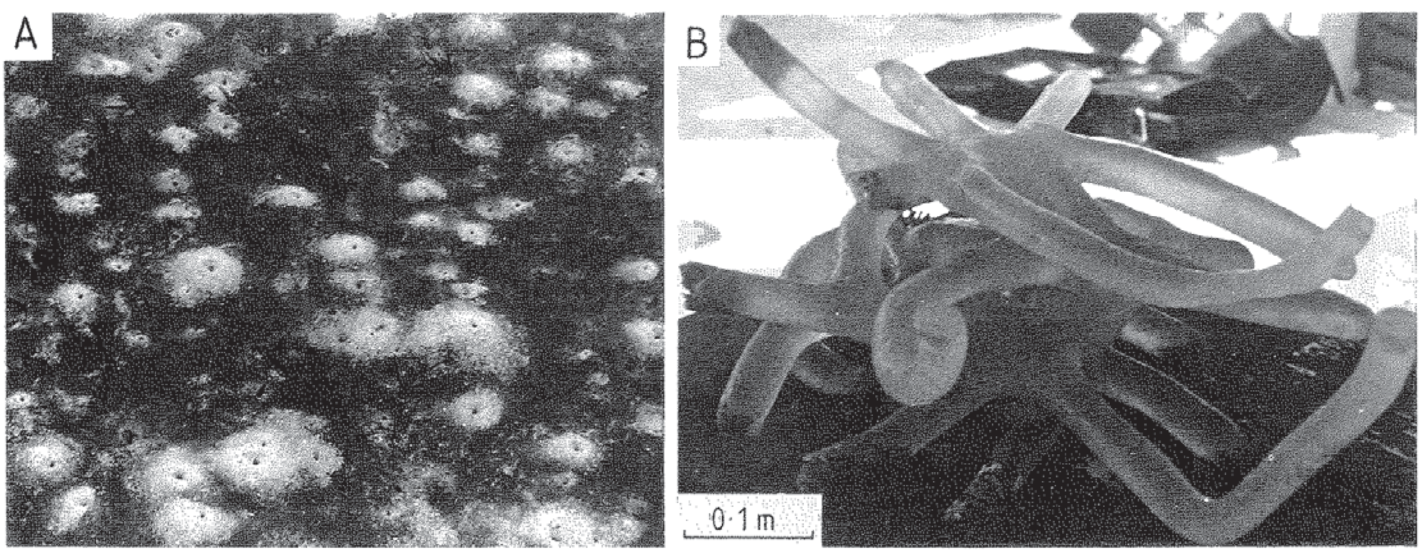


Fig. 3.17 *(A) Conical mounds, 0.1–0.2 m across, produced by the crustacean* Callianassa, *in 0.5 m water depth. The mounds are composed of white muddy sand brought up by the crustaceans on to a lagoon floor covered by a thin, dark surficial microbial mat. (B) Burrow system of* Callianassa *preserved in resin.*

lithofacies and the associated resedimented carbonates are considered further in Section 5.11.

The distribution of subtidal lithofacies on the Bahama Platform largely reflects water movement high-energy platform margins with reefs, associated skeletal sands and oolite shoals, contrast with stabilized sands, pelleted muds and lime muds of the quieter-water platform interior. Most sediments are deposited close to where constituent grains were formed, and long-distance sediment transport is only important at the platform margins where material may be taken on to the adjoining slope to basin by storms, or transported deep into the basin by turbidity currents and other resedimentation processes. Microbial micritization of skeletal debris is widespread and

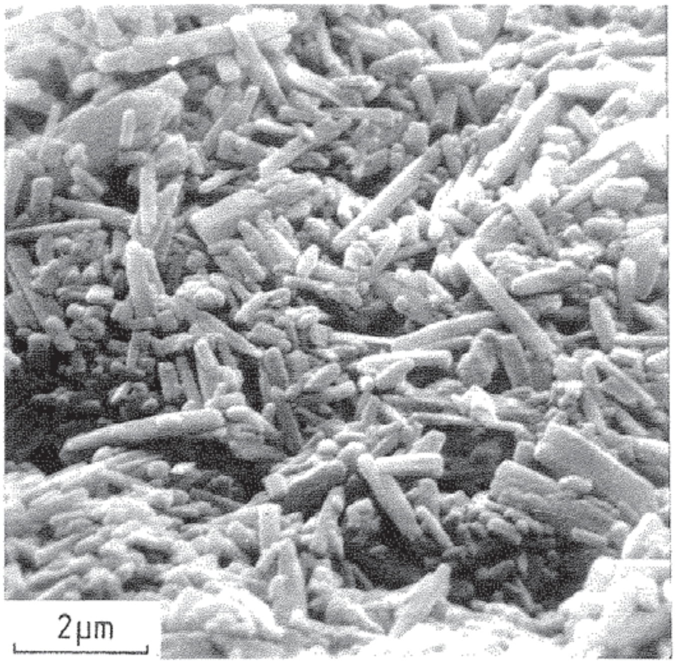


Fig. 3.18 *Bahamian mud. SEM view showing predominance of aragonite needles. Photo courtesy of Jean-Paul Loreau.*

inorganic processes of precipitation are important, giving aragonite needle muds, ooids and seafloor cementation of sands to form grapestones and hardgrounds. In spite of this, biogenic processes of carbonate formation still contribute most material to the sedimentary package.

3.2.2 **Intertidal–supratidal carbonate sediments of the Bahamas**

A wide variety of lithofacies are deposited in intertidal and supratidal settings on the Bahama Platform and they fall into two broad groups: those of relatively high-energy shorelines where narrow sandy beaches are backed by aeolian dunes, and those of low-energy shorelines where broad, laterally extensive muddy tidal flats are backed by supratidal freshwater marshes.

3.2.2a High-energy shorelines

High-energy shorelines are developed along windward coasts such as the eastern sides of Andros, the Berry Islands, Great Abaco and Eleuthera. Lime sand comprising the beach is derived from the shoreface and most consists of skeletal debris and ooids. Sedimentary structures are identical to those of clastic beaches: flat-bedding in truncated sets dipping seaward at a low angle, with some wave ripple cross-lamination and crustacean burrows. Shoreface (low tide to wave-base) sands are also rippled and bioturbated. Synsedimentary cementation in the intertidal zone gives rise to beachrock (Scoffin & Stoddart, 1983), well known from Bimini. In the backshore, beach berms and dunes are the result of storm waves and onshore winds blowing sand off the beach. Anchoring of sand by vegetation is common and soils may develop, giving rise to crusts, local cementation and rhizocretions. Shoreline carbonate sands are discussed at length in Section 4.1.

3.2.2b Tidal flats on west side of Andros

The tidal flats on the west side of Andros Island have been the focus of numerous studies (for example, Black, 1933; Shinn *et al.*, 1965; Shinn *et al.*, 1969; Gebelein, 1974a; Hardie, 1977; Multer, 1977; Gebelein *et al.*, 1980; Shinn, 1983a). In this area, the tidal range is very low (0.46 m) and wind–wave activity is weak since Andros Island acts as a barrier to the dominant easterly winds. Occasional winter storms from the west to north produce strong waves in spite of the shallowness of the platform.

The Andros tidal flats are complex with many subenvironments including tidal channels, beach ridges (hammocks), levees, ponds, intertidal flats themselves, areas of surficial crusts, areas of algal mats, mangrove clumps and swamps, and freshwater algal marsh (Figs 3.19 and 3.20). Parts of the tidal flat are permanently subaqueous, the ponds and channels for instance, whereas other areas are exposed for some of the tidal cycle or for certain seasons of the year (Fig. 3.23). To describe the fluctuations in water cover, an exposure index was introduced by Ginsburg *et al.* (1977) to indicate the percentage exposure of a subenvironment over a year (Fig. 3.21, also see Section 4.3).

Two distinct types of tidal flat occur on the west side of Andros Island (Gebelein, 1974a; Hardie, 1977): to the northwest of Williams Island, the tidal flats are 5 km wide and are dissected by many tidal channels (comprising 15% of the flat complex) which drain ponds, flats and algal marshes (Figs 3.19, 3.20A and 3.22); to the southwest of Williams Island, the tidal flats are up to 35 km wide but they have few tidal channels, and consist instead of broad depressions separated by former beach ridges rising 1–2 m above normal high water (Fig. 3.20B). The depressions are variably occupied by water to form ponds which are surrounded by intertidal flats, algal marshes and areas of surficial crust.

Fig. 3.19 *Aerial view of the intertidal flats on the northwest side of Andros Island, Great Bahama Bank, showing tidal channels, levees, patchy areas of mangroves and algal mats (dark) and depressions, ponds and flats (grey areas). Photo courtesy of Ian Goldsmith.*

At their seaward margins, the Andros tidal flats have a low beach ridge which is constructed of sediment thrown up from the shallow subtidal zone during storms. The sediment of the present and former beach ridges is largely skeletal–peloidal sand with fine laminae and small irregular and laminoid fenestrae. Levee sediments along channel banks are similar.

Sedimentation on the tidal flats mostly takes place during major storms, which are very sporadic. Waters move on to the tidal flats via the tidal channels every tidal cycle, but these waters are clear, carrying little suspended sediment. During storms, sediment from the adjacent nearshore is put into suspension and carried on to the tidal flats via the channels, to be deposited as a thin pelletal lime mud blanket. In the ponds and low intertidal parts of the flats, this thin layer is mixed into the sediment packet by burrowing and grazing animals (gastropods, annelids and crustaceans). In the high intertidal and supratidal areas, the storm layer is incorporated into algal mats.

Channels meander across the tidal flat and rework the sediments. Erosion takes place on the outside of meander bends and deposition occurs on point bars and levees. Intraclasts occur in channel bottoms along with many gastropods. Pelleted lime muds occur in quieter reaches of the channels. The ponds and intertidal areas are composed of pelletal lime muds, with polychaetes and gastropods providing the pellets. Birdseye vugs are common in higher tidal flat sedi-

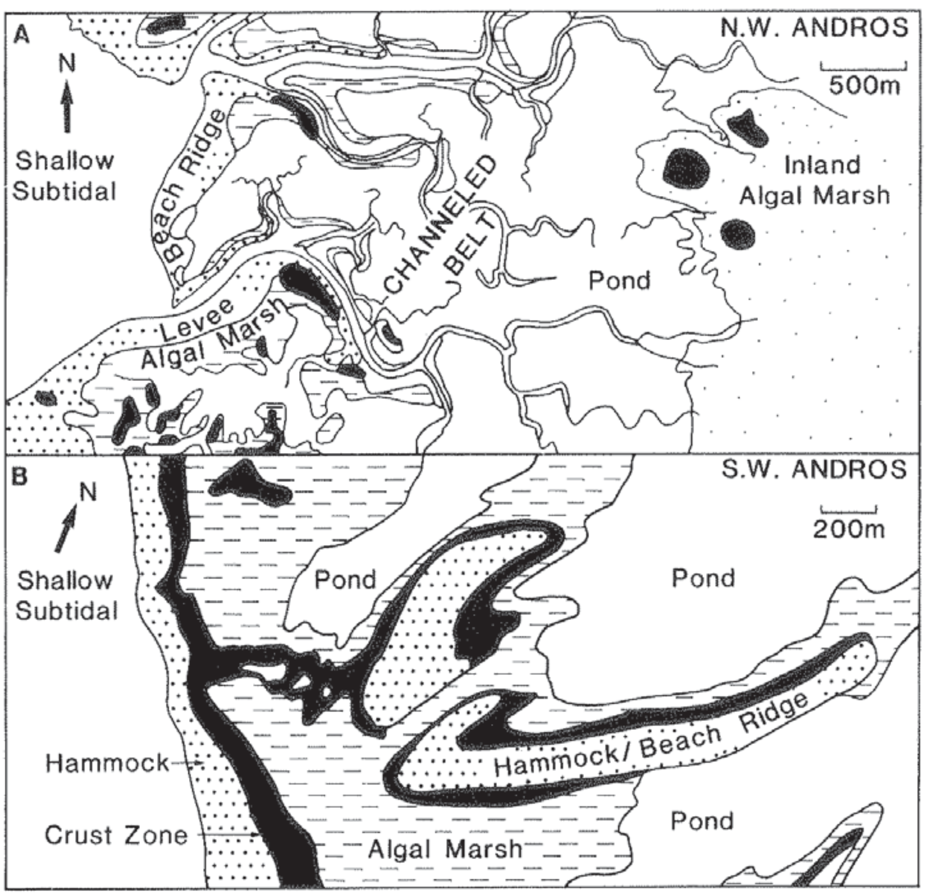


Fig. 3.20 *Tidal flat environments of western Andros, Great Bahama Bank. (A) Northwest Andros, with well-developed channelled belt. After Hardie (1977). (B) Southwest Andros, with beach ridges. After Gebelein* et al. *(1980).*

