Depositional Environments of Carbonate Platforms*

Paul M. Harris[1]

Search and Discovery Article #60032 (2009)
Posted May 15, 2009

*Reprinted from P.M. Harris, C.H. Moore, and J.L. Wilson, 1985, Carbonate Platforms, *in* J.E. Warme and K.W. Shanley, eds., Carbonate Depositional Environments, Modern and Ancient, Part 2: Carbonate Platforms: Colorado School of Mines Quarterly, v. 80, no. 4, p. 31-60 (also in P.M. Harris and C.H. Moore, 1985, Modern and Ancient Shelf Carbonates - Facies, Diagenesis, and Petroleum Potential: The Modern of South Florida and the Bahamas, The Ancient of the U. S. Gulf Coast: Colorado School of Mines Short Course Notes, 172 p.)
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[1]Chevron Oil Field Research Company, La Habra, California; currently 1ETC, Chevron, San Ramon, CA, USA. (MitchHarris@chevron.com)

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DEPOSITIONAL ENVIRONMENTS OF CARBONATE PLATFORMS

Paul M. Harris

INTRODUCTION

The following discussion of carbonate platform depositional environments relies heavily on well-studied modern areas from South Florida and the Bahamas (Multer 1977). These modern examples have proven to be valuable analogs for interpreting their ancient counterparts. The sections on reefs, sand shoals, and muddy lagoon and tidal flats do not provide all the answers or models that may be needed in a particular subsurface study; however, the depositional environments and resultant facies emphasized have proven to be exceedingly important. They comprise the majority of subsurface carbonates that contain substantial quantities of hydrocarbons.

Shallow-water carbonate sediments in South Florida have accumulated to substantial thicknesses only from the shelf edge to the Florida Keys (the Florida reef tract) and in Florida Bay (Ginsburg 1956; Enos and Perkins 1977) (figure 39). The carbonates are mixed with quartz sands to the north and are eventually replaced by them, because quartz sands are being transported southward along both the east and west Florida coasts.

In general the South Florida shallow shelf can be subdivided into an inner shelf (Florida Bay) and an open shelf (the reef tract), separated by the Florida Keys (figure 40). The landward boundary of the inner shelf is mangrove swamps and the supratidal flats of the Everglades. The inner shelf is a shallow bay where water circulation is restricted by the Florida Keys and numerous mud banks and islands.

The open shelf (or reef tract), an arcuate belt 5 to 10 km wide, lies east of the Florida Keys and is subject to relatively open circulation. The inner portion of the shelf (Hawk Channel) is blanketed by a thin veneer of grass-covered muddy sands with localized, mound-like buildups (Rodriguez Key) and patch reefs. The outer portion is a more agitated environment covered with clean skeletal sands (White Bank) or coral reefs.

Great Bahama Bank, as well as the other shallow-water platforms that comprise the Bahamas, displays a facies zonation that differs from that of South Florida (Purdy 1963) (figures 41 and 42). The margins of Great Bahama Bank, with more open circulation, are rimmed by coral reefs, ooid shoals, or coralgal sands. The interior of the platform, which is subject to poorer water circulation and shielded from predominantly easterly winds by Andros Island, is covered by lime muds and pelleted muds. The more open shelves intermediate between the platform margin and interior are sites of grapestone deposition.

As the energy is focused along the margins of the platforms, the facies are more variable there than in the interiors. The margins can be classified as windward-, leeward-, or tide-dominated (Hine and others 1981). The windward margins are the most complex: reefs and associated skeletal sands form along open margins; tidal deltas are associated with interisland gaps; wide belts of tidal bars form in re-entrants; and skeletal sands may be transported seaward. Leeward, open margins are dominated by offbank sand transport. These margins are characterized along their edges by wide belts or sheets of nonskeletal sands. Finally, large tidal-bar belts commonly form at the ends of embayments, where tidal currents are more rapid.

Deposition of Holocene, shallow-platform carbonates in

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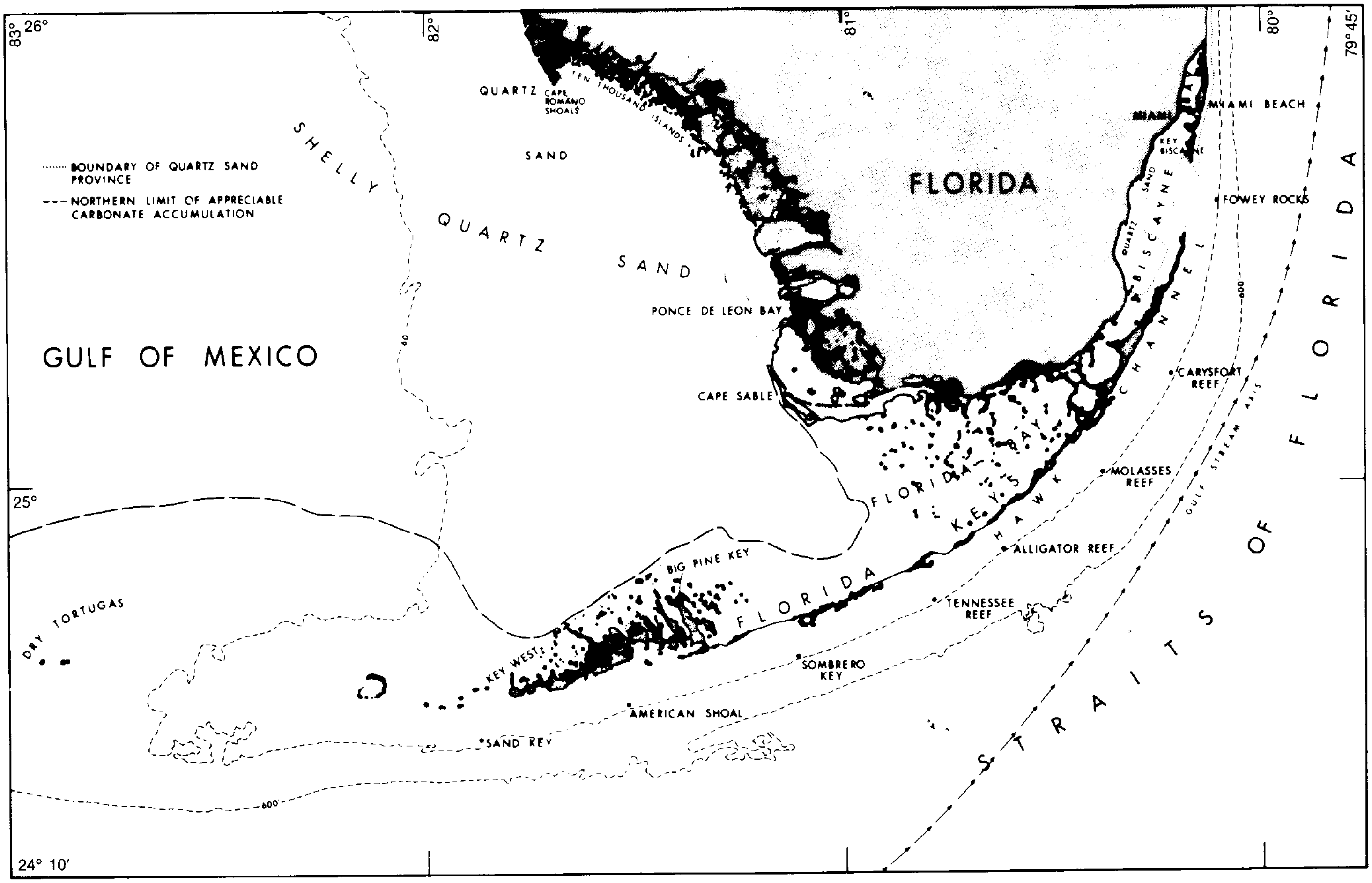


Figure 39.—Index map for South Florida (from Enos and Perkins 1977).

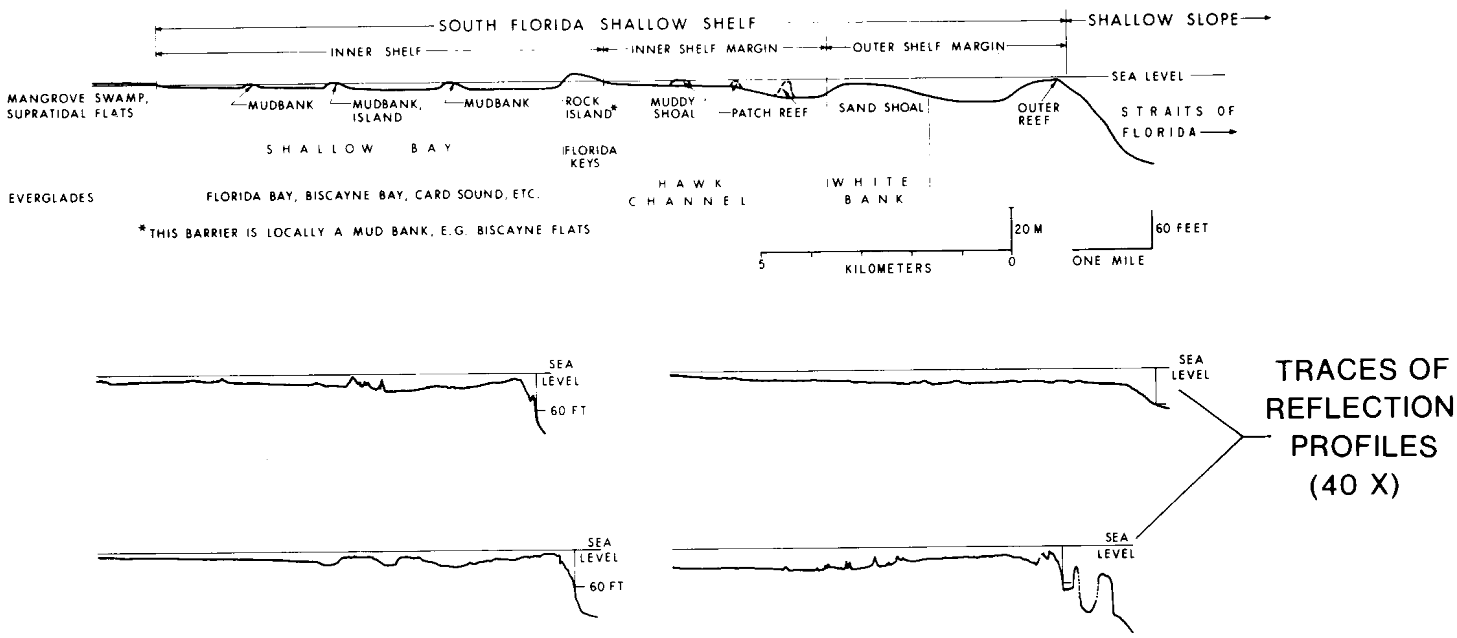


Figure 40.—Bathymetry of South Florida shelf (from Enos and Perkins 1977). See figure 39.