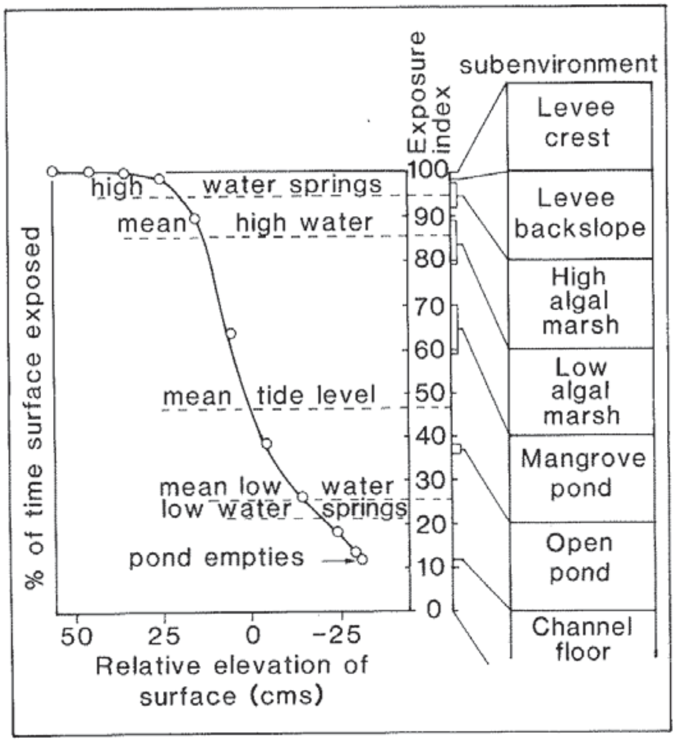


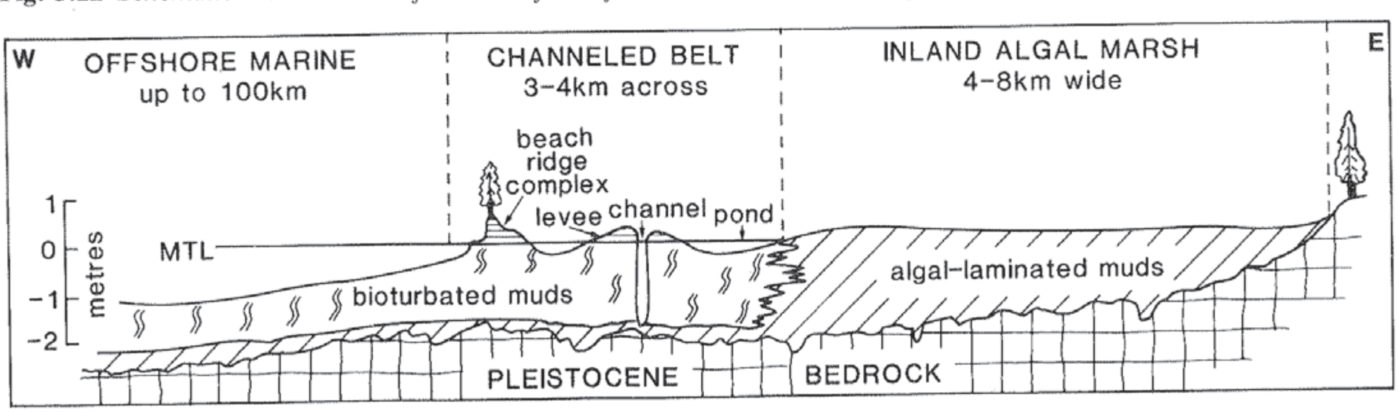
Fig. 3.21 *Exposure index (percentage of time tidal flat surface exposed) and subenvironments of western Andros. After Ginsburg* et al. *(1977).*

ments along with root-moulds and burrows (Fig. 3.23; Shinn, 1968a, 1983b). Faunal diversity is low on the tidal flats; apart from numerous gastropods, benthic foraminifera and polychaete worms are common. Microbial (algal–bacterial–diatom) mats (Fig. 3.24) are widespread in the upper intertidal and supratidal zones of the channelled belt of Andros and these give rise to stromatolitic laminae by the trapping of sediment within the mat. Laminoid fenestrae are common between microbial layers. The organic mats are often broken up or disrupted into polygonal structures by desiccation; small domes may develop through buckling up of the mat. Thin graded storm beds are common in the supratidal sediments, where there are few burrowing organisms to destroy them.

In the freshwater, inland algal marsh (Black, 1933; Monty, 1972; Monty & Hardie, 1976), an area 4–8 km wide and 50 km long, the microbial mats are dominated by the tufted blue–green alga *Scytonema*, which can form a peat several centimetres thick. A range of mat types is again developed, from flat sheets to polygons and domes, with exposure and desiccation determining gross morphology. Periodically, storms bring marine sediment on to the freshwater marsh, covering the microbial mat with a peloidal–foraminiferal layer. Another alga *Schizothrix* then colonizes the surface forming a thin sheet before the *Scytonema* mat is re-established. An important feature of the freshwater marsh is that calcification of algal filaments takes place here (Monty & Hardie, 1976). Micritic calcite (low Mg) is precipitated around the algal filaments and within the mats through evaporation and biochemical effects. This ‘algal tufa’ does not form in microbial mats of areas frequently inundated by seawater.

In the high intertidal and supratidal zones, aragonite and dolomite are being precipitated to cement the surficial sediments and form crusts (Fig. 3.25; Shinn *et al.*, 1965; Shinn, 1983a). This is happening on the sides of beach ridges and levees and around ponds. Frequently crusts are broken up to produce intraclasts and these can be reworked to form edge-wise conglomerates or flakestones. Dolomitization of surficial sediments is probably an evaporitic process, caused by porewaters with increased Mg/Ca ratio resulting from precipitation of aragonite and possibly gypsum in the sediment. The dolomitic crusts form just above high tide mark where marine groundwaters are drawn up to the surface by capillary action and

Fig. 3.22 *Schematic cross-section of the tidal flats of northwestern Andros. After Hardie (1977).*



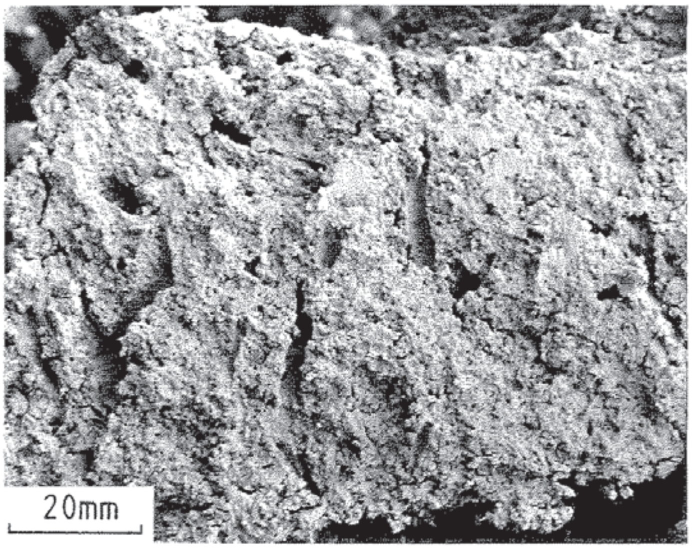


Fig. 3.23 *Tidal flat sediment showing crumbly, peloidal muddy sediment with small irregular fenestrae (birdseyes), some horizontal, laminoid fenestrae, and prominent burrows.*

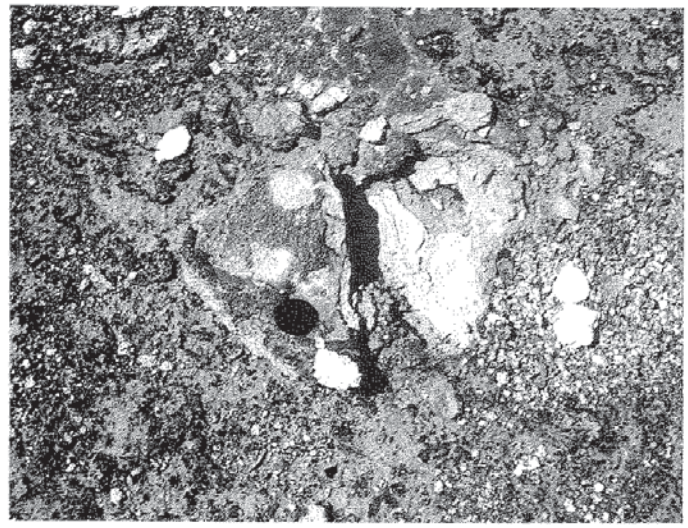


Fig. 3.25 *Cemented dolomite crust and surface covered with small intraclasts. Andros tidal flat. Lens cap 6 cm diameter. Photo courtesy of Ian Goldsmith.*

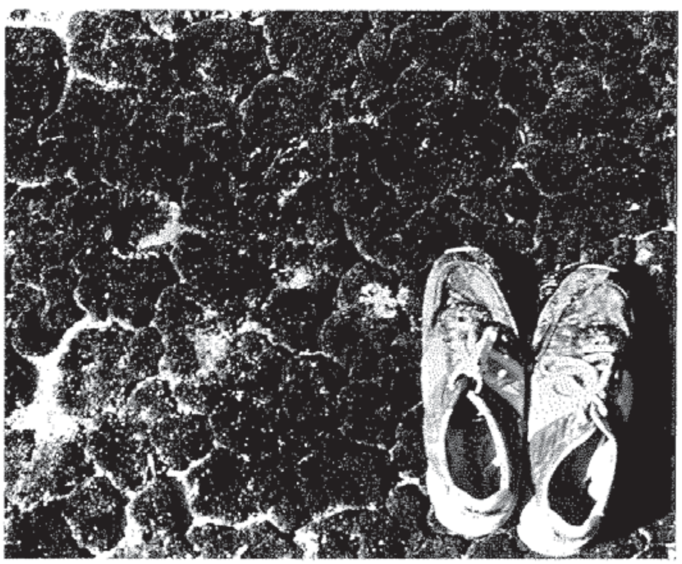


Fig. 3.24 *Algal mats with polygonal cracks from Andros tidal flat. Photo courtesy of Ian Goldsmith.*

evaporated. The dolomite is Ca-rich, poorly ordered and very fine grained (crystals 2–4 μm). A poorly-ordered dolomite is also forming in the subsurface of the southwestern Andros tidal flats (Gebelein *et al.*, 1980). It is possible that the dolomite is forming where marine groundwaters are mixing with fresh-water beneath porous beach-ridge sediments (also see Section 8.7.3).

Cores taken through the tidal flats on the northwest coast of Andros reveal a thin (3 m) seaward-thickening sediment wedge over Pleistocene limestone (Fig. 3.22; Shinn *et al.*, 1969; Gebelein, 1974a; Shinn, 1983a). Immediately overlying the Pleistocene is a marsh deposit (peat) laid down prior to and during the Holocene sea-level rise across the platform, and this passes into the modern algal marsh deposits at the back of the present-day tidal flat. Above the basal transgressive marsh deposit, there occurs a shallowing-upward sequence of bioturbated pelleted muds with a good marine fauna (offshore subtidal) passing up into pelleted muds with a restricted fauna and vertical burrows (low intertidal) and then algal laminated sediments of the upper intertidal. The Holocene sequence upon the Pleistocene thus records the initial transgression (marsh deposits, onlapping), followed by depositional regression as the tidal flats began to prograde seawards. It is thought that progradation took place until about 1000 years BP and that now the tidal flats are being eroded (Gebelein, 1974a). It has been suggested that this erosional state accounts for the abundance of tidal channels in this northwestern area of Andros. In fact, the orientation of this north-western coastline is such that it receives the full force of storm winds and waves, i.e. the meteorological tidal range can be high (Gebelein, 1974a; Hardie, 1977). It also appears that the adjacent subtidal area has a low sediment production rate. By way of contrast, the tidal flats on the southwest side of Andros face away from the direction of major storms and have continued to prograde until the present time, although the rate appears to have slowed down over the last 1000 years (Gebelein, 1974a). The few tidal channels may reflect the more protected nature of this shoreline. The tidal flats here appear to have prograded seaward in discrete jumps, with the old shorelines marked by hammocks (former beach ridges). The depressions between hammocks are being filled with pond and bay sediments (low intertidal) and then by algal laminites (high intertidal) and supratidal sediments (marsh with algal tufa).

3.3 RECENT CARBONATES OF THE FLORIDA SHELF

The Florida carbonate shelf extends some 300 km south and southwest from Miami curving towards the west past Key West to Dry Tortugas (Figs 3.26 and 3.27). A discontinuous string of elongate islands, the Florida Keys, delineates the inner shelf margin. Behind these islands and connected to the shelf by tidal channels is Florida Bay, a very shallow region with numerous mud banks and mangrove-covered islands. The shelf itself is 5 to 10 km wide with a shelf-break in 8–18 m of water where a seaward slope of 1–10° descends and then gradually flattens into the Straits of Florida with a water depth of 800–1000 m. Along some parts of the shelf margin there is a deeper shelf in 200–400 m of water (the Pourtales Terrace and the Miami Terrace) with a generally steeper slope (8–18°) up to the modern shelf. The deeper terraces are thought to be remnants of a Miocene carbonate platform with marginal reefs. The Florida Shelf–Florida Bay region is underlain by Pleistocene limestones which were deposited during the last high stand of sea-level (the Sangamon), some 120 000 years ago (Perkins, 1977). The present topography of south Florida broadly reflects the Pleistocene facies distribution, and on a local scale, the karstic weathering which affected the Pleistocene limestones after their deposition, when sea-level fell to around −100 m. The antecedent topographic control is well seen around Miami, where the small hills and troughs (glades) reflect an oolite beach-barrier and tidal channel system in which the Pleistocene Miami Oolite was deposited (see Section 4.2.3; Evans, 1984). The elongate Middle and Upper Florida Keys, which run parallel to the shelf margin, are formed upon a Pleistocene patch reef complex (the Key Largo Limestone) and in the Lower Keys (Big Pine to Key West), the more shelf-normal arrangement of islands (well seen in Fig. 3.27) is due to the underlying Pleistocene oolite having been deposited in a more tide-dominated oolite complex (Section 4.2.3). Small lagoons and bays occur within the Keys belt, particularly in the Lower Keys, where the configuration of the rocky islands provides protection from waves and tides. Strongly affecting Holocene carbonate sedimentation, the modern shallow shelf-break is the former site of a Pleistocene shelf-marginal reef.

The prevailing winds in southeast Florida (Fig. 3.26) are the same as affect the Bahamas; southeast trade winds during the summer, and winds from the northeast during the winter. These generate onshore waves and currents which can be strong along the shelf margin (0.5 m s^{-1} bottom currents), but are much weaker on the inner shelf. For much of the time, wave-base on the shelf is very shallow, at less than 3 m. The Gulf Stream moves northwards in the Florida Straits at a mean velocity of around 1.3 m s^{-1} and this gives rise to a southward-flowing counter current along the shelf margin. The tidal range along the Florida Shelf is very low, at 0.7 m, so that tidal currents are only significant within and near the channels which connect Florida Bay with the shelf. Of great importance to sedimentation along the Florida Shelf, and in the Caribbean generally, are the effects of hurricanes. These short-lived intense but rare storms give rise to extreme waves and currents, which can break large coral stands, transport much coarse sediment, cut channels through barrier islands and mud banks, and put vast quantities of lime mud into suspension (Ball *et al.*, 1967; Perkins & Enos, 1968).

Water temperatures on the Florida Shelf generally range from 18 to 30°C, but occasionally much lower values pertain and these can have a detrimental effect on the corals. Salinities on the shelf are generally the normal marine values of 35–38‰. However, in Florida Bay salinity ranges from 6 to 58‰ with the low values reflecting extreme freshwater run-off from the Everglades during a very wet summer. Lowered salinities can then occur on the shelf, especially in the vicinity of the tidal channels between the Keys, connecting the bay to the shelf. Reviews of the Florida Shelf carbonates are given in Ginsburg (1956, 1964), Bathurst (1975), Multer (1977) and Enos (1977a).

3.3.1 Subtidal carbonate sediments of the Florida Shelf

The carbonate sediments of the Florida Shelf are almost entirely biogenic in origin. There are no oolites and much of the lime mud is probably of codiacean algal origin. Pleistocene limestone lithoclasts make up a small percentage of the carbonate sediments, and a

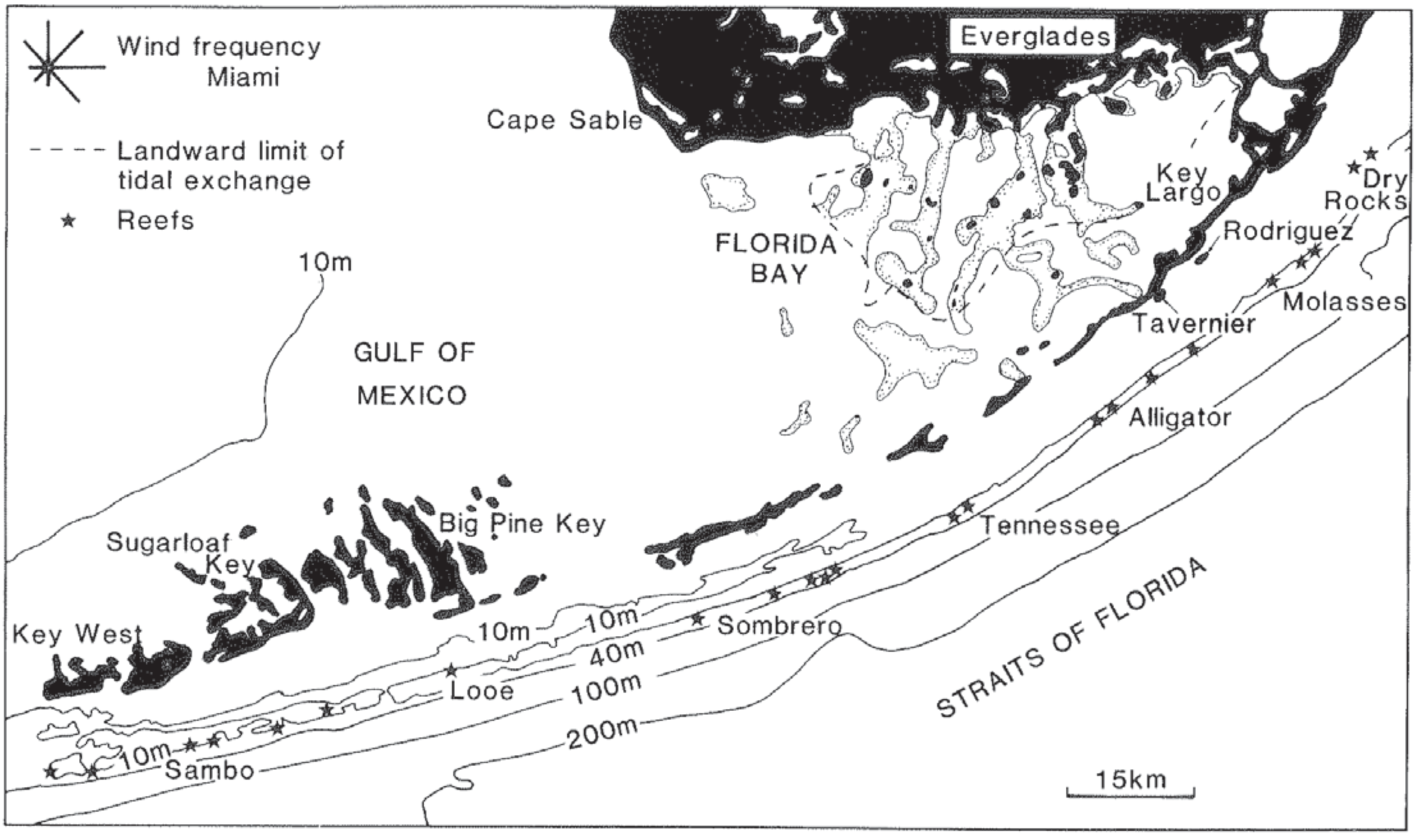


Fig. 3.26 *Map of southern Florida showing the Florida Keys located along a line of Pleistocene reefs and oolite shoals, the modern reefs along the shelf-break, the two mud banks at Tavernier and Rodriquez, and Florida Bay with network of banks and 'lakes'.*

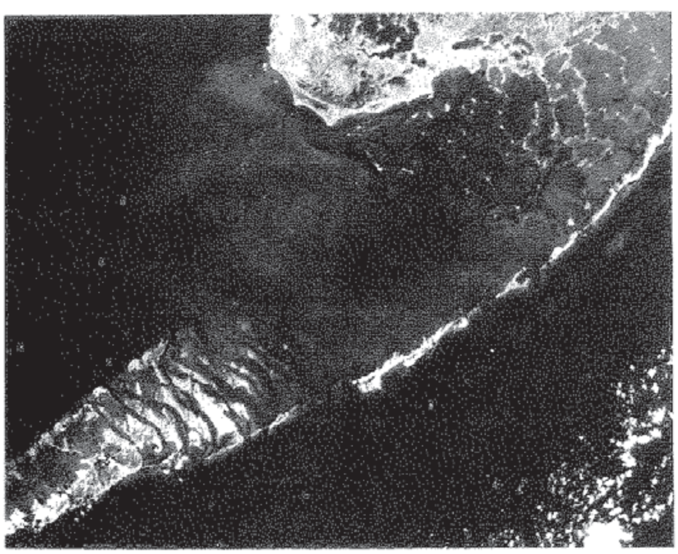


Fig. 3.27 *Satellite photo of southern Florida showing Florida Bay with network of 'lakes' and banks, the linear Upper and Middle Florida Keys which are present-day islands developed upon a line of Pleistocene patch reefs (Key Largo Limestone) and the Lower Florida Keys which are modern islands developed upon a Pleistocene tide-dominated oolite shoal complex (the Miami Oolite).*

little quartz is present, increasing northwards to form a substantial part of the sediment in the Key Biscayne region. As for the Bahamas, the sediments of the Florida Shelf have been discussed in terms of their lithofacies, and organism communities and habitats have been defined (Enos, 1977a; Multer, 1977). Enos recognized eight habitat communities: (1) rock and dead reef, (2) lime mud, grass-covered or bare, (3) lime sand, grass-covered or bare, (4) patch reef, (5) outer reef, (6) fore-reef muddy sand, (7) shoal fringe (mound), and (8) reef rubble. Each habitat has a distinctive group of organisms living there, although many organisms can live in a number of habitats. The habitats are broadly zoned across the Florida Shelf (Fig. 3.28) and there is an associated variation in sediment composition across the shelf and into Florida Bay (Fig. 3.29).

3.3.1a The rock and dead reef habitat

The rock and dead reef habitat occurs along the inner shelf margin where Pleistocene limestone crops out, from the shoreline to up to 3 km seawards. Many

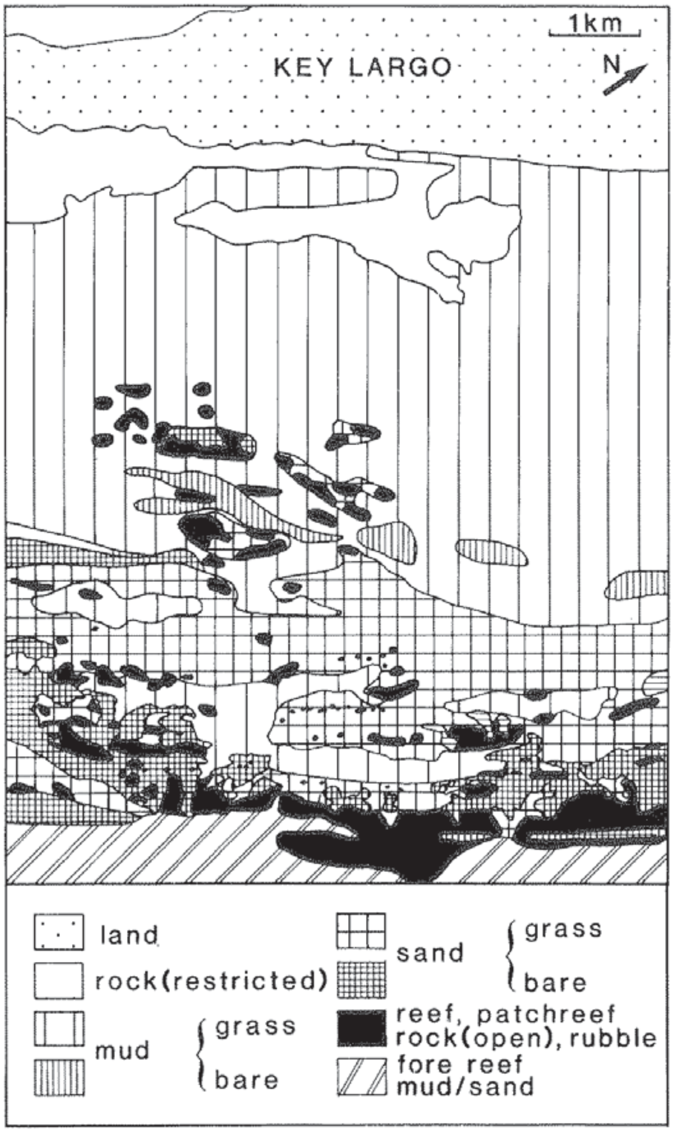


Fig. 3.28 *Lithofacies distribution across the Florida Shelf off Key Largo. After Enos (1977a).*

organisms live in this area, particularly those liking a hard substrate and able to tolerate the low energy of the inner shelf, such as some corals (*Siderastrea, Porites*), sponges, echinoids, gorgonians and some algae. Patches of sand and muddy sand are common.

Dead reef is common along the outer shelf margin and forms a similar hard substrate for encrusting organisms. Red algal sheets are particularly common over dead coral and play a major role in generating reef rock. Bioerosion of dead coral by boring bivalves, sponges and algae is intensive. Some areas of dead reef along the shelf edge are situated opposite major channels draining from Florida Bay. These reefs, such as Alligator, were living until about 4000 years BP. Florida Bay came into existence from about that time, and water draining out from the bay on to the shelf through gaps in the chain of islands (the Florida Keys) had a detrimental effect on the outer reefs.

3.3.1b Lime mud and sandy mud habitat

Areas of lime mud and sandy mud habitat mostly occur along the inner shelf margin where water circulation and wave action is at a minimum. Where water depths are less than about 8 m, there is a dense cover of turtle grass. Many rooted algae grow here, such as *Penicillus*, *Halimeda incrassata* and *Udotea*, and they contribute vast quantities of aragonite needles and sand-sized grains to the sediment. Molluscs and echinoids are common, along with some soft sediment-living corals and red algae such as *Goniolithon*. Crustacean burrows are abundant. Small bays and lagoons in the Lower Florida Keys area (e.g. Fig. 3.30) such as Coupon Bight, are also locations of lime mud deposition (see Section 4.1.5) but the faunal diversity is lower.

As on the Bahama Platform, the dense sea-grass carpets stabilize mud and sand, and baffle waves and currents; they also trap sediment. The stabilization effect of sea-grass is clearly demonstrated when hurricanes strike the Florida Shelf (Ball *et al.*, 1967). Little erosion takes place in the sea-grass areas, whereas areas of lime sand and the outer reefs are considerably affected. Storms can generate 'blowouts' in grass-covered areas (Wanless, 1981), and the subsequent migration and recolonization of these flute-shaped hollows gives rise to fining-upwards units, around 1 m thick. A shell lag begins the unit, formed in the base of the erosional hollow, and this is overlain by skeletal sand and then sandy mud trapped by the grass-covered advancing leeside of the hollow.

3.3.1c Lime sand

Lime sand occurs in a belt several kilometres wide inshore from the shelf margin. Sea-grass colonizes areas of lower wave energy; elsewhere sand is rippled and sand waves occur. The latter are well seen on White Bank, 3–5 km offshore from Key Largo, and 1–3 km in from the shelf edge (Fig. 3.31). Sediment is entirely skeletal in origin, being derived from the outer reef tract, and from molluscs and algae particularly, but also from echinoids and foraminifera, which live in the bare sand habitat. Grains are commonly micritized but there are no ooids forming on the Florida Shelf.