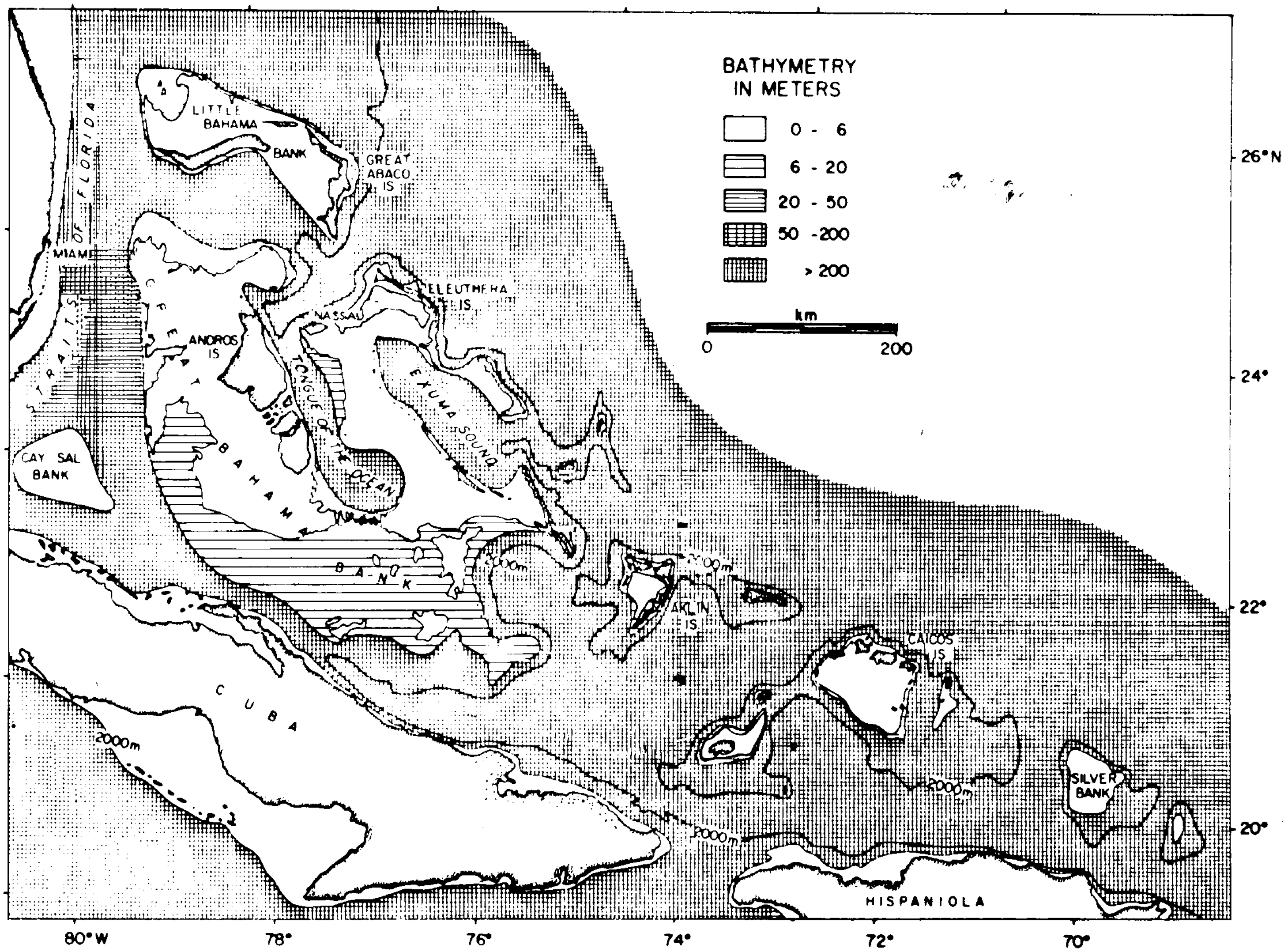


Figure 41.—Bathymetric chart of the Bahama banks (from Ginsburg and James 1974a).

South Florida and the Bahamas did not begin until the margins and flat-topped interiors of the platforms were submerged during the latest rise in sea level. The submergence curve of Scholl and others (1969) for South Florida suggests that the rise in sea level has slowed from 30 cm in 250 years between 5500 and 3500 y.b.p. to less than 30 cm in the last 1000 years. Based on data from stable areas of the world (Shepard 1963), the Florida curve represents the tail end of a flooding event that began as early as 15,000 y.b.p. (figure 43).

The facies succession, as revealed by coring of modern environments, is a function of changes in the rates of sea-level rise and sedimentation. Such relative sea-level changes may be the products of eustatic fluctuations, but may also be a response to subsidence or uplift. Relative sea-level rise has an obvious effect on the sediment type and nature of deposition; the rate and extent of relative sea-level fall markedly affect the diagenesis and erosion of carbonate sequences.

Shoaling-upward cycles in carbonates are common in stable platforms and shelves (James 1979a). The shelf interior has few complete shoaling-upward cycles because hiatuses are common and not all sea-level rises extend all the way across the shelf interior. In contrast, shelf-margin and basin centers may lack shallow-water sediments because subsidence was so rapid that evidence of the progradation cycles is obscured.

Thus, where subsidence was extremely fast, as on a basin margin immediately following continental breakup, cycles can be hidden by rapid subsidence. Instead of the asymmetric shoaling-upward cycles common to stable shelves, symmetrical shoaling and deepening cycles might be produced.

REEFS AND ORGANIC BUILDUPS

Reefs and organic buildups commonly form where there is a break in slope on the seafloor (figure 44), or landward of this break, within the slightly deeper water of platform interiors and epeiric settings. Most reefs and buildups are either continuous and parallel to the depositional strike of the shelf edge or a series of isolated buildups on either side of the shelf break.

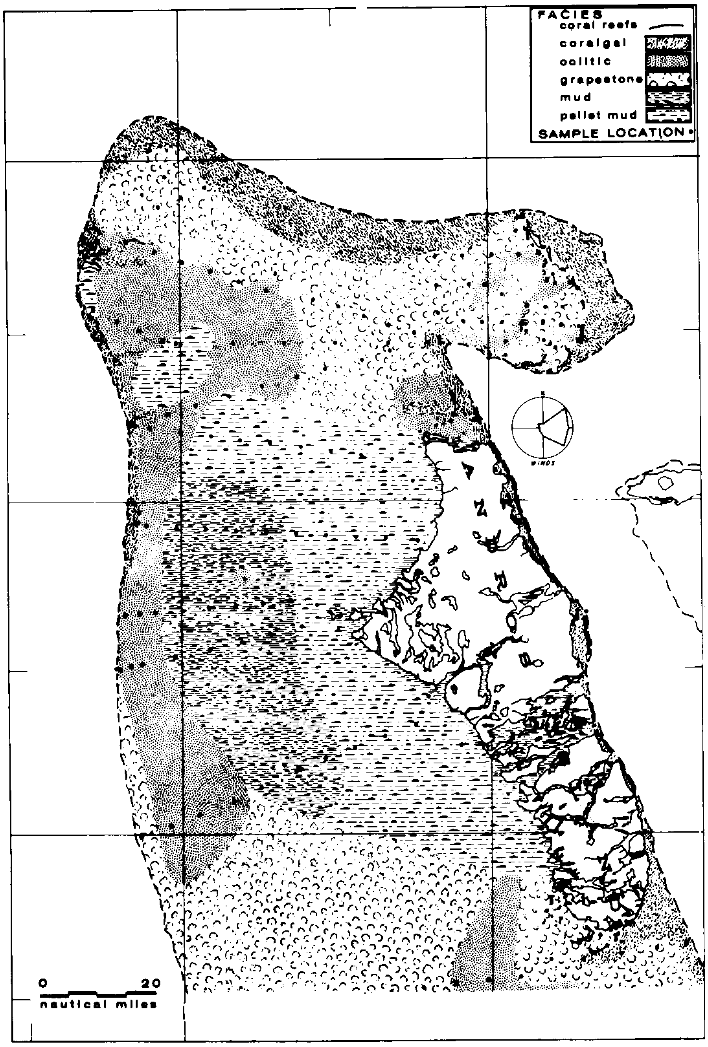


Figure 42.—Sediment distribution patterns on Great Bahama Bank (Purdy 1963).

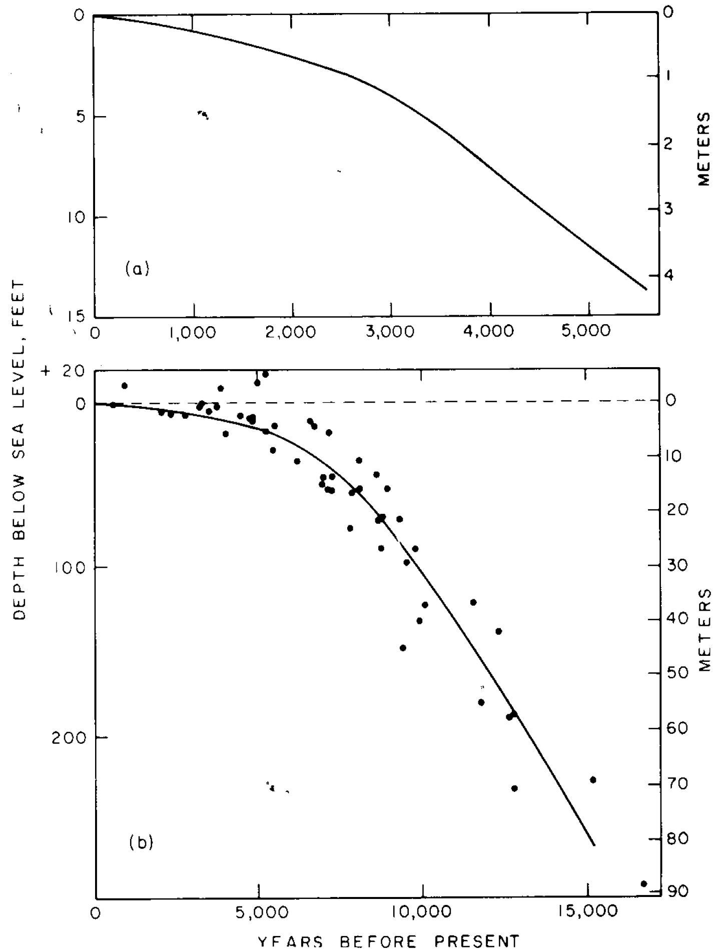


Figure 43.—Holocene sea-level curve for South Florida (from Enos and Perkins 1977).

Barrier Reefs and Mud-Skeletal Banks

Reefs and mud-skeletal buildups are best developed where open marine waters shoal against a basin margin. The slope of the seafloor on which the buildup grows is controlled by antecedent topography, faulting or the juxtaposition of active shallow-water accumulation and deeper-basin starvation. Barrier reefs tend to be massive, but have associated, discontinuous, thin beds of sediment.

Reef geometry is expressed as thick sheets or ribbons that parallel depositional strike. The major subenvironments are the reef frame (the reef crest and seaward wall), seaward reef apron, back reef, and barrier islands (sand cays). Reefs act as sediment sources for areas both landward and seaward. Major reef contributors through geologic time have included corals, stromatoporoids, calcareous sponges, algae, and rudists. Associated fauna are very diverse. (See James and Macintyre 1985.)

The reef frame is characterized by in situ growth of calcareous organisms interbedded with calcareous sands, silts, and muds that form as the result of bioerosion and episodic storms (figure 45). The frame is usually massive and cavernous, with voids filled by bladed and fibrous marine cements and by internal sediment that is commonly perched on or within these cements. Within the reef crest, the skeletal framework may vary from 20 to 80 percent of the rock volume, with a reciprocal distribution of sediment-cement infill.

The reef apron is composed of silt- to boulder-sized debris derived from the reef frame and mixed with in situ fore-reef biota. It typically has a chaotic texture, but may locally exhibit cross-bedding. Many cited examples of Holocene fore-reef and upper-basin slope deposits contain huge blocks of reef rock that slumped from the cliff-like fore-reef face. Precipitous fore-reef slopes are characteristic of Quaternary and Holocene reefs, which owe much of their relief to antecedent topography. Similar fore-reef cliffs occur in the Upper Devonian of the Canning Basin in Australia and in parts of the Mesozoic margin of the East Coast of United States. Most pre-Holocene coral-reef buildups, however, lack this steep fore-reef cliff, and have correspondingly less reef-core rock rubble.