Lime sand also occurs in tidal channels and associ-

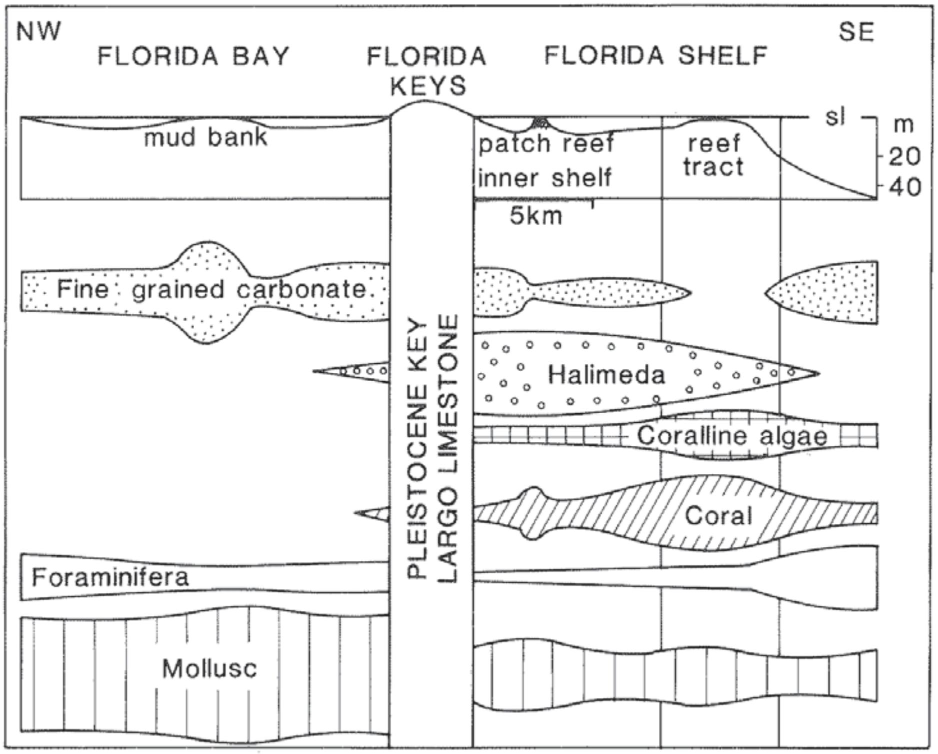


Fig. 3.29 *Distribution of sediment grain size and type across the Florida Shelf and Florida Bay. After Ginsburg (1956).*

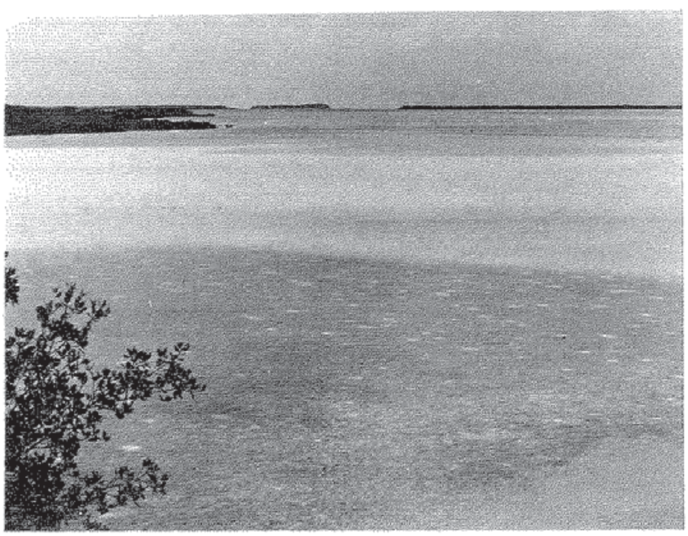


Fig. 3.30 *Shallow, protected lagoonal area of the southern Florida Keys showing* Callianassa *mounds in the foreground, shallows and channels in the middle part, and low islands and barriers colonized by mangroves in the distance.*

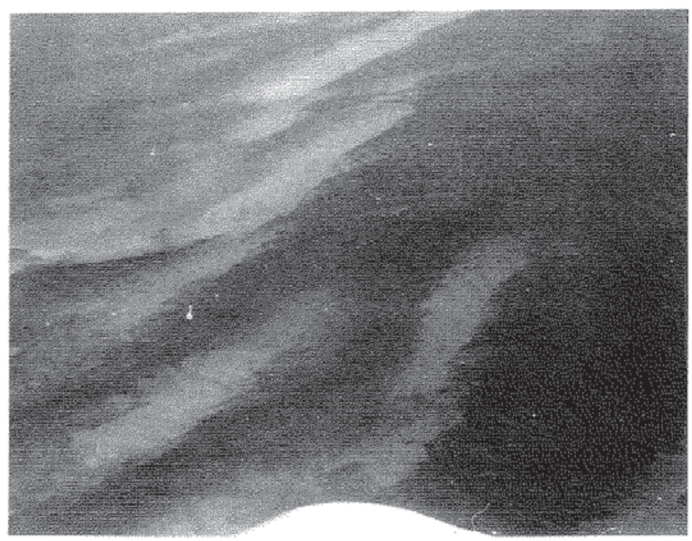


Fig. 3.31 *Aerial view of White Bank, ridges of skeletal sand near the shelf margin off Key Largo, Florida. Areas between the sand waves are covered in sea-grass. Photo courtesy of Ian Goldsmith.*

ated tidal deltas developed between islands of the Florida Keys, where there is substantial drainage from the bay and there are strong tidal currents (e.g. the sand complex between Lower and Upper Matecumbe Keys, Ebanks & Bubb, 1975; and in Bluefish Channel in the Lower Keys, Jindrich, 1969). Beaches of lime sand are not well developed along the inner shelf margin (on the east side of the Florida Keys) since it is mostly a low-energy shoreline. This is a function of shelf dynamics (Section 2.5), where maximum wave action is concentrated at the shelf margin and wave energy is progressively damped across the shelf. Along most of the shoreline, there occur mangrove swamps against and upon the Pleistocene limestone which forms the Florida Keys. Sandy beaches occur close to major tidal channels cutting through the Keys. Examples include Bahia Honda in the Lower Keys and the southern part of Long Key (Multer, 1977). Even so, these beaches are narrow, with a 1–3 m wide foreshore and 1–10 m wide backshore before vegetation, because of the low tidal range.

3.3.1d Patch reefs

Patch reefs occur upon the Florida Shelf, several kilometres back from the shelf margin. They are generally elongate and stand several metres above the surrounding seafloor, which is usually bare lime sand close to the reefs. The patch reefs, such as Hens and Chickens, Matecumbe Coral Gardens and Mosquito Banks, are similar to those described in Section 3.2.1a from the Bahamas. *Porites*, *Montastrea*, *Siderastrea* and *Diploria* are the dominant corals, along with the hydrocoral *Millepora*. Gorgonians and *Halimeda opuntia* are abundant, and molluscs and echinoderms are common.

3.3.1e Outer reef tract

The outer reef tract is a belt up to 1 km wide extending along the shelf margin from off Key Biscayne for 200 km south-west to the Dry Tortugas area. Flourishing reef growth is only taking place where islands occur along the inner shelf margin, for example along the shelf margin seaward of Key Largo, where Molasses Reef, Grecian Rocks, Key Largo Dry Rocks and Carysfort Reef occur. Where there are no Keys and Florida Bay water drains through on to the shelf, then reefs are impoverished or dead along the shelf margin (e.g. Alligator Reef, see Fig. 3.26). Some of the Florida outer reefs, such as Looe Key and Molasses, show a well-developed spur and groove topography (Section 3.2.1a), and this is ascribed to a constructional process, being related to zones of coral growth and strength of prevailing wave energy (Shinn *et al.*, 1981).

Subenvironments of the outer reefs are similar to those described from the Bahamas (Section 3.2.1a). As an example, Shinn (1980) has described five ecologic zones across the reef at Grecian Rocks, which occurs 1 km back from the shelf margin (Fig. 3.32): (1) a deep seaward coral rubble zone at depths of 6–8 m where there is still some strong coral growth and many soft corals (alcyonarians), (2) an area of spurs and grooves in 3–7 m of water where massive coral heads, especially *Montastrea* and the hydrocoral *Millepora* are common, (3) a zone of oriented *A.*

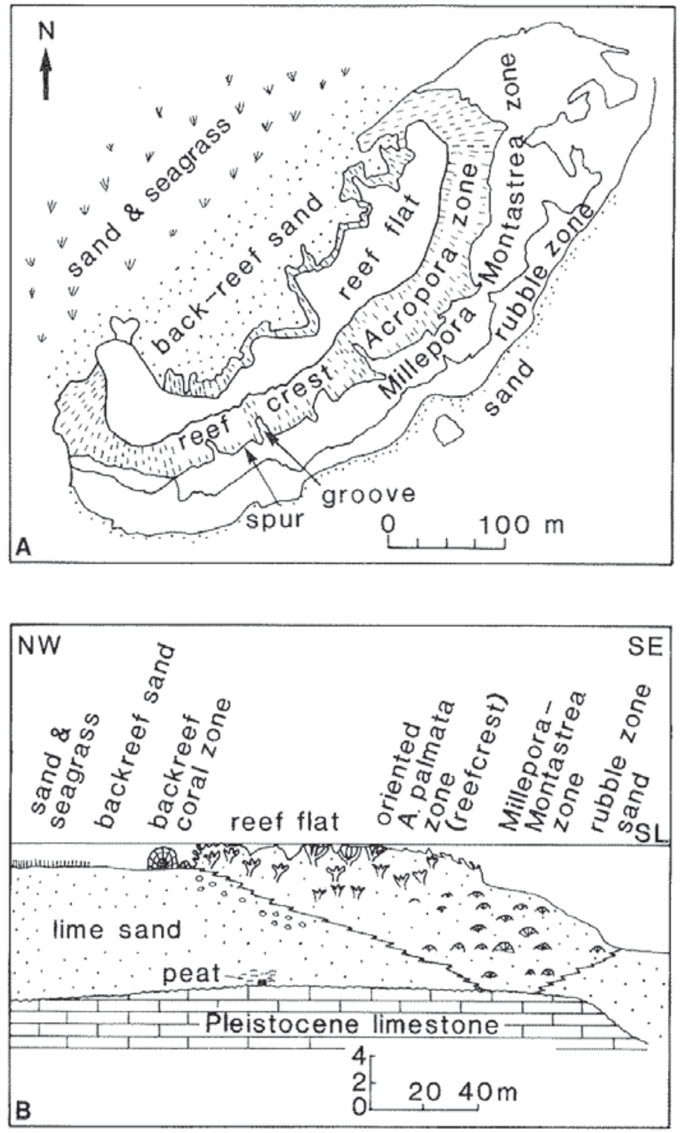


Fig. 3.32 *The reef at Grecian Rocks, near Dry Rocks, off Key Largo. (A) Zonation of the reef. (B) Cross-section of the reef based on shallow coring and radiocarbon dating. After Shinn (1980).*

palmata, in water depths of 0.5–4 m, which receives the brunt of waves and oceanic swells. Branches of this coral are mostly oriented landward, away from incoming waves, (4) the reef flat, composed of unoriented *A. palmata* which has grown to spring low tide level. Much of this coral is dead and there is rubble around, but active growth does occur on the leeward side of the reef flat, by *A. palmata* and many other corals, (5) the back-reef zone, consisting of scattered colonies of *A. palmata*, many thickets of *A. cervicornis*, large heads of *Montastrea annularis* and *Diploria*, with rubble and lime sand between. Some debris in this back-reef area is transported from the fore reef during hurricanes (Ball *et al.*, 1967). Reef debris extends back on to the shelf for up to 1 km in places as back-reef talus lobes (Enos, 1977a) and rubble islands are formed locally (e.g. the Sambo reefs and rubble islands 7 km south of Boca Chica Key, Fig. 3.26). Seaward transport of reef debris occurs down chutes between coral buttresses. Coring through Grecian Rocks Reef (Shinn, 1980) has shown that the reef is around 5 m thick and founded on 5 m of lime sand which rests on a flat surface of Pleistocene limestone (Fig. 3.32B). Reef development at Grecian Rocks appears to have involved the seaward growth of the oriented *A. palmata* (reef crest) zone over the *Montastrea–Millepora* zone, and it also appears that the reef has been extending leewards, mainly by corals colonizing the storm-derived rubble (Fig. 3.32B). A similar trend has been documented from other Florida shelf-margin reefs (Shinn *et al.*, 1977) where it can be shown that underlying Pleistocene bedrock topography controlled the location and trend of Holocene reef development. The Florida Shelf was flooded by the post-glacial transgression 6000–7000 years BP and coral reef growth followed soon after. However, it appears that more tolerant massive corals such as *Montastrea* and *Porites* were the dominant forms until 3000–4000 years BP, and then the more sensitive *Acropora* genera were able to grow when more open oceanic conditions were established.

3.3.1f Fore-reef muddy sand belt

An extensive fore-reef muddy sand belt occurs seaward of the outer reefs on the gentle fore-reef slope at depths greater than around 20 m. Sediment is derived from molluscs, green algae (*Halimeda* can exist to depths in excess of 50 m), echinoids, foraminifera and corals. Coarse reef debris and rippled sand, with patches of live coral, occur in the fore-reef zone between this muddy sand blanket and the outer reef itself.

3.3.1g Mud banks

One important feature of the inner Florida Shelf is the presence of mud banks or *mounds* (Fig. 3.33) and the associated *shoal fringe habitat*. The best documented are Rodriguez Bank (Turmel & Swanson, 1976) and Tavernier (Bosence *et al.*, 1985), location shown on Fig. 3.26. The banks are mostly composed of sandy mud but there is a distinct zonation of lithofacies and habitats around the banks (Fig. 3.34). The banks are emergent and covered by red and black mangroves, which baffle and trap sediment during storm flooding, and give a peaty soil. Around the banks in the shallow subtidal there is a grass and green algal zone where much lime mud is produced by *Penicillus* and lime sand by *Halimeda*. Sea-grass stabilizes the sediment and prevents erosion and acts as a baffle to trap suspended sediment. On the windward margin there next occurs a *Neogoniolithon* zone where this red branching alga grows profusely. Seawards is the *Porites* zone and the branching finger coral *Porites divaricata* forms a dense intergrowth of colonies (Fig. 3.33B), with many other animals, grass and algae associated, which is a wave-resistant hedge. The red algae and finger corals are brittle and quite easily broken to form coarse sand and gravel. In the deeper water around the bank, grass-covered muddy sand is present.

The mounds have developed through sediment trapping by mangroves, rooted green algae, sea-grass, red algae and branching corals to form a buildup which rises from water depths of around 3–5 m to just above HWM.

In the subsurface of Tavernier mound, a molluscan-rich gravelly mud facies dominates (Fig. 3.34B) with minor foraminifera, ostracods, *Halimeda* and sponge debris, and some 40% cryptocrystalline grains (Bosence *et al.*, 1985). Vertical *Thalassia* roots are common in this facies which is thought to represent the development of the mound by sea-grass trapping of sediment produced by the local molluscan, green algal community and epibionts upon the grass blades. In fact, the mound originated in a valley within the Pleistocene surface as sea-level was rising from 8000 years ago, and once the whole shelf was flooded, this area continued to be a grass-dominated site of high organic productivity. Aragonite and Sr contents and SEM study (Fig. 3.33C) show that the mud in the mound is derived from mixing *Penicillus*, *Halimeda* and *Porites* on the one hand with *Thalassia* epibionts and *Neogoniolithon* on the other, with the subsurface molluscan mud facies having a higher green algal input than the surface muds (Fig. 3.34C). Breakdown of molluscs contributes little to the mud fraction. More open marine conditions in the last few thousand years have resulted in growth of the windward *Neogoniolithon* and *Porites* zones and the seaward (eastwards) growth of the mound as a whole.

Mud banks also occur in *Florida Bay*. This is a large, rock-floored lagoon occurring behind the Florida Keys and south of the Everglades, connected to the open Atlantic Ocean through tidal channels between the Keys. Florida Bay consists of a network

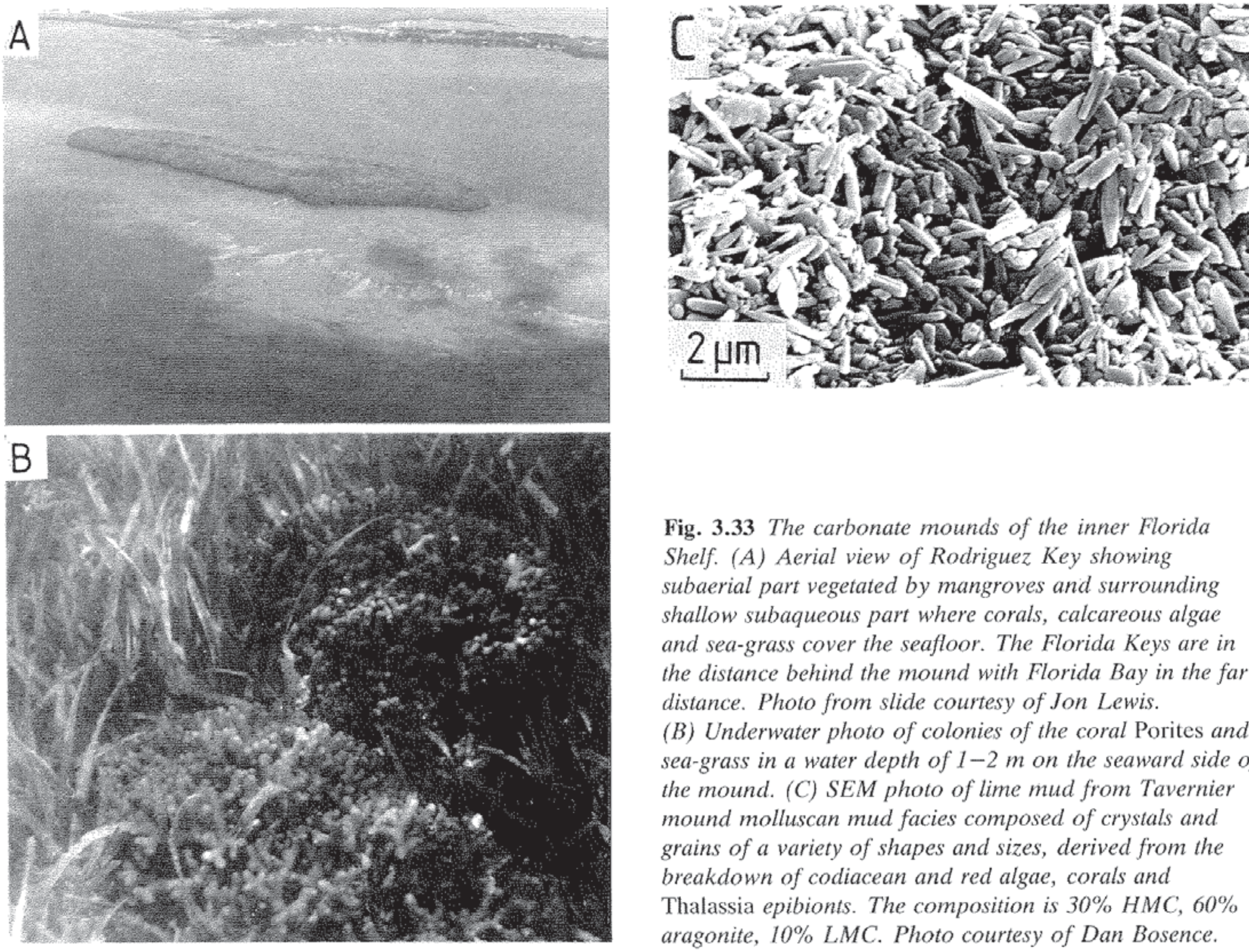
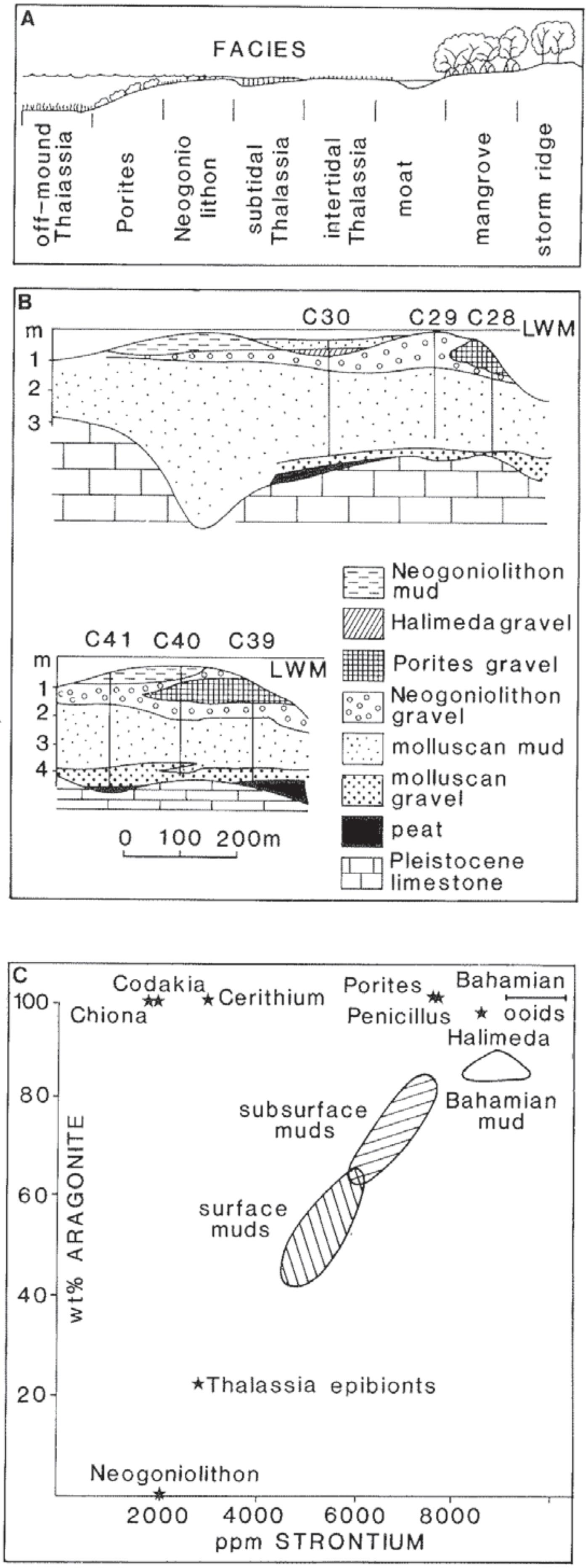


Fig. 3.33 *The carbonate mounds of the inner Florida Shelf. (A) Aerial view of Rodriguez Key showing subaerial part vegetated by mangroves and surrounding shallow subaqueous part where corals, calcareous algae and sea-grass cover the seafloor. The Florida Keys are in the distance behind the mound with Florida Bay in the far distance. Photo from slide courtesy of Jon Lewis. (B) Underwater photo of colonies of the coral* Porites *and sea-grass in a water depth of 1–2 m on the seaward side of the mound. (C) SEM photo of lime mud from Tavernier mound molluscan mud facies composed of crystals and grains of a variety of shapes and sizes, derived from the breakdown of codiacean and red algae, corals and* Thalassia *epibionts. The composition is 30% HMC, 60% aragonite, 10% LMC. Photo courtesy of Dan Bosence.*

of mud banks and mangrove keys which are interconnected to divide the bay into many shallow basins, commonly referred to as 'lakes' (Enos & Perkins, 1979). The area of Florida Bay was flooded about 4000 years ago and since then sediments have been produced largely by the breakdown of carbonate skeletons, especially the algae *Penicillus* and *Halimeda*, molluscs which dominate the bottom fauna, and foraminifera. Close to the channels connecting with the shelf, a more diverse fauna occurs, including some corals. Much of the muddy bay floor is covered with turtle grass and this has helped construct the banks (Enos & Perkins, 1979). Although vertical accretion has generally been assumed for the Florida Bay mud banks, recent coring has shown that in some banks there are important differences in the sediments of leeward and windward sides. Also, a crude, large-scale cross-stratification suggests that the banks in effect may be bed forms, which migrate during major storms (Bosence, 1989).

3.3.2 Intertidal and supratidal carbonates of the inner Florida Shelf

Intertidal–supratidal flats are poorly represented along the inner Florida Shelf; much of the shoreline is occupied by mangroves, sandy beaches are rare (see Section 3.3.1c), and there are a few rocky shores (rock habitat, features noted earlier, Section 3.3.1a). Mangroves (Fig. 3.35) contribute greatly to the development of new land by their baffling effect on sediment in transit. The black mangrove *Avicennia nitida* occurs in the highest intertidal and supratidal zones and is distinguished by its aerial roots. The red mangrove *Rhizophora mangle* with prop roots occurs more in the upper intertidal. The latter produces more peat than the former. The thick tangle of roots traps sediment carried into the swamp during high tides and storms. Many organisms are associated with the mangrove swamps: molluscs and barnacles on the stems and branches, and crustaceans between the



roots and in burrows. Mangrove swamps are well developed on the Florida Bay side of the Florida Keys and they cap some mud banks in Florida Bay. The main development, however, is along the southern and western coasts of mainland south Florida in the Everglades National Park. These swamps are considered analogues for ancient coal-forming environments (e.g. Cohen & Spackman, 1980).

Tidal flats are not as well developed in the Florida Keys area as on the Bahamas. They occur mostly in the Lower Keys, such as Sugarloaf and Big Pine Keys, and then generally on the bay side, where there is most protection from the waves and storms which come dominantly from the east. They also occur on some of the larger islands in Florida Bay, such as Crane Key and Cluett Key. The tidal flats are quite small, being only a few square kilometres in area, and they do not have tidal channels. Depressions permanently occupied with water are common. Clumps of small mangroves are widely distributed. Microbial (algal) mats are the main surface feature of these tidal flats and they vary from extensive carpets where the tidal flat is permanently wet, to desiccated and buckled mats where they are exposed for long periods of time. These mats give rise to stromatolitically-laminated pelleted mud with laminoid fenestrae. Birdseyes (irregular fenestrae) are common in unlaminated pelleted lime muds (Shinn, 1968a). Storms, especially hurricanes, transport much sediment on to the flats to give a distinctive layering. A 50 mm bed was deposited on Crane Key supratidal flat during Hurricane Donna in 1960 (Ball *et al.*, 1967; Perkins & Enos, 1968).

The supratidal flat sediments of Sugarloaf contain up to 80% dolomite, and this mostly occurs within surficial crusts which are cemented storm layers (Shinn, 1968b). Dolomitic intraclasts are common. The origin of this dolomite has been put down to evaporation of seawater and an increase in Mg/Ca ratio, perhaps induced by the precipitation of aragonite. Recent work by Carballo *et al.* (1987) has suggested that the dolomite is formed by tidal pumping of Florida Bay water through the thin Holocene sediment package during spring tides (see Section 8.7.5).