Reef-apron sediments may be stabilized or encrusted by foraminifera, sponges, or algae. A typical fore-reef toe facies of late Mesozoic and Cenozoic reef buildups is composed of a

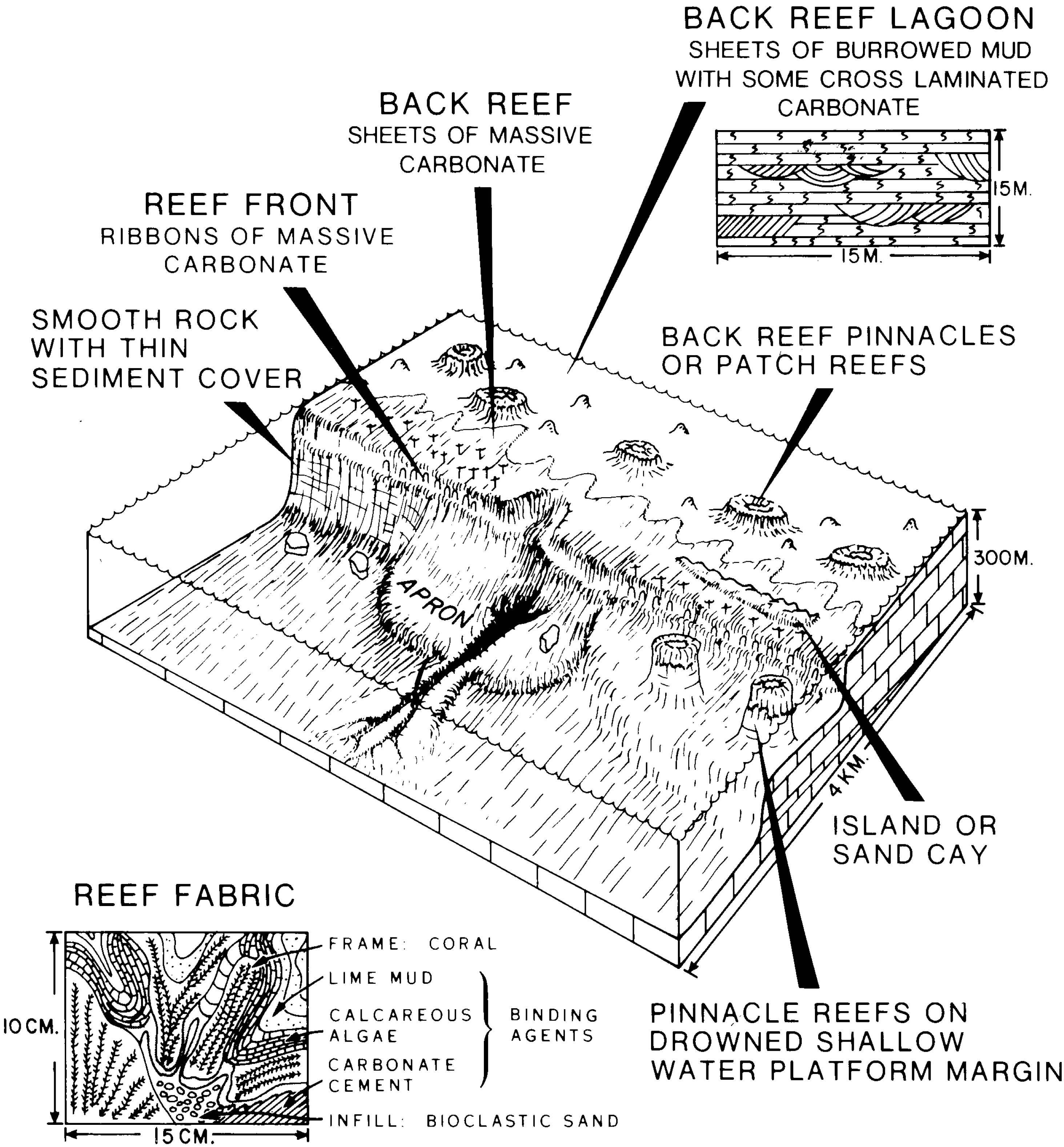


Figure 44.—Generalized view of reef environments and facies.

"gravel" of irregular red algal nodules.

Back-reef sediment is formed by both localized patch-reef framework, which grew as carpets or patch reefs, and by skeletal debris transported from the reef crest. The patch reefs tend to be massive and lens-like, while the adjacent back-reef sediments are generally burrowed, widespread, sheet-like, and varied in

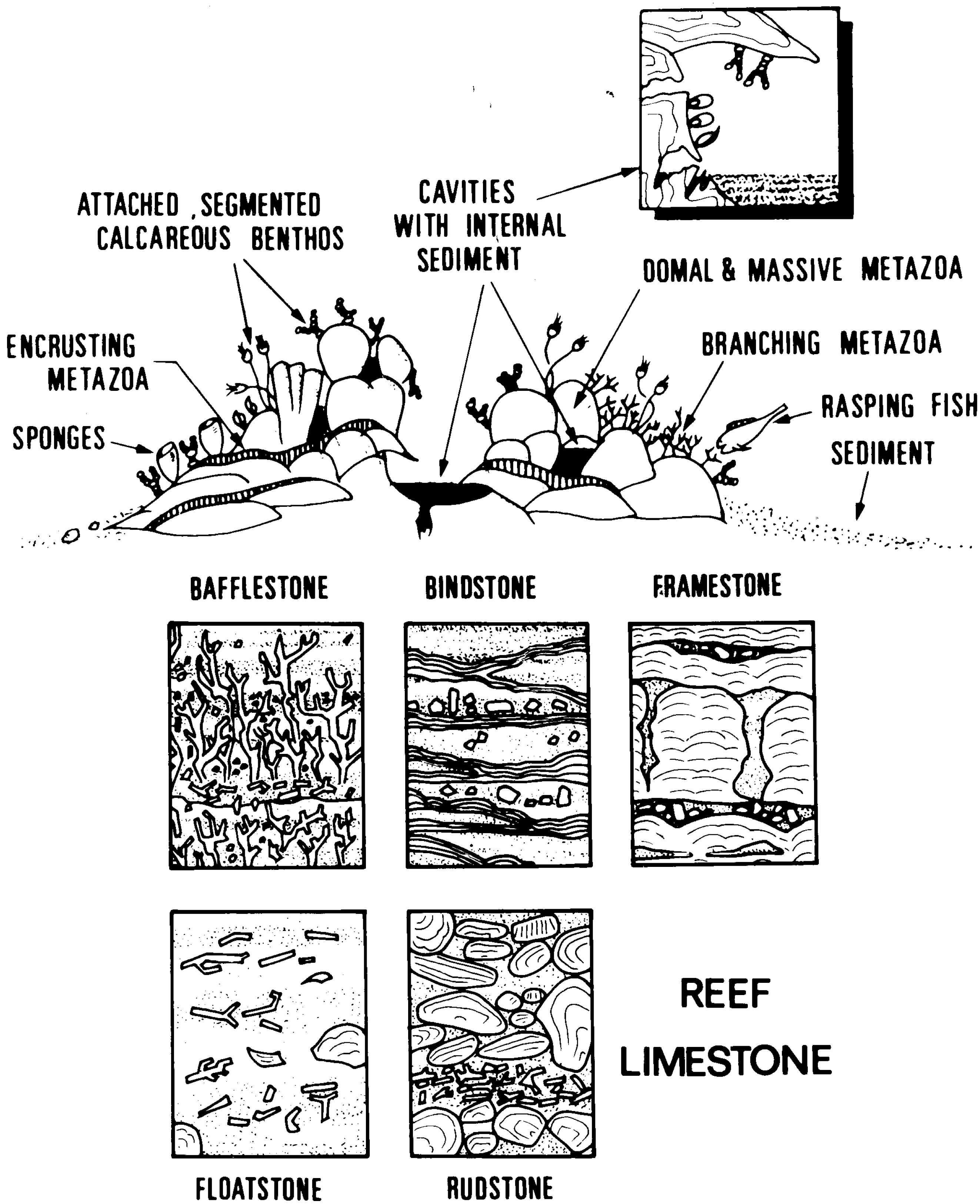


Figure 45.—Reef mosaic of organisms and sediment and the various limestones it can produce (from James 1979b).

grain size from sand to mud. Barrier islands or beaches may occur just behind the reef crest and show many of the characteristics of siliciclastic barrier islands. They form by a complex of linear carbonate sand bodies parallel to depositional strike.

The seaward margin tends to be smooth, but the lee side is serrated by storm washover fans and flood-tidal deltas. Sedimentary structures associated with the islands include cross-laminated carbonate sands from the beach face, lamellar fenestral (bird's-eye) limestone, algal stromatolites, and storm washover layers. Early diagenetic changes and cementation may produce beach rock and tepee structures.

Mud-skeletal banks are massive elongate bodies that form both parallel and perpendicular to the seaward edge of the platform margin. They range from knoll-like mounds of a few square meters to massive linear belts trending for hundreds of kilometers along depositional strike. The thickness of the banks varies from one to over 100 meters. Beds may be thick to massive and range from horizontal to clinoform. Modern carbonate mud banks form in conjunction with sea grasses and green calcareous algae that bind and trap fine-grained sediments derived from breakage in more turbulent water. Sediment in ancient banks of this kind varies from lime mud to fossiliferous sand, is commonly neomorphosed, and may contain cavities filled by sediment and cement.

Pinnacles, Patch Reefs, and Mounds

Pinnacles form during relatively rapid sea-level rises, when carbonate production only locally keeps pace. Bottom agitation is not as great over pinnacles, patch reefs, and sediment mounds as on shelf-edge reefs; therefore, organisms tend to be different, and winnowing and frame-building are less important. These structures are also more symmetrical than shelf-edge reefs, and relatively less oriented with respect to waves and winds. Pinnacles and patch reefs are formed by frame builders. "Mounds" are designated as accumulations of lime silt and mud trapped by sponges, octocorals, algae, and crinoids.

Pinnacles, reefs, and sediment buildups are localized landward or seaward of the crest of the basin margin. They may be localized on highs formed by previous karst topography or some other local irregularity that causes waves to shoal and break or focuses swift tidal currents. Core facies of these bodies are generally massive to thick-bedded, while the flank beds have thin, irregular beds. Changes in texture tend to radiate outward from the buildup core. Seaward buildups commonly contain more porous, coarse-grained carbonates than the more shelfward mounds, but pore-filling marine cements occur more readily in a seaward direction.

As with barrier-reef buildups, the pinnacle reefs are characterized by in situ boundstones of calcareous organisms and sediments. The reef frame is massive and cavernous, and the voids are filled with sediment and marine cement. These sedimentary features are exquisitely displayed in Silurian pinnacles from the Michigan Basin. Major facies variation occurs as buildups that initiated in deep water grew upward into shallow water. Their basal sediments are usually finer grained than their crests. The fauna at the base are usually a pioneer community of low diversity, while the fauna of the crest may be a more diverse climax community. Lower contacts are gradational with the platform sediments below the bodies. Most pinnacles are sharply overlain by basinal marls and shales similar to those deposited on the deeper parts of the platform. In rare instances the bodies coalesce upward and are sharply overlain by tidal-flat sediments.

Potential Reservoir and Source Rocks

The belt of reef and mud buildups at the depositional surface tends to be a narrow, ribbon-like feature less than about 100 m wide. The apron of skeletal sand shed back of the reef may be even narrower, while lagoonal sediments may stretch for tens of miles back of these buildups. All these facies may be quite extensive in the subsurface, due to basinward progradation.