Fig. 3.34 *The carbonate mound at Tavernier. (A) Subenvironments across the mound. (B) Two cross-sections showing distribution of sediment types. (C) Strontium–aragonite contents of mound muds and of sediment-producing organisms. Also shown are Bahamian ooids and muds. After Bosence* et al. *(1985).*

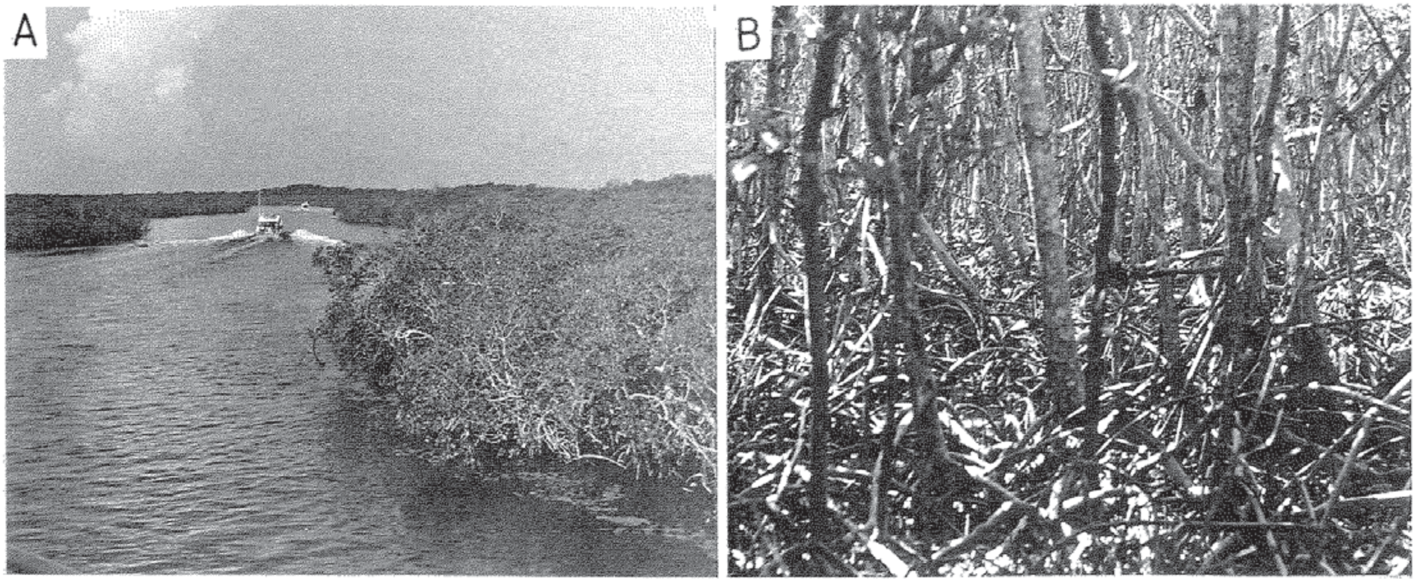


Fig. 3.35 *Mangrove swamps along the Florida Shelf shoreline. (A) Tidal channels through the swamp. (B) Dense network of subaerial roots and branches of the mangrove.*

3.4 CARBONATE SEDIMENTS OF THE TRUCIAL COAST

The Trucial Coast of the Arabian Gulf is an area of extensive subtidal and intertidal carbonate sedimentation, and supratidal carbonate, dolomite and evaporite precipitation (papers in Purser, 1973a, and reviews in Bathurst, 1975 and Schreiber *et al.*, 1986). It is a modern example of a carbonate ramp, where the seafloor gradually slopes from sea-level to many tens of metres, without any major break of slope (Fig. 3.36; see Section 2.6). This ramp, in fact, is not a smooth surface; there are many local positive areas, shoals and islands, which are structurally controlled, some being due to movement of salt (halokinesis). The Trucial Coast is a mesotidal area with a tidal range of 2.1 m along the shoreline, dropping to 1.2 m within lagoons. The northeast–southwest oriented coast directly faces strong winds (Shamals) coming from the north–northwest. Because of a very arid climate, and the partly enclosed nature of the gulf (Fig. 3.36), salinity (40–45‰) is a little higher than in the Indian Ocean (35–37‰), and in the lagoons it may reach 70‰.

The offshore outer ramp, in deep water below wave-base where fine sediments accumulate, gives way to a complex of nearshore inner-ramp sedimentary environments of sand shoals, beach-barrier islands and coral reefs (Fig. 3.37). Behind the beach-barrier island system of the Trucial Coast there occur lagoons which are connected to the open gulf via tidal channels through the barriers. Ebb and flood tidal deltas occur at the gulf and lagoonward ends of the channels. Extensive intertidal flats dissected by tidal creeks occur on the landward side of the lagoons and these are partly covered by microbial mats. Still further landwards there occur the broad supratidal flats or sabkhas, wherein gypsum–anhydrite is precipitating. Dolomite is being precipitated in high intertidal–supratidal zones. The back-barrier environments are particularly well developed along the Trucial Coast because this area is a tectonic depression.

3.4.1 Shoals, barriers and reefs of the inner ramp

In the Arabian Gulf, skeletal sandy muds of the deeper, outer ramp give way to carbonate sands and reefs in the shallow subtidal to intertidal zones of the inner ramp along the Trucial Coast (Wagner & Van der Togt, 1973). The outer-ramp skeletal sandy mud lithofacies consists largely of bivalve and foraminiferal debris, often unbroken and unabraded, with mud which has come largely from the Tigris and Euphrates

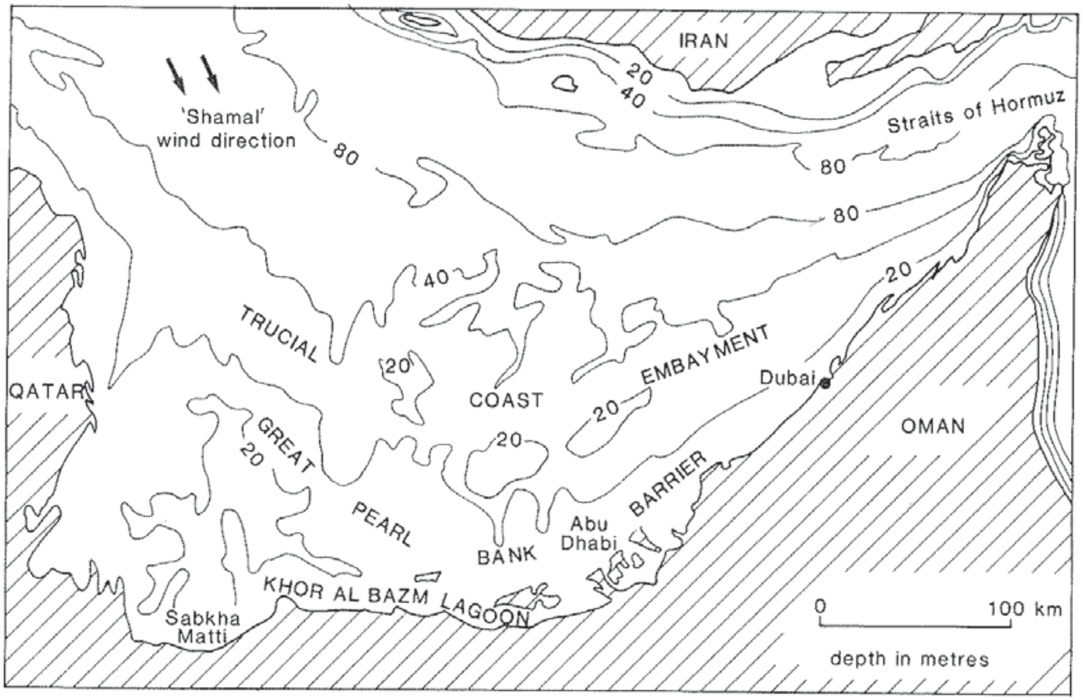


Fig. 3.36 *The Trucial Coast Embayment of the Arabian Gulf. After Purser (1973a).*

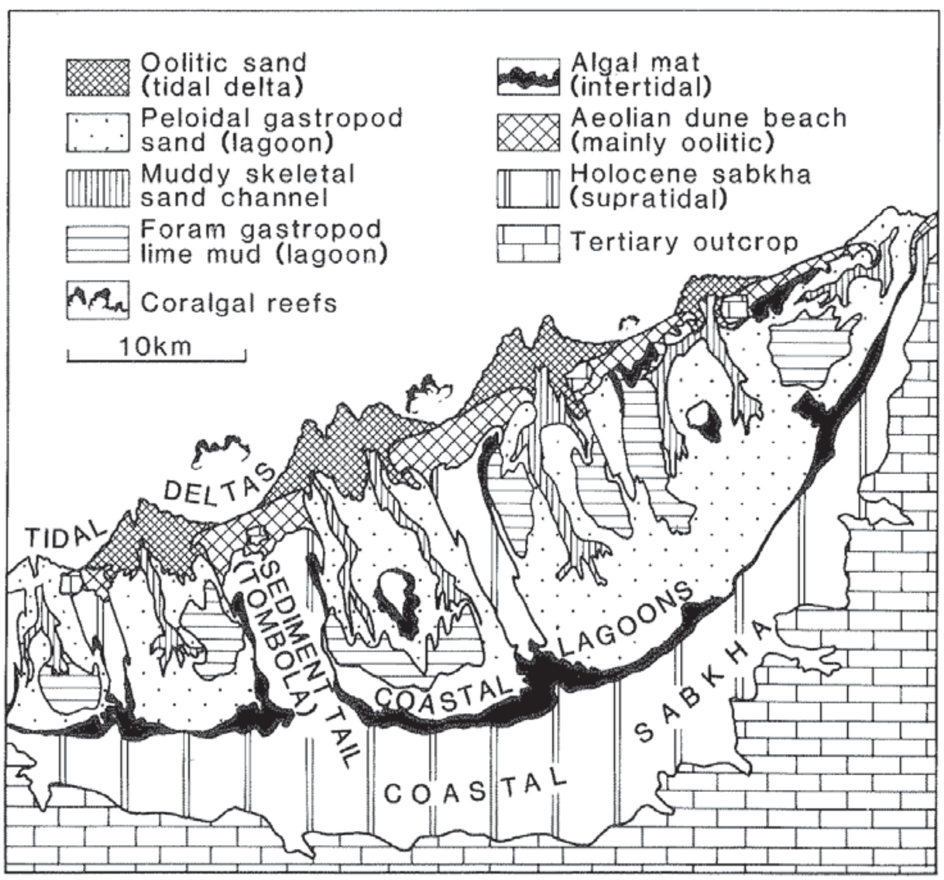


Fig. 3.37 *Schematic map of Abu Dhabi region of the central part of the Trucial Coast showing depositional environments and sediments. After Purser (1973a).*

Rivers at the head of the Arabian Gulf. The inner-ramp sands form an east–west oriented barrier, whose location is structurally determined, with a major lagoon, the Khor al Bazm, located behind (Fig. 3.38; Purser & Evans, 1973). In the west, the sand shoal is submerged and forms the Great Pearl Bank; towards the east it becomes emergent to form a beach-barrier island system, capped by aeolian dunes and dissected by tidal channels. Outcrops of Pleistocene limestone along the structural high are foundations to some of

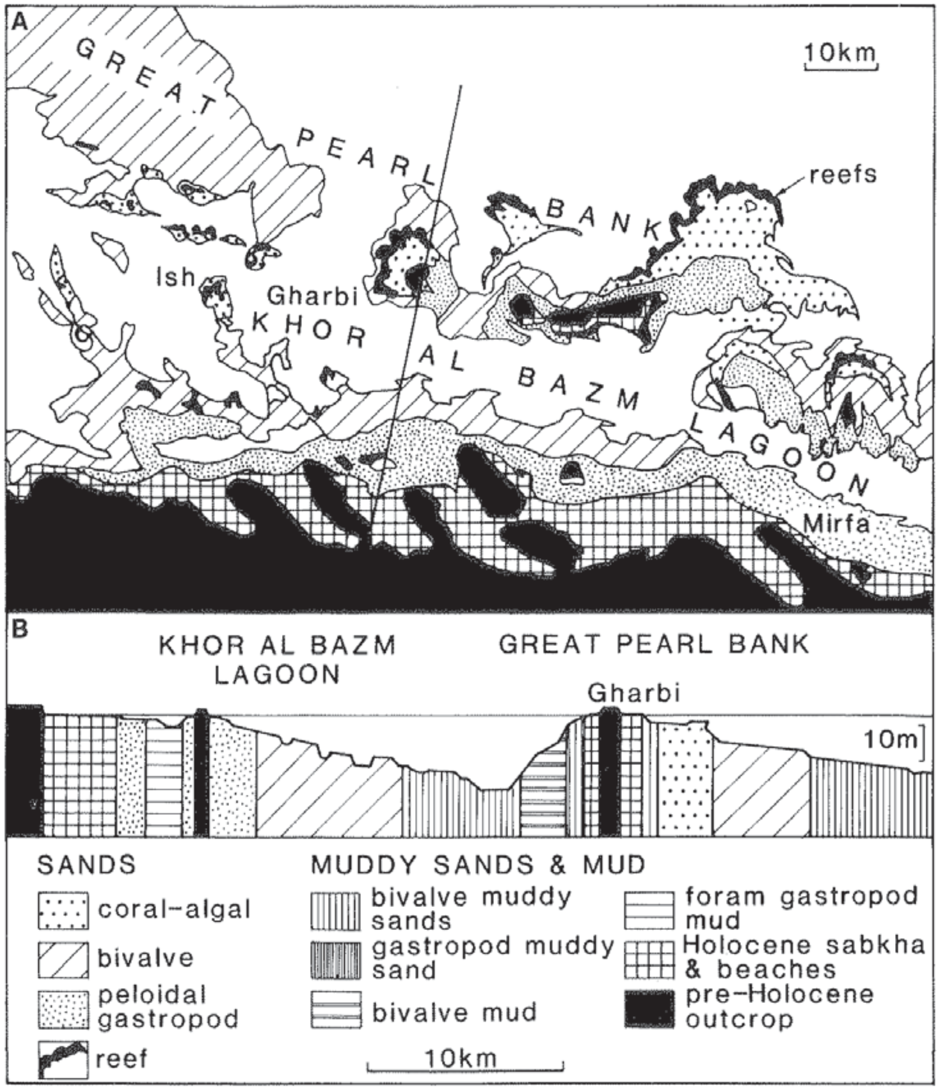


Fig. 3.38 *Sediment distribution and schematic cross-section of the western part of the Trucial Coast, in the region of the Great Pearl Bank and Khor al Bazm Lagoon. After Purser (1973a).*

the islands. The asymmetric Great Pearl Bank is steeper on the lagoon side as a result of cross-bank transport during storms.

Coral patch reefs are developed on the lagoonward side of the bank on shoals and around islands rising up from the deeper parts of the ramp in the central part of the gulf (Purser, 1973b). Prolific coral growth is inhibited by the slight hypersalinity of gulf waters and occasional low (<20°C) water temperatures. Three coral types are common: the branching *Acropora*, the massive *Porites* and the brain coral *Platygyra*. Encrusting red melobesoid calcareous algae are important reef binders, occurring especially on dead corals. Sponges, serpulids and oysters are also common. Various gastropods, regular echinoids and fish graze on the reefs and along with wave action they generate much sand- and mud-sized skeletal debris.

Towards the east, the Great Pearl Bank widens to 20 km and consists of many sand banks and islands. Once the latter reach several kilometres in size, they develop tidal flats on their lagoonward sides. The barrier is traversed by tidal channels, some of which have tidal deltas building into the lagoon. Further east around Abu Dhabi, the coastline becomes a complex of barrier islands, peninsulas and small lagoons (Figs 3.37 and 3.39). Island growth around Pleistocene outcrops has taken place by the lagoonward growth of tails of carbonate sand and lateral accretion of sand spits by longshore currents. Channels between islands have spectacular oolite tidal deltas at their seaward ends, with minor bars and deltas developed within the channels and at their lagoonward ends (Fig. 3.39). The channels are the sites of ooid precipitation and the high rates of production are the main reasons for

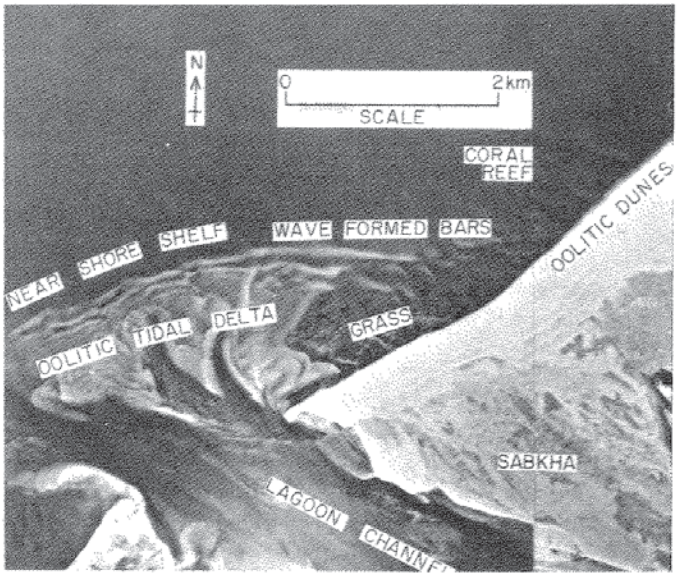


Fig. 3.39 *Tidal channel with bars and shoals, ebb-tidal delta with sand waves and spillover lobes and beach-dune ridges with sabkha behind. In protected areas of the open marine environment, sea-grass is developed and in high-energy areas, coral reefs are present. After Schreiber* et al. *(1986).*

the rapid barrier growth (Loreau & Purser, 1973). Fringing coral reefs just offshore from the barrier islands also supply some sand. Sea-grass occurs in some quieter-water areas seaward of the barriers.

In many shallow-water areas of the inner ramp, carbonate sediments are being cemented on and just below the seafloor to form surficial crusts and hardgrounds. Off Qatar especially, seafloor cementation by acicular aragonite and high-Mg calcite has produced pavements with polygonal crack patterns and pseudo-anticlinal structures (*tepees*) where the crust has fractured and expanded through cement precipitation (Shinn, 1969). Cemented sands form hard substrates (hardgrounds) for encrusting organisms and they are frequently penetrated by borings.

From the foregoing, three principal lithofacies can be distinguished for the Trucial Coast inner-ramp environments:

1 Skeletal and oolitic grainstones of the relatively high-energy shoals, beach barriers, tidal deltas and tidal channels. Skeletal material is generally well abraded and is derived from molluscs, echinoids, foraminifera and corals. Sedimentary structures are common, mainly various types of cross-stratification and some burrows, and these together with data on lithofacies geometry and sequence, palaeocurrents and subtleties of the sediment composition, would enable the various environments to be distinguished in a fossil situation.

2 More muddy skeletal sands of the quieter-water locally grass-covered areas seaward of the barrier.

3 Coral–algal bindstones and framestones of the fringing and patch reefs.

3.4.2 Lagoons

The major Khor al Bazm Lagoon behind the Great Pearl Bank decreases in width and depth towards the east and salinity increases from 40‰ to 50‰. Along with this increasing restriction, corals, echinoids and algae decrease in importance and gastropods and foraminifera dominate. The sediments are mostly skeletal–pelletal sands, with little lime mud. The small protected lagoons at the eastern end of the Trucial Coast (e.g. Fig. 3.40) have salinities up to 60‰ and sediments are pelleted lime muds rich in imperforate foraminifera and gastropods. Locally, swamps occur around the edges of the lagoons, colonized by the black mangrove *Avicennia marina* and other halophytic plants. Here lime muds and pelleted muds accumulate and they are extensively bioturbated by crustaceans.

The lime mud and pellets of the lagoons are composed of aragonite needles and chemical evidence suggests that these are a direct precipitate (Kinsman & Holland, 1969). The needles are 1–4 μm long and have a strontium content of around 9400 ppm. Calcareous green algae, the source of much lime mud in the Caribbean (Section 3.2.1d), are rare in the gulf and the main aragonite-producing organisms, the molluscs, have 1000–4000 ppm Sr. Corals have high Sr but they do not contribute much sediment to the lagoon floor. Furthermore, around 9400 ppm Sr is close to the expected value for aragonite precipitated from seawater at the known temperature, and inorganically-precipitated, aragonitic ooids in the gulf have a similar Sr content (9600 ppm). The occurrence of 'whitings' (Section 3.2.1d) in the open gulf and lagoons could be the inorganic precipitation taking place, although Ellis & Milliman (1986) documented evidence against this. Milkiness in gulf waters is also produced by shoals of fish stirring up the bottom sediment and discharge from oil tankers.

The dominant lithofacies of the lagoons are thus skeletal-pelletal sands in areas of moderate circulation and lime mud and pelletal mud in the most protected parts. Faunal diversity is low compared with seaward lithofacies but certain species of gastropods and

Fig. 3.40 *View over lagoon and tidal flats behind the Trucial Coast barrier showing well-developed polygonal mud cracks in the lime mud. Photo from slide courtesy of John Powell.*

foraminifera are abundant, and burrow structures produced by annelids and crustaceans are common, giving a mottled appearance to the sediments. Mangrove swamps give rise to a lime mud and pelleted mud lithofacies with characteristic root structures and possibly some peat.

3.4.3 Tidal flats

The intertidal zone of the Trucial Coast is up to 5 km across and is dominated by microbial mats in the upper part. Tidal creeks dissect the flats and patches of halophytic plants occur at their lagoonward ends. Although the flats themselves are mostly composed of lime mud and pelleted mud, low beach ridges, strand-line deposits and some tidal creeks consist of skeletal–pelletal sand, dominated by cerithid gastropods. As in the Bahamas, the tidal flat sediments are laminated as a result of storm and spring tide deposition, and laminoid, irregular and burrow fenestrae are common. In some parts of the tidal flat, active precipitation of aragonite is leading to the formation of cemented crusts. In a similar way to the subtidal hardgrounds, polygonal cracks and tepee structures are generated (Evamy, 1973). Fracture and breakage of the crusts gives rise to intraclasts which can be reworked into edge-wise conglomerates or flakestones.

A distinctive feature of the Trucial Coast tidal flats is the development of a microbial (algal) mat belt, up to 2 km wide, along protected lagoon margins where surface slopes are very low. In the low intertidal and lagoonal areas, mat growth is mostly precluded by the grazing activities of cerithid gastropods. However, the latter cannot tolerate high salinities, so where these occur, small domal stromatolites up to 60 mm high and 180 mm across form, in waters up to 3 m deep.

The Trucial Coast algal–bacterial mats are dominated by two basic algal communities, those of *Microcoleus* and *Schizothrix* (Park, 1976). The microbial mat types, however, are determined by environmental factors, rather than the structure of the algal–bacterial community. Four principal types of mat are recognized and the major control on their distribution is the frequency of flooding by tidal and storm-driven lagoon waters (Fig. 3.41). The most important type is the *smooth mat* of Kinsman & Park (1976), also called polygonal mat (Kendall & Skipwith, 1968) and flat mat (Park, 1977). It covers 30–60% of the mat belt and is best developed in shallow pool and channel sites (Fig. 3.42). The dominant alga is the filamentous *Microcoleus chthonoplastes*. Smooth mat gives rise to stromatolites consisting of alternations of organic and sediment laminae (Fig. 3.42B). The thicknesses of the carbonate laminae are generally a millimetre or two, but they increase to 10 mm in the direction of the lagoon. In areas of lower flooding frequency, in the upper parts of the tidal flat, the smooth mat is more of an algal peat, with little sediment, and the mat is broken into polygonal structures through desiccation (Fig. 3.42A). Smooth mat has a high preservation potential. *Pustular mat* of Kinsman & Park (1976), equivalent to cinder mat of Kendall & Skipwith (1968), is the initial mat type in low to mid intertidal areas and is dominated by the coccoid alga *Entophysalis major*. The fabric produced by sediment trapping on a pustular mat has a clotted or thrombolitic appearance. *Pinnacle mat* occurs in the higher, usually well-drained parts of the tidal flat and has small tufts, formed by the large filamentous *Lyngbya aestuarii*. *Blister mat* (crinkle zone of Kendall & Skipwith, 1968) occurring in the highest intertidal zone, consists of a leathery, domed mat with little trapped sediment, but it has a low preservation potential.

Although a little sediment is deposited on the microbial mats during every diurnal tidal flooding, the thicker and thus more significant laminae are invariably the product of major storms (Park, 1976). Growth rates of living mats appear to be of the order of 2–2.5 mm yr^{-1}, but buried mat horizons suggest that after desiccation and compaction, the accretion rate is of the order of 0.2 mm yr^{-1}.

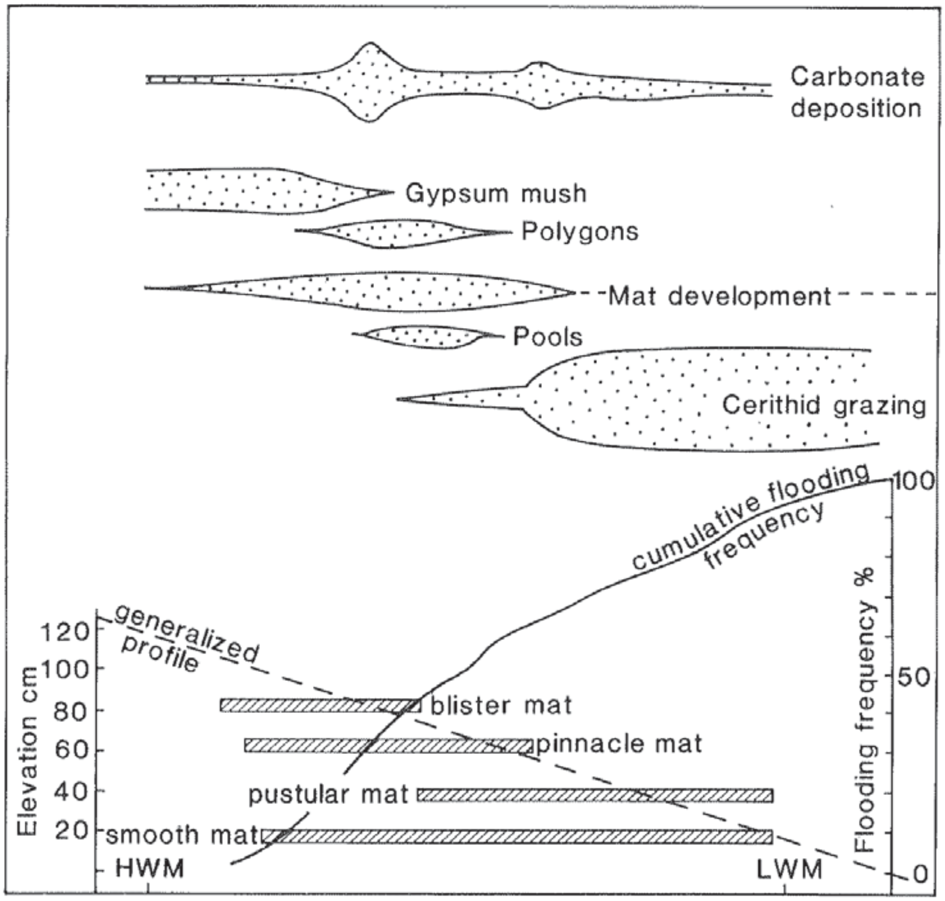


Fig. 3.41 *Distribution of algal mat types and sedimentary features across the tidal flats of the Trucial Coast. After Park (1977).*