The reservoir potentials of reefs and buildups are widely assumed to be high. However, studies indicate that the porosity of reef boundstones is more often than not plugged by both primary and secondary cements and internal sediments. Proximal back-reef sand deposits can retain significant amounts of primary porosity, especially in reef tracts where accumulation of skeletal rubble was rapid. Quiet-water carbonates of deeper lagoons tend to be muddy sediments (that is, wackestones and packstones) with relatively low porosities and permeabilities. Fore-reef deposits and the aprons of mud buildups may have somewhat greater reservoir potential, especially if the reef itself is plugged by carbonate cements and acts as the updip seal in a stratigraphic trap.

Some barrier-reef deposits have proved to be major hydrocarbon reservoirs, although most lack an immediate updip trap. An excellent example of a giant field in a barrier reef is the Oligocene reef complex at Kirkuk, Iraq. Part of the Devonian Leduc reservoir trend of Western Canada has the characteristics of both a linear mud-skeletal margin and barrier-reef complex.

Reef-tract and linear mud-skeletal sediments typically have relatively low source potential. This is in part due to the shallow, turbulent environment, but also to the efficient recycling of organic detritus within the reef community's trophic structure. Thus, little organic debris "leaks" from the crest community into the apron or the back margin lagoon.

In contrast to barrier facies, major hydrocarbon discoveries are common in ancient pinnacle-reef and mud buildups. Reservoir volumes of pinnacles tend to be more sharply limited than for shelf-edge reservoirs. This is because the porous core facies is typically bounded by either relatively impermeable flank and margin deposits, basinal shales, or basinal evaporites. These deposits form the seal, but make recharge of reservoir hydrocarbons unlikely. Examples of oil fields in pinnacles are the Silurian of the Michigan Basin and the Devonian of Western Canada. Pinnacles may be associated with fairly rich source rocks at their flanks and in basinal sediments enclosing them. Organic productivity and preservation in the sedimentary column tend to be high for pinnacles situated on the lower basinal slope, but decline updip.

South Florida Reefs and Banks

The sedimentary package of the open shelf of South Florida shows upward changes to progressively less restricted environments of deposition (Enos and Perkins 1977). The vertical sequence closely parallels lateral changes in depositional textures, sedimentary structures, and organic populations, all of which can be observed in surface sediments from the inner shelf to the shelf break. Upward in the sequence, grain size increases erratically, the percentage of fine sediment decreases, sorting improves, sedimentary structures are less disrupted by bioturbation, and the organisms present indicate open circulation. *Halimeda*, corals, and red algae are more abundant in the top of the sequence, whereas mollusks are more common toward the base.

Stratigraphic cross sections show that the sequence is basically transgressive, reflecting progressively more open circulation and greater agitation up to the present (figure 46). The relative restriction in the lower part of the sequence probably resulted from the depositional topography, such as reefs at the shelf break, early in the Holocene cycle (Enos and Perkins 1977). Deposition occurred during a continuous but decelerating eustatic sea-level rise; however, the overall transgressive sedimentary sequence may have been broken by laterally discontinuous regressive sequences due to restriction behind local depositional topography.

Rodriguez Bank. The inshore zone of the open shelf is covered with a thin veneer of grass-covered muddy sands and localized, mound-like buildups. One such buildup, Rodriguez Bank, is a mound of unconsolidated calcareous sediments deposited with no rigid organic framework (Turmel and Swanson 1976). Major contributors were green algae, red algae, and corals (figure 47). Surface sediments vary from skeletal sands and gravels to variable mixtures of lime mud and skeletal sand.

The bank was probably initiated by an embayment in the Pleistocene rock floor which acted as a trap for lime muds. During its early stages the bank developed in quiet water where circulation was restricted. Sediments were primarily lime muds with sparse skeletal fragments (figure 48). With rising sea level, the depositional environment changed to more open circulation. Coral-algal sands and gravels accumulated, forming the sediment distribution and ecological zonation that typify the present bank.

This nearly flat-topped, intertidal, algal bank has a distinct zonation along its windward (east-facing) side (figure 49). Immediately surrounding the mangrove-covered island (Rodriguez Key) is the bank top, where marine grasses and calcareous green algae are common, along with bivalves and burrowing crustaceans. The bank is fringed by narrow bands of intergrown, branched finger corals (*Porites porites* var. *divaricata*) and branched, twig-like coralline algae (*Goniolithon strictum*). These narrow bands are relatively grass-free. Off-bank, in 2 to 3 m of water, the bottom is similar to the bank top, but with the addition of sponges, additional small-branched corals, and sea urchins.

Rodriguez Bank shows one possible evolutionary path that a mud bank can take. It is similar to those in Florida Bay (discussed below). Rodriguez Bank has developed a cap of coral-algal framework as a result of changing circulation patterns and water depth on the open shelf. A similar pattern exists in many ancient mud mounds, of which good examples are the Upper Paleozoic phylloid algae buildups and Cretaceous rudistid mounds.

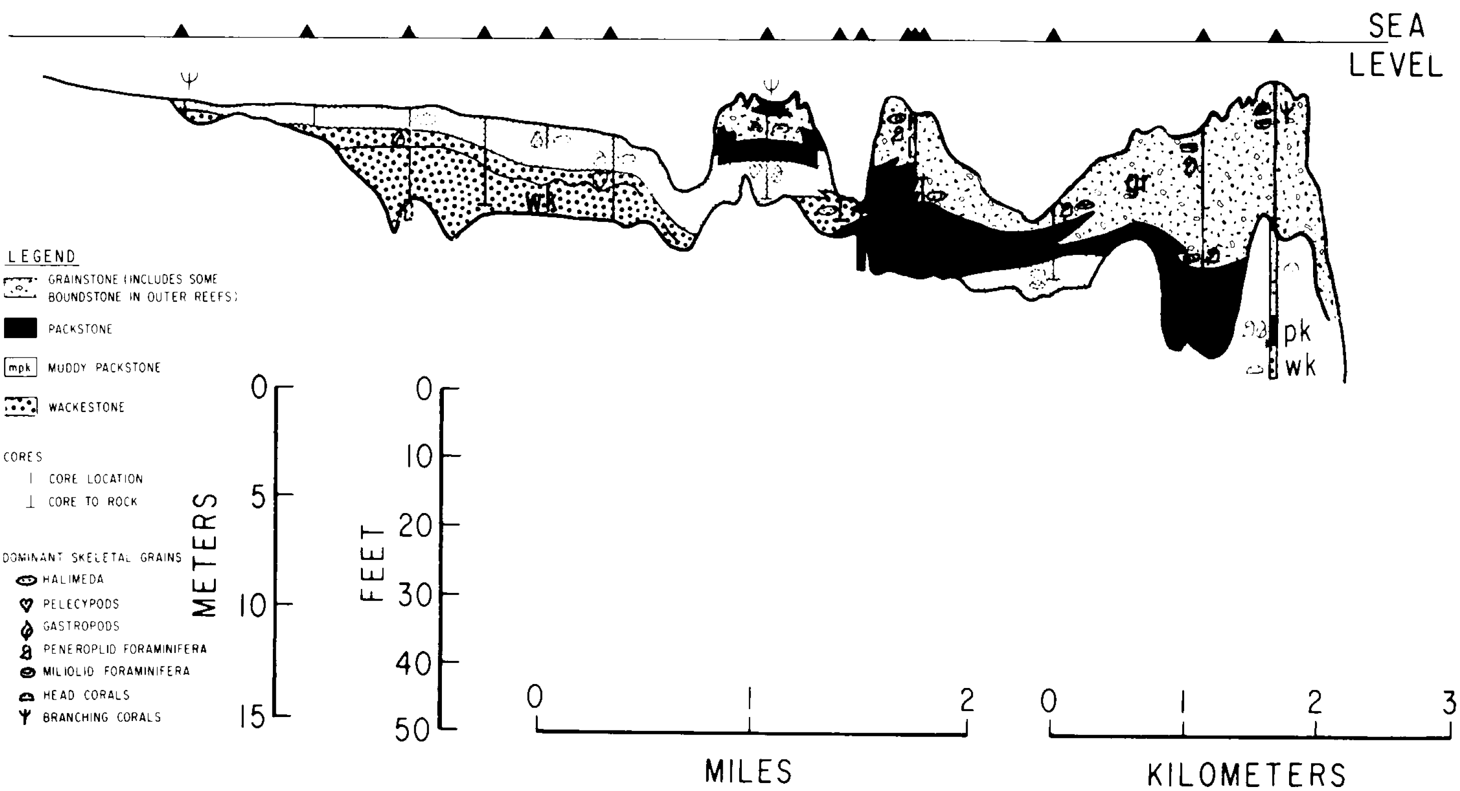


Figure 46.—Cross section of open shelf of South Florida (from Enos and Perkins 1977).

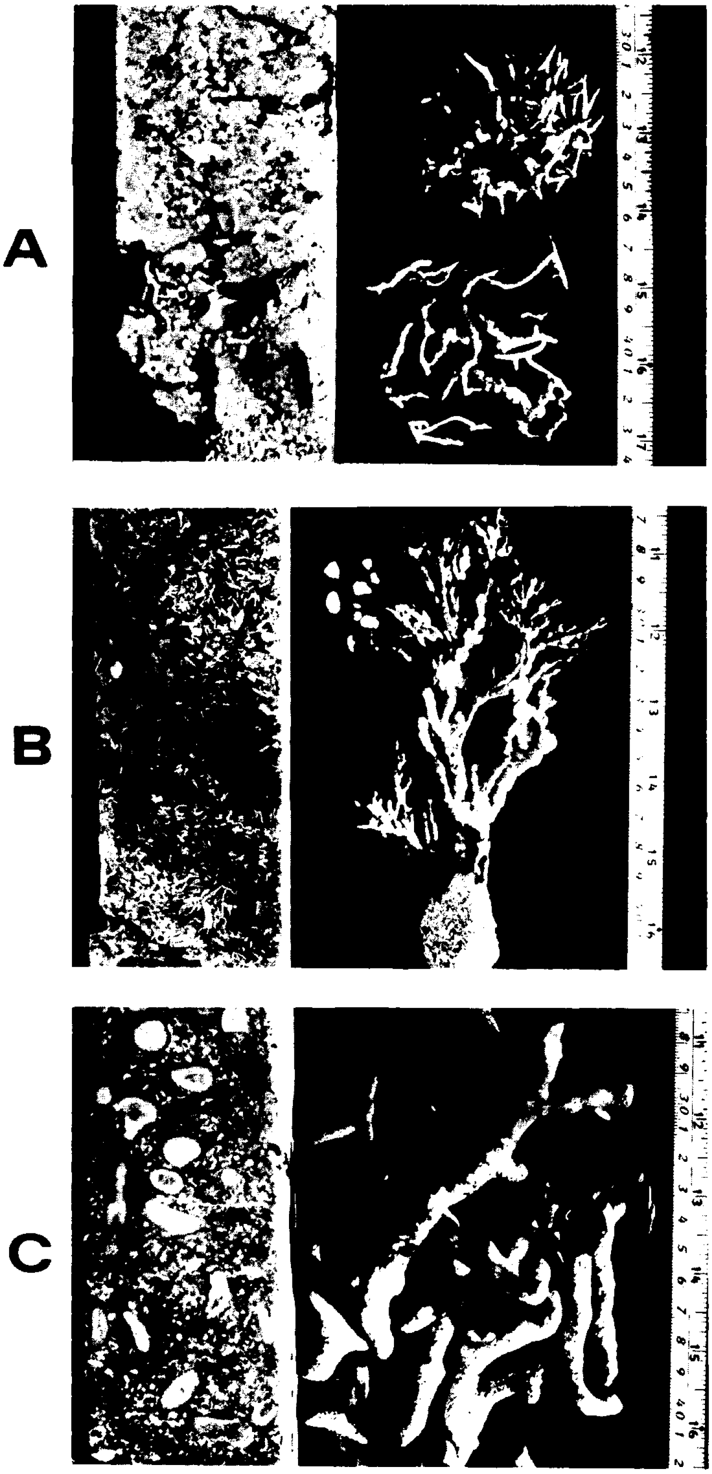


Figure 47.—Important skeletal constituents in sediments of Rodriguez Bank: branched coralline algae *Goniolithon strictum* (A), Codiacian algae *Halimeda tridens* (B) and branched coral *Porites porites* (C) (from Ginsburg and James 1974b).