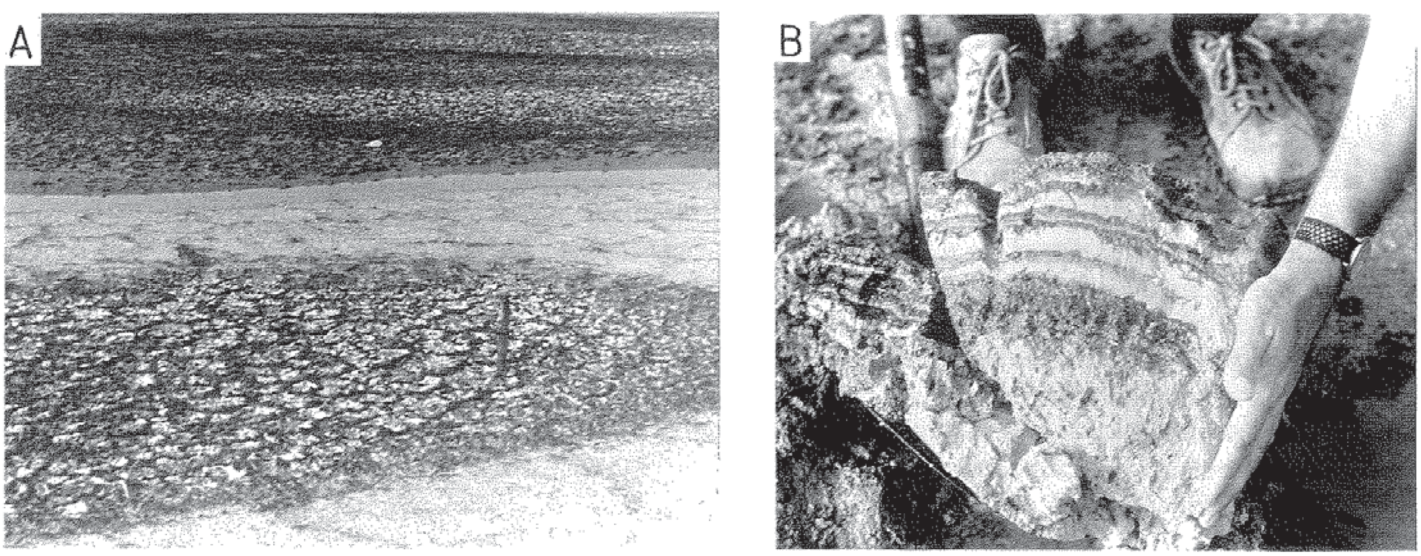


Fig. 3.42 *Algal mats of the Abu Dhabi tidal flats. (A) View showing desiccated polygonal mats, and larger polygonal structures close to tidal channel. Knife 0.2 m long. (B) Section through tidal flat sediment showing algal mat layers alternating with sediment layers. Photos from slides courtesy of John Powell.*

In the higher parts of the tidal flat, high evaporation results in the precipitation of gypsum crystals within and beneath the microbial mats. Seaward progradation of the supratidal sabkha surface progressively buries the mats so that an organic horizon occurs beneath the sabkha (Fig. 3.43). Within this buried algal bed and in the sabkha sediments above, more gypsum and anhydrite are precipitated. The precipitation of these evaporite minerals has a disruptive effect upon the mat laminae, and may destroy the

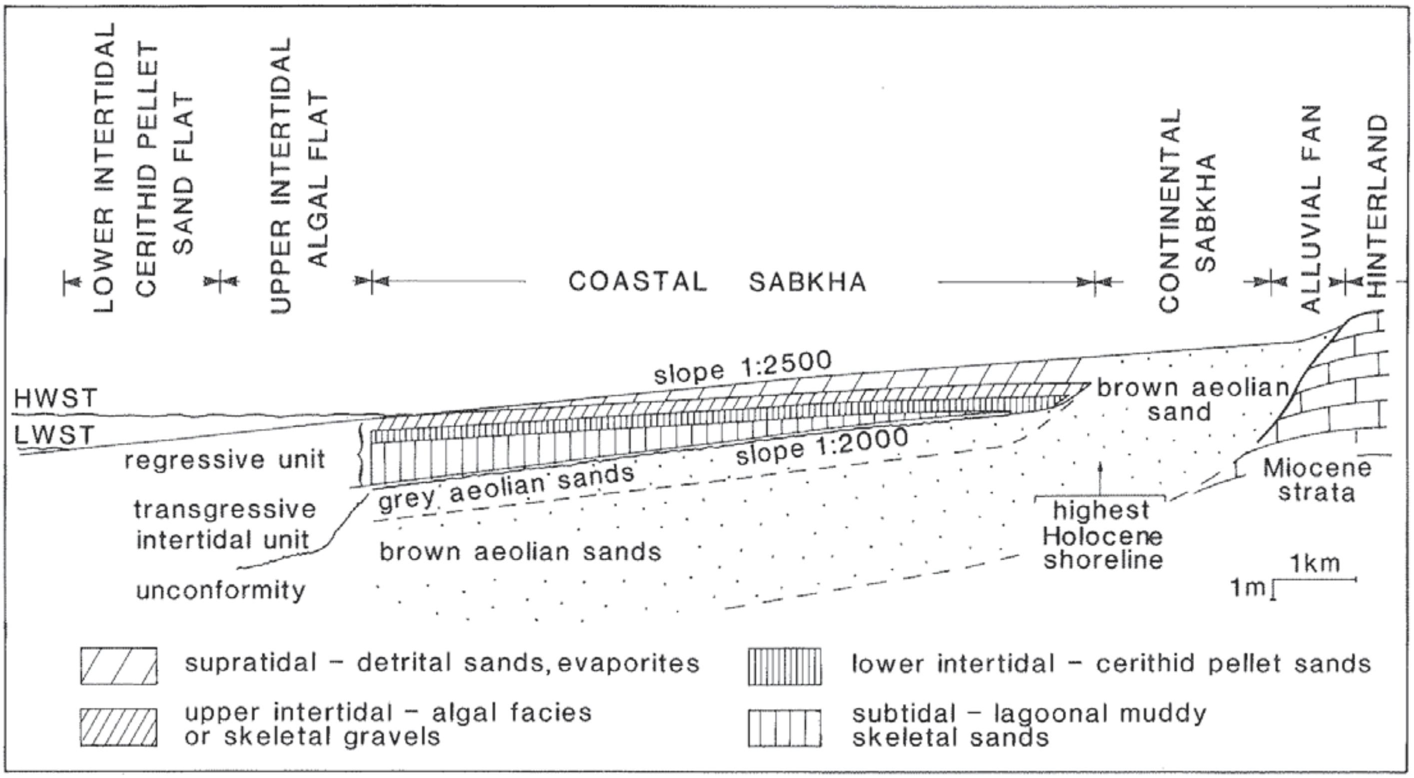


Fig. 3.43 *Cross-section of the Abu Dhabi tidal flat–sabkha showing sediment types. After Kinsman & Park (1976).*

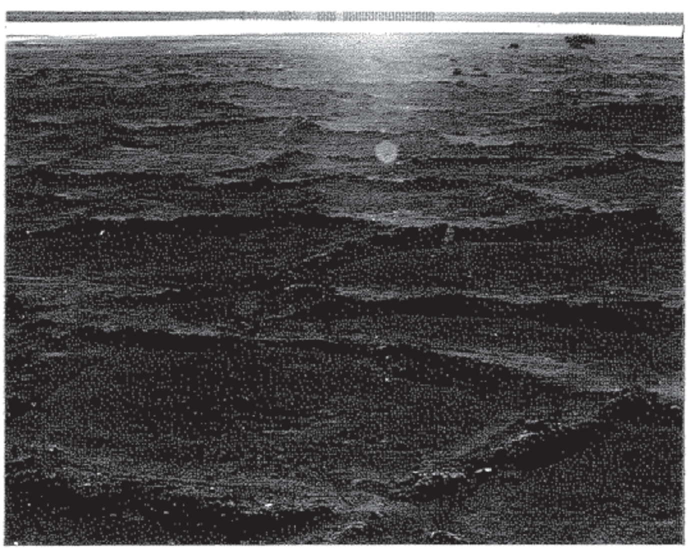


Fig. 3.44 *Large polygonal structures in supratidal sediments. Abu Dhabi sabkha. Photo from slide courtesy of John Powell.*

horizon altogether (Park, 1977).

The two characteristic tidal flat lithofacies of the Trucial Coast are pellet mud and lime mud with storm laminae and fenestrae, and stromatolitically-laminated lime mud with a variety of fabrics reflecting the microbial mat type. Minor lithofacies are skeletal, mostly gastropod, sands and gravels and skeletal–pelletal sands of tidal creeks and channels and former beach ridges.

3.4.4 Supratidal flats and sabkhas

Landwards from the intertidal zone of the Trucial Coast occurs an extensive supratidal area known as the sabkha. The sabkha extends for up to 10 km inland and has an imperceptible seaward slope of around 1:2500. The sabkha surface has formed in the last 5000 years as a result of the gradual seaward migration of the shoreline. Progradation was mostly brought about by the filling of lagoons and sedimentation in the supratidal–intertidal zones, but in addition to this sedimentary offlap, there has been a relative fall in sea-level of 1.2 m. The offlap has proceeded at an average rate of 2 m yr^{-1}, so that a broad supratidal zone (the sabkha) has been generated (Patterson & Kinsman, 1981).

Seawater flooding of the supratidal flat occurs during high tides and storms and transports much lime mud and sand on to the sabkha. Carbonate laminae are deposited but these are mostly disrupted by desiccation and evaporite mineral growth and large polygons develop on the sabkha surface (Fig. 3.44).

Flooding frequency decreases quickly across the extensive supratidal flat, so that seawater may only reach the most landward parts of the sabkha once every few years. As a result of the intense evaporation (1.5 m yr^{-1}), a suite of minerals is successively precipitated from the high intertidal zone across the sabkha and pore-fluid chemistry changes systematically (see Section 8.7.1 and Fig. 8.20). In the upper intertidal zone, aragonite is precipitated as a cement to generate surficial crusts and lithified subsurface layers. Dolomite is also being precipitated within the intertidal sediments (Illing *et al.*, 1965; McKenzie, 1981; Patterson & Kinsman, 1982), and it appears to be concentrated near remnant channels close to and above the present strandline. The dolomite itself is poorly ordered and Ca-rich, with an oxygen isotopic composition ($\delta^{18}O$ + 2‰) indicating an evaporative origin (see Section 8.6). As a result of the sabkha progradation over the intertidal sediments, a zone of dolomite is present beneath the sabkha surface, especially in the buried upper intertidal sediments. The absence of aragonite in these sediments suggests that the dolomite has formed by replacing this mineral. Analyses of porewaters have shown that dolomitization is taking place from fluids with a high Mg/Ca ratio (>6), pH of 6.3–6.9 and temperature of 25–40°C (Patterson & Kinsman, 1982).

Gypsum is also being precipitated in the upper intertidal sediments. The crystals are mostly less than 1 mm long, lens-shaped and flattened normal to the c-axis. A little further inland, in the outer supratidal zone where flooding occurs at intervals of a month or more, gypsum crystals up to several centimetres long form a surface mush which reaches 0.3 m in thickness. The gypsum is precipitated displacively within the sediment, but frequently includes carbonate grains and even microbial lamination. Replacement of aragonitic shells by gypsum also occurs.

In the mid sabkha, where flooding is more frequent than once a month, the gypsum mush gives way to anhydrite, which occurs as nodules of lath-shaped crystals, 1–100 μm in length (Butler, 1970). As a result of displacive growth, the original sediment occurs as stringers between the nodules to give a net or chicken-wire texture. Anhydrite also forms seams of coalesced nodules in the sabkha sediment above the former gypsum mush horizon, and these beds usually show contortions in the form of ptygmatic and disharmonic folds, referred to as enterolithic structures (Butler, 1970; Butler *et al.*, 1982). Lower in the sediment profile, in the buried algal and lagoonal sediments, lenticular gypsum crystals, up to 0.25 m long, are common.

Apart from dolomite, gypsum, anhydrite and halite, two other minerals, which are precipitated in minor quantities, are celestite and magnesite (Bush, 1973). Although some celestite ($SrCO_3$) is precipitated through the evaporative concentration of seawater, most is precipitated as a result of the release of Sr during the dolomitization of aragonite and during the secondary gypsification of anhydrite. Magnesite ($MgCO_3$) occurs within the sediments of the mid sabkha, above the buried microbial mats. In this area, brines have a very high Mg content; there is no aragonite in the sediments but some dolomite. The magnesite probably forms after dolomitization has removed all aragonite. Although much of the Ca^{2+} for the gypsum–anhydrite comes from seawater, another source is dolomitization. During this process, Ca^{2+} is released according to the equation:

$$2CaCO_3 + Mg^{2+} \rightarrow CaMg(CO_3)_2 + Ca^{2+}$$

and the Ca^{2+} combines with SO_4^{2-} in the porewaters to give more $CaSO_4$.

The typical sabkha lithofacies, as being formed in the Trucial Coast, is thus a nodular anhydrite and gypsum crystal mush in lime mud and muddy sand, with some dolomite. It is a distinctive facies of arid–semi-arid supratidal flats.

As a result of sedimentation upon the intertidal flats and sabkhas, and the early diagenetic precipitation of evaporites within the sediment, this coastal belt has prograded up to 10 km in the last 5000 years. The product of sabkha progradation is a definite sequence of sediments: the evaporitic supratidal sediments come to overlie the tidal flat sediments, and these in turn overlie lagoonal deposits. A pit dug in the outer to mid sabkha of the Trucial Coast reveals this sequence beneath the surface (Fig. 4.48), with the tidal flat part around 1.2 m thick, reflecting the tidal range, and the supratidal nodular and enterolithic anhydrite reaching 2 m. In the geological record, there are now many carbonate–evaporite formations interpreted as the product of sabkha sedimentation and precipitation, and most of these consist of repeated sabkha cycles (see Sections 2.10.2, 4.3.3g and 4.3.6).