Shelf-edge reefs. The outer part of the Florida open shelf (or reef tract) is made up of coral reefs and clean skeletal sands (figure 50). Typical of the Florida reefs, the reef flat is composed of skeletal sands and coral fragments and is covered with only a few feet of water at low tide. A typical profile of Florida reefs is compared with profiles from Jamaica and Belize in figure 51.

A terrace of oriented branching corals slopes from the reef flat down to about a 3-m water depth, and a second terrace covered with massive head corals extends to about 7 m deep. Both terraces are incised by grooves that run normal to the trend of the reef. These sand-bottomed grooves are 3 to 6 m wide and 3 m deep. Coralline limestone spurs between the grooves may be 60 m wide. The spurs form by differential coral growth; the shifting substrate within the grooves is unfavorable for coral colonization.

Except for the absence of the elkhorn coral (*Acropora palmata*), the organisms on the Florida patch reefs are the same as those observed on the shelf-edge reefs. The significant differences between the two types of reefs are their positions on the open shelf and the geometries of the buildups. The shelf-edge reefs form as discontinuous, relatively narrow belts along the strike of the actual shelf edge and develop back-reef and fore-reef deposits. In contrast, patch reefs are scattered across a very wide belt covering the outer half to two-thirds of the open shelf. They occur as isolated or coalescing buildups within a more widespread sheet of skeletal sands.

Sand produced by reefs frequently forms sand bodies landward of reefs. White Bank, a shoal area about 5 km east of Key Largo, Florida, is such a sand body (Enos and Perkins 1977) (figure 52). Algal, mollusk, and coral sands form a belt 1 to 2 km wide and 40 km long. Although large portions of the belt are stabilized by sea grass, rippled sands are common. Both active and stable bottoms occur at similar water depths (1 to 3 m), suggesting that scattered large sand waves must be due to localized hydrologic conditions. The sand body is asymmetrical in profile; the steeper lagoon side suggests landward transport of sand.

Most Holocene reefs of South Florida have developed over pre-existing highs. These highs may be earlier Holocene dune or reef topography or may lie along a major break in slope formed by reef growth during Pleistocene time (Enos and Perkins 1977). The reefs have grown at rates of 1 to 5 m per 1000 years (Shinn and others 1977). In some areas the rise in sea level eventually outpaced the reef's ability to grow. In other areas faster buildup formed thicker accumulations. Where reefs reached the surface, the accumulation rate eventually ceased because the rate of relative sea-level rise slowed during the latest Holocene.

Coring of the Florida reefs shows that they have a coral-algal framework, plus accumulations of sand-size sediment veneered with coral rubble and scattered in-place colonies (figure 53 and 54). The rate of upbuilding of coral reefs is mostly controlled by the type of coral, *Acropora palmata*, which dominates most Caribbean shallow-reef communities, is capable of vertical rates that match the rate of sea-level rise at any time during the Holocene (Adey 1975). Coral reefs comprised of *Diploria* and *Montastrea* grow much slower, and thus generally lagged behind a rapidly rising sea level. A declining sea-level rise or stillstand might allow these slower growing reefs to reach close enough to the surface to develop *A. palmata* communities.

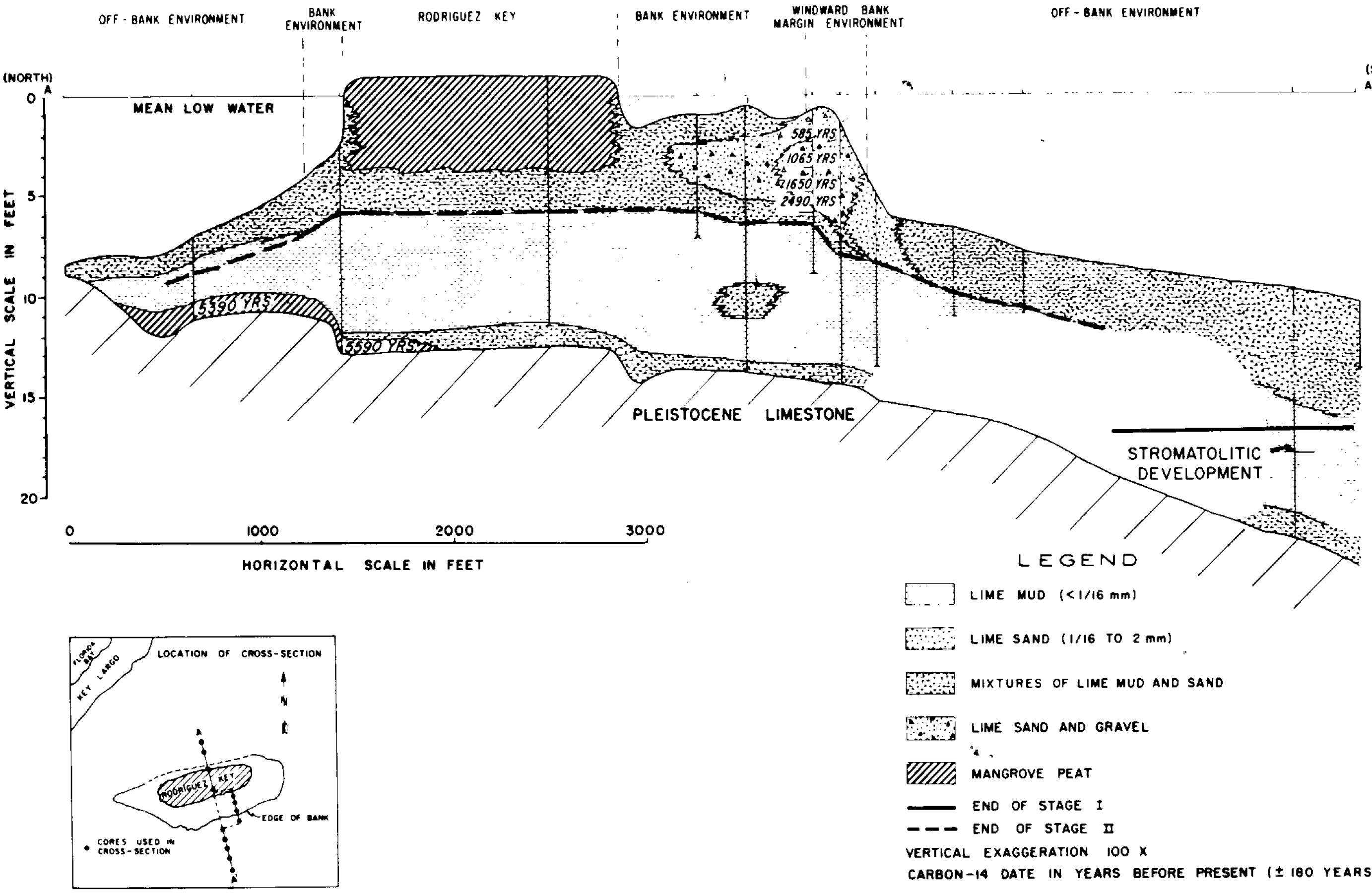


Figure 48.—Stratigraphic cross section of Rodriguez Bank (from Turmel and Swanson 1976).

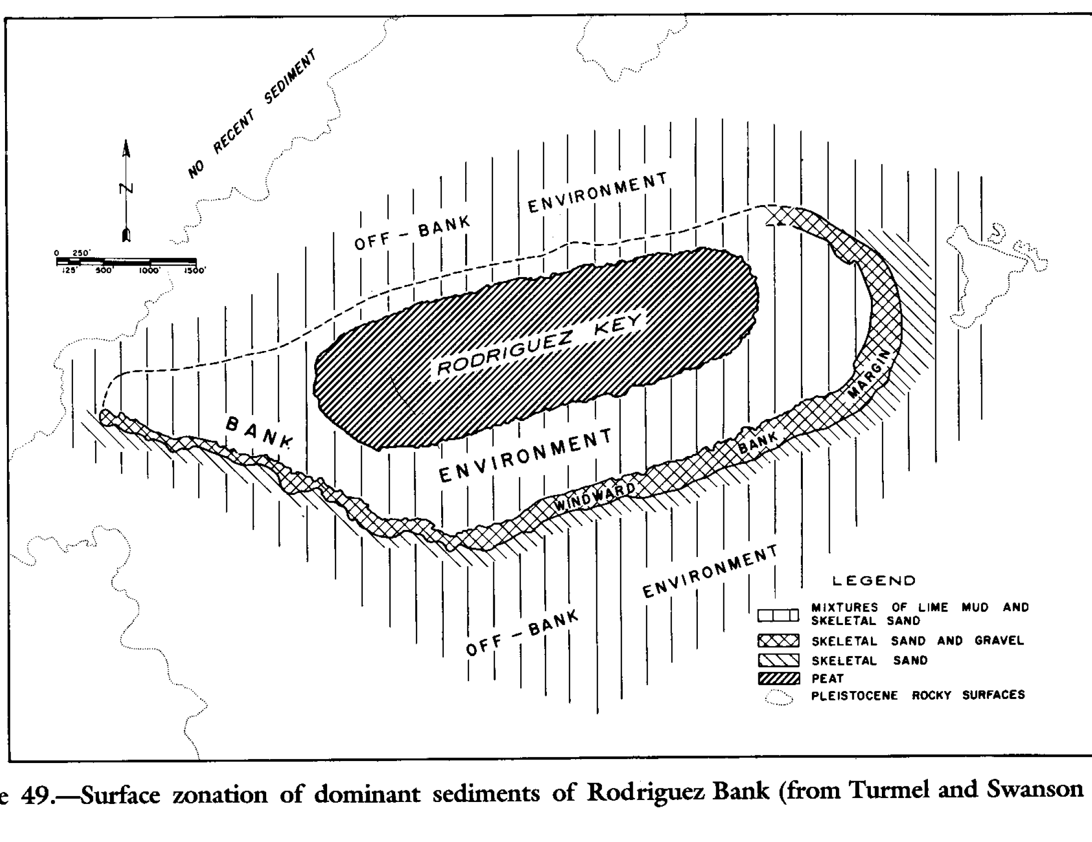


Figure 49.—Surface zonation of dominant sediments of Rodriguez Bank (from Turmel and Swanson 1976).

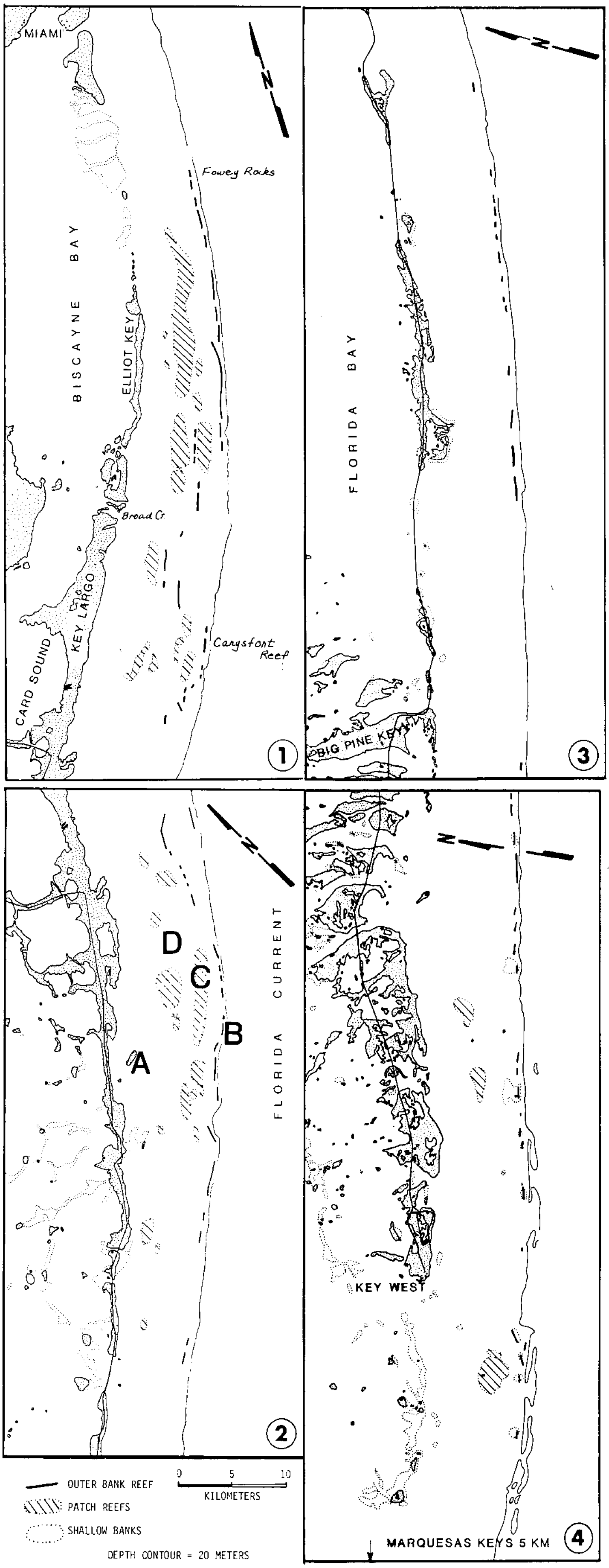


Figure 50.—Four contiguous maps from Miami to Key West showing reef distribution in the Florida Keys (from Marszalek and others 1977): (A) Rodriguez Bank, (B) shelf-edge reef, (C) patch reef, and (D) skeletal sand shoal (White Bank).

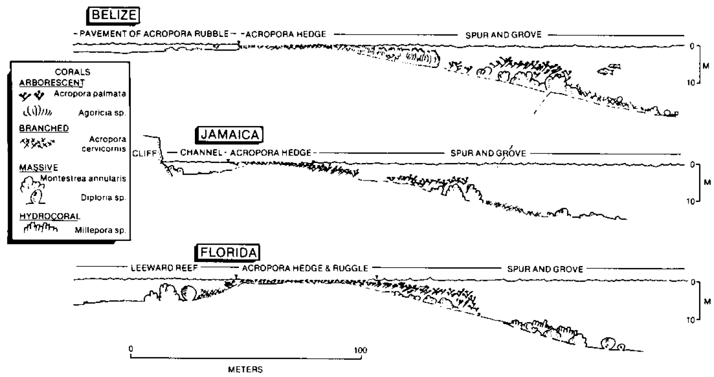


Figure 51.—Morphology and zonation across three zoned marginal reefs (from Ginsburg and James 1974b).

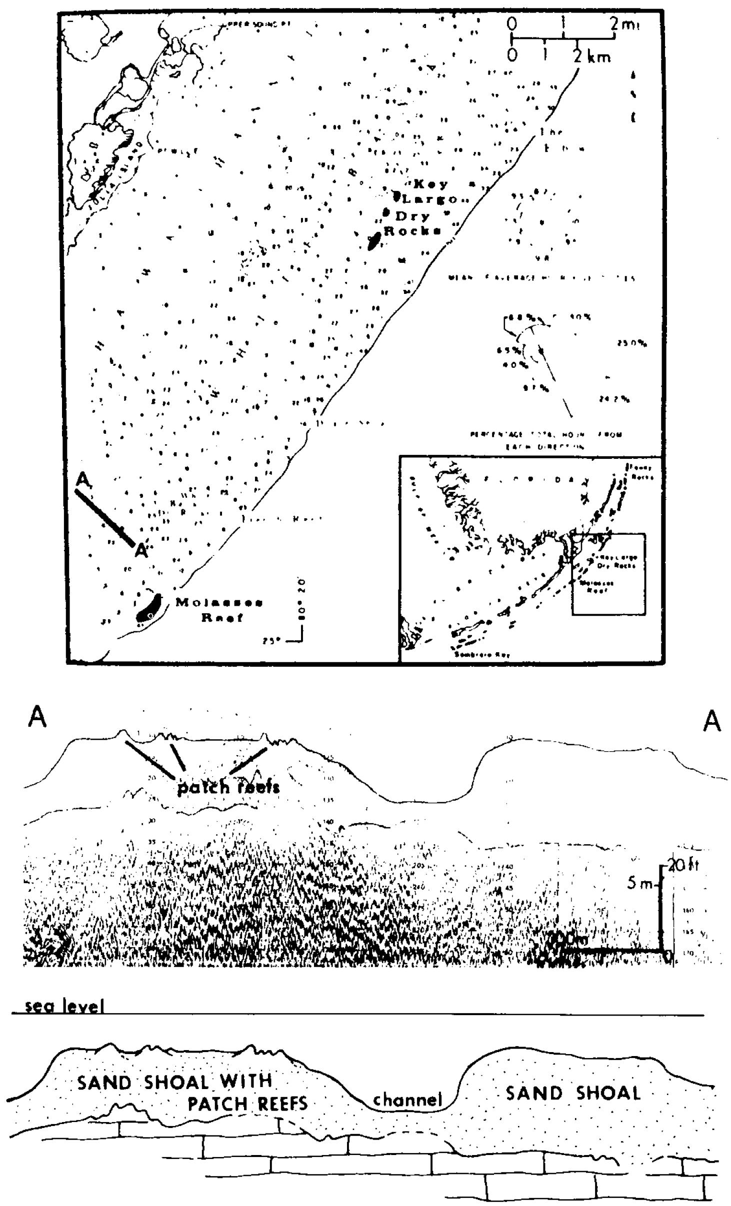


Figure 52.—White Bank skeletal sands and patch reefs (from Enos and Perkins 1977). Location map (top), sparker profile and interpretation (bottom).

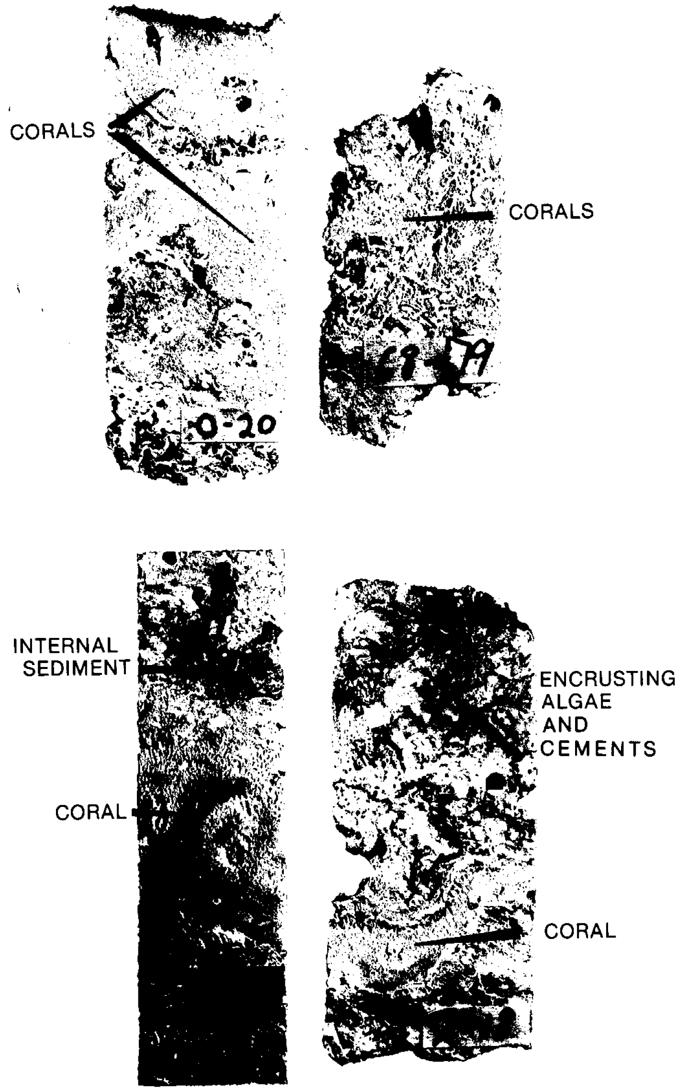
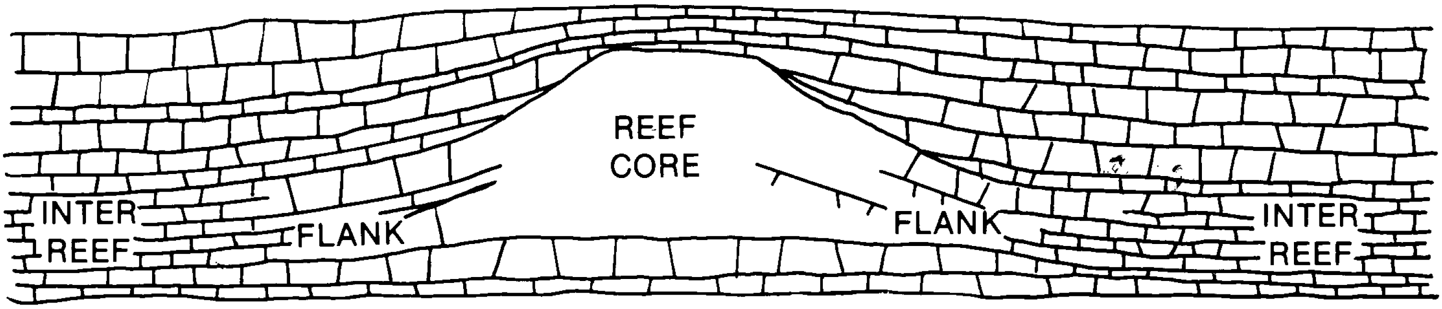


Figure 53.—Core slabs showing reef boundstone fabric. Note the variability in size, shape, and distribution of porosity.

Reefs are important exploration targets in many areas. Our discussion of the open shelf of South Florida introduces the various reef and reef-related lithologies that are likely to be encountered in an ancient example. The size, shape, and orientation of individual buildups, as well as the relationships between buildups and associated sediments, are the large-scale patterns that should prove useful in analyzing subsurface shelf-edge deposits.

SAND SHOALS

Carbonate sand accumulations of reservoir size commonly occur near the seaward edges of banks, platforms, and shelves. Less commonly they form in platform interiors and over topographic highs in regional deep-water settings. Bank-margin sand accumulations may grade over short distances landward



STAGE	TYPE OF LIMESTONE	SPECIES DIVERSITY	SHAPE OF REEF BUILDERS
DOMINATION	bindstone to framestone	low to moderate	laminate encrusting
DIVERSIFICATION	framestone (bindstones) mudstone to wackestone matrix	high	domal massive lamellar branching encrusting
COLONIZATION	bafflestone to floatstone (bindstone) with a mudstone to wackestone matrix	low	branching lamellar encrusting
STABILIZATION	grainstone to rudstone (packstone to wackestone)	low	skeletal debris

Figure 54.—Major reef deposits in schematic cross section and reef core facies in detail (from James 1979b).

or seaward into other environments.

The development of these bodies requires sand-sized sediments and a means of removing smaller or larger material. These requirements are met where a change in shelf slope coincides with wave action or strong tidal currents in a zone of high carbonate production.

In modern carbonate settings, sand bodies occur in many different forms, nearly all of which have ancient counterparts (figure 55). Back-reef sheets, belts, and lobes of skeletal sand form along open platform margins where sediment transport is toward the bank. Gaps between small islands may be the sites of tidal deltas. Commonly the flood tide delta is enlarged as a result of storm-created currents. If re-entrants or embayments occur along a margin, tidal and storm-generated currents can generate wide belts of tidal bars.

Along windward margins, which are dominated by large islands, skeletal sands generated within the fore-reef environment can be carried seaward to the marginal escarpment. There they can accumulate behind rocky barriers or be carried farther seaward into deep water.

In contrast to the variety of sand bodies that form along windward margins, leeward open margins are dominated by offbank sand transport. Here wide belts or sheets of nonskeletal sands form at the bank edge.

The vertical sequence of deposits in modern sand shoals usually records progressive shallowing because these shallow-water sites provide optimum conditions for carbonate production (figure 56). Therefore, platform carbonate sediments usually accumulate at greater rates than that of relative subsidence and repeatedly build up to sea level or above. Cyclical packages form, each a few meters thick. Similar shoaling sequences are recognized in thick, ancient carbonate sand deposits.

Sand shoals do not have good source potential, being oxygenated and highly agitated. Lateral facies equivalents must therefore be called upon to act as the sources of hydrocarbons trapped in sand-shoal reservoirs. In terms of reservoir quality,

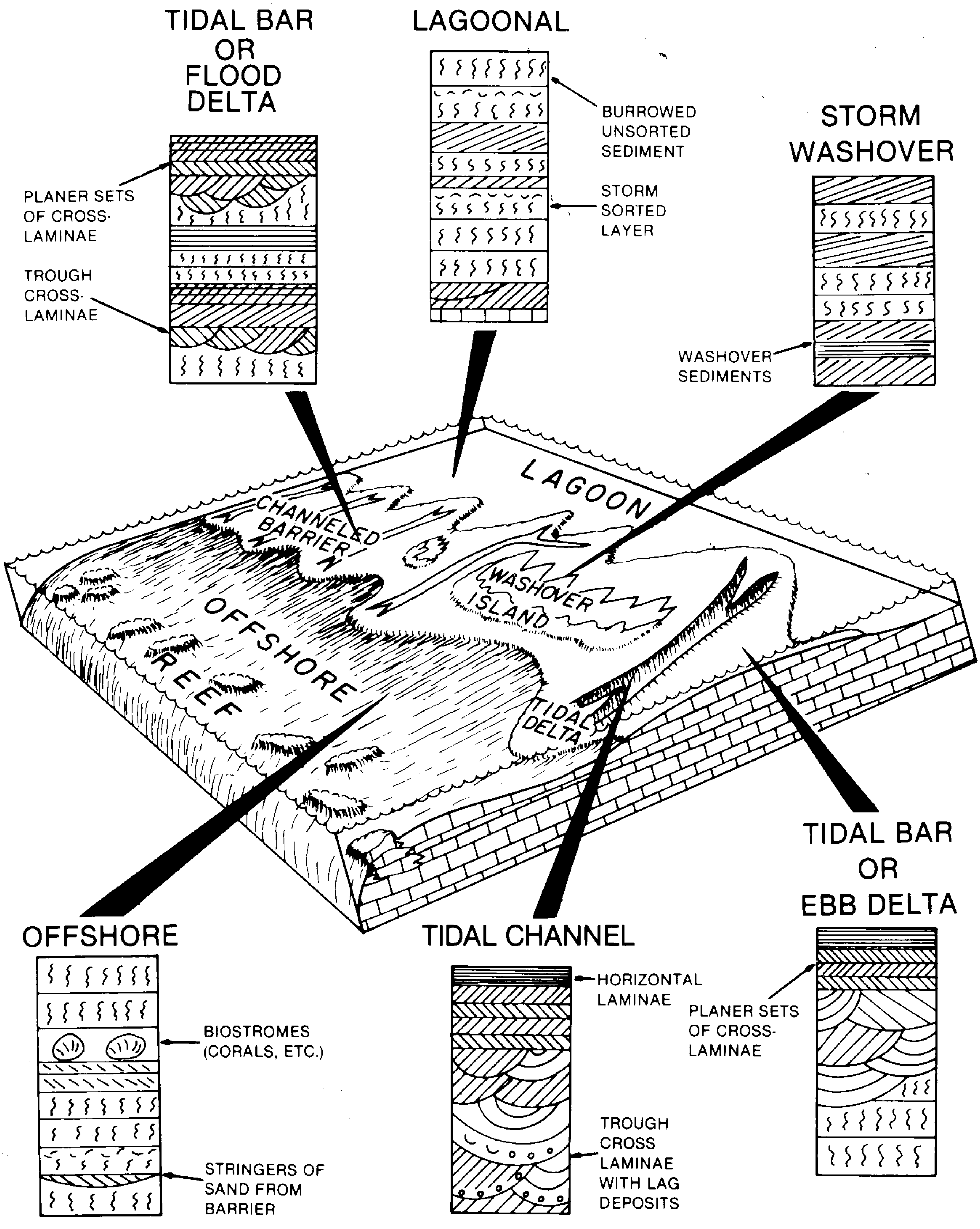


Figure 55.—Generalized view of sand body environments and facies.

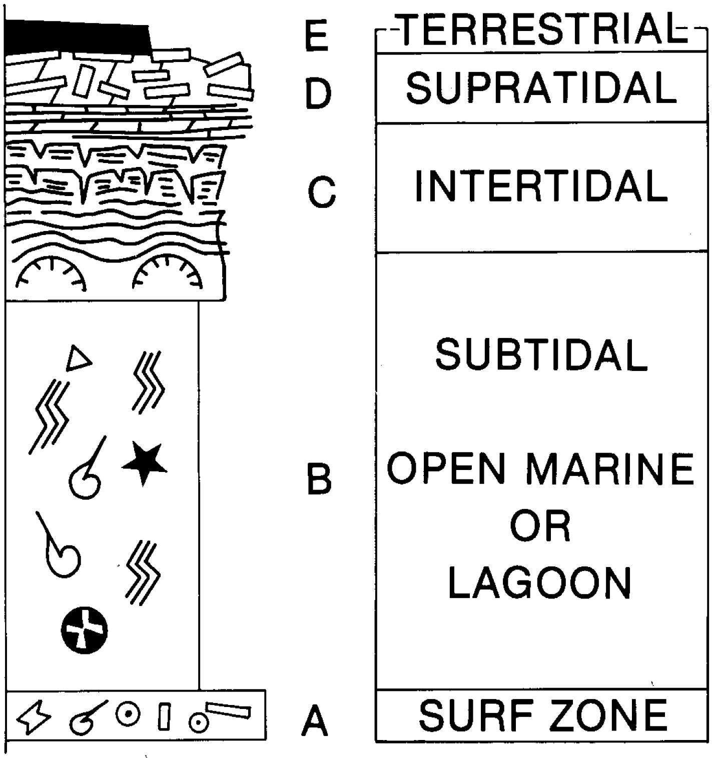


Figure 56.—Schematic of shallowing-upward carbonate sequence. Lithoclast conglomerate recording initial transgression over preexisting deposits (A), fossiliferous or oolitic limestone of normal marine conditions (B), stromatolitic, mud-cracked limestone or dolomite deposited in intertidal environment (C), laminated dolomite or limestone signifying supratidal deposition (D), and calcrete recording subaerial exposure (E) (from James 1979a).

the sands have high initial porosities that may be preserved in the subsurface. Marine cementation, leading to the formation of hardgrounds and intraclasts, rarely cements a thick continuous section of sands. Instead, only localized zones are cemented in otherwise unlithified sediments. Because sand shoals are commonly localized on paleohighs and generally build up to and above sea level, secondary porosity usually develops shortly after deposition.

The best-documented example of production from sand-shoal reservoirs occurs in the Jurassic Smackover Formation of the U.S. Gulf Coast. Other well-known examples are the San Andres Formation in the Permian Basin and the Arab D in the Persian Gulf.

Bahama Ooid Shoals

Both ooid shoals and coral reefs occur along most of the platform margin in the Bahamas. The two environments may co-exist or laterally replace one another along windward margins (figure 57). Along leeward margins, sands transported from the bank may bury older or contemporaneous reefs (Hine and Neumann 1977; Palmer 1979; figure 58).

Ooid sands can form in lobe-shaped or elongate submarine bars, in beaches, in tidal deltas between islands, or in subenvironments associated with these sites. Ooids mixed with other sand grains or with lime mud can be found as islands and dunes, as well as channel bars and levees within vast sand flats. An ooid shoal is a succession of these environments produced during the Holocene sea-level rise.

Coring and seismic profiling of ooid shoals in the Bahamas has been done by Ball (1967), Buchanan (1970), Hine (1977), Dravis (1977), Harris (1979), Palmer (1979), Halley and Harris (1979), and Hine and others (1981). Their studies document facies, sedimentary structures, vertical sequences, and the geometry of bank-margin ooid accumulations. These are the critical sedimentary features in our interpretation of ancient oolite deposits seen in outcrop or in borehole cores.

The primary difference between the tidal-bar and marine-sand types of linear sand belts of Ball (1967) and Halley and others (1983) is their orientation to the trend of the bank-margin. Tidal-bar belts develop perpendicular to the margin, whereas marine sand belts form parallel to the bank edge. Both types respond primarily to daily tidal flows and to wave- and storm-generated currents; but significant differences exist between them, as well as between individual sand accumulations within each sand belt. Such differences result from antecedent topographic control, response to sea-level change or storms, bed-form distribution, the role of diagenesis, development of benthic communities, sediment type, sediment thickness, and lateral facies changes.

Joulters ooid shoal. The Joulters ooid shoal, north of Andros Island on Great Bahama Bank, was described in detail by Harris (1979). This shoal is important because it displays a variety of subenvironments in which ooid sands can accumulate, some of which are quite different from environments where ooids are generated.

Facies recognized in coring studies of the Joulters shoal include ooid grainstone, ooid packstone, fine-peloid packstone, pellet wackestone, and lithoclast packstone. Skeletal grainstone is also present on the shelf seaward of the shoal, but was not recovered in cores.

The area of the present-day shoal includes depositional environments that produce each distinctive facies. Ooid grainstone (figure 59) forms on current-swept, rippled sea bottoms such as the crests of active sand bars; ooid packstone (figure 60) forms on stabilized burrowed bottoms, including sand flats and relict sand bars; fine-peloid packstone collects on stabilized and burrowed bottoms farther from sources of ooid sands; pellet wackestone forms in restricted bottoms (such as in the lees of islands); and lithoclast packstone accumulates in active tidal channels. Each of these facies can occur in other settings as well, collectively forming what we refer to as an ooid shoal in the geological record.

The fence diagram of figure 61 shows the important facies relationships revealed by coring the Joulters shoal. The relief of the shoal above the surrounding seafloor is primarily a result of ooid sands in one of three facies: ooid grainstones in a narrow belt along the ocean-facing borders of the shoal, where ooid accumulation coincides with formation, and the more widespread ooid packstone and fine-peloid packstone facies which are the result of mixing of ooids with other grain types.

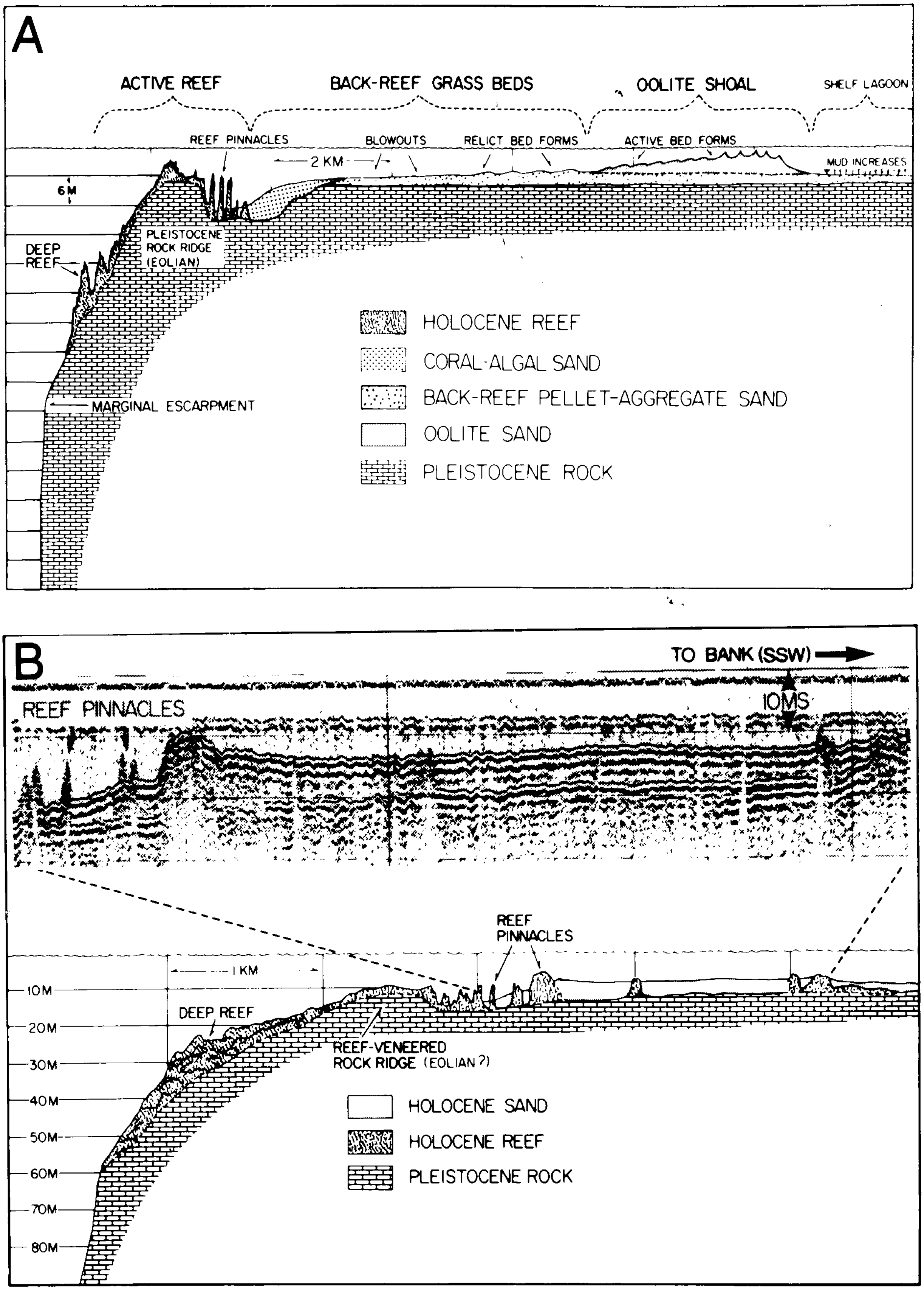


Figure 57.—Cross sections of Bahamian Bank margins: windward margin with reefs and oolite shoals (A) and outer windward margin (B) (from Hine and Neumann 1977).

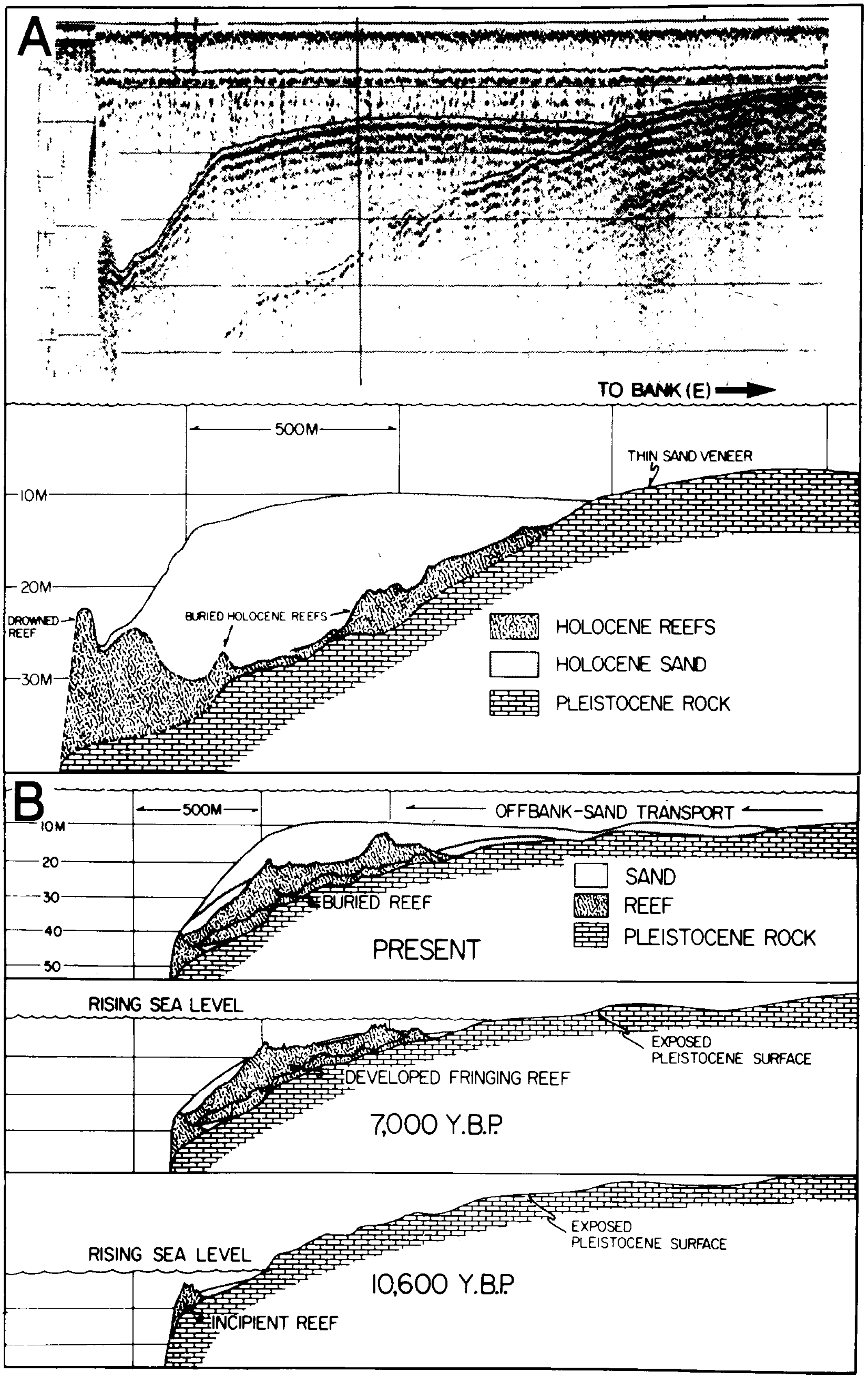


Figure 58.—Cross sections of Bahamian Bank margins: Holocene reef buried by sand along a leeward margin (A) and growth history of leeward margin (B) (from Hine and Neumann 1977).

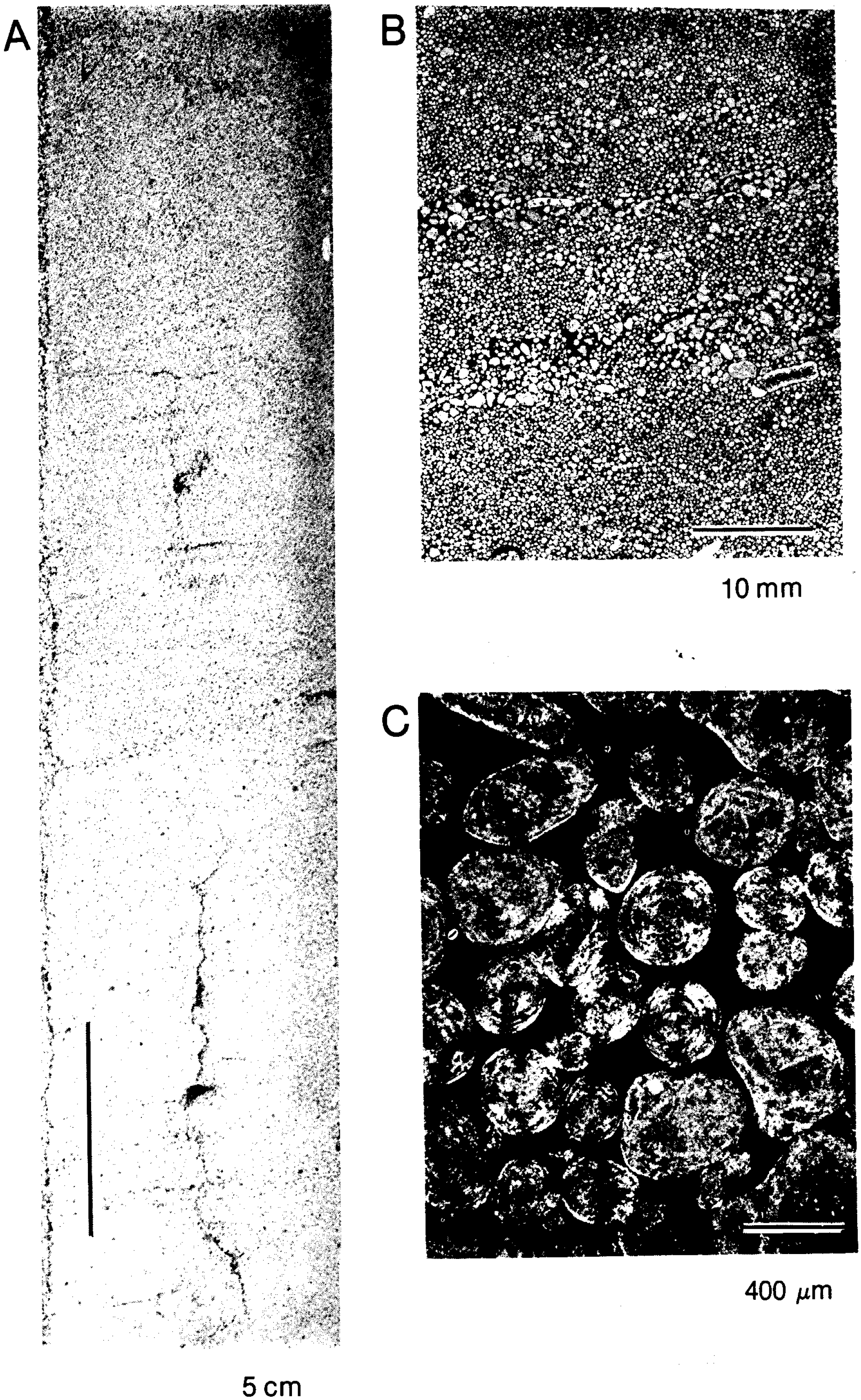


Figure 59.—Ooid grainstone shown in core (A), entire thin section (B), and close-up of thin section (C) (from Harris 1979).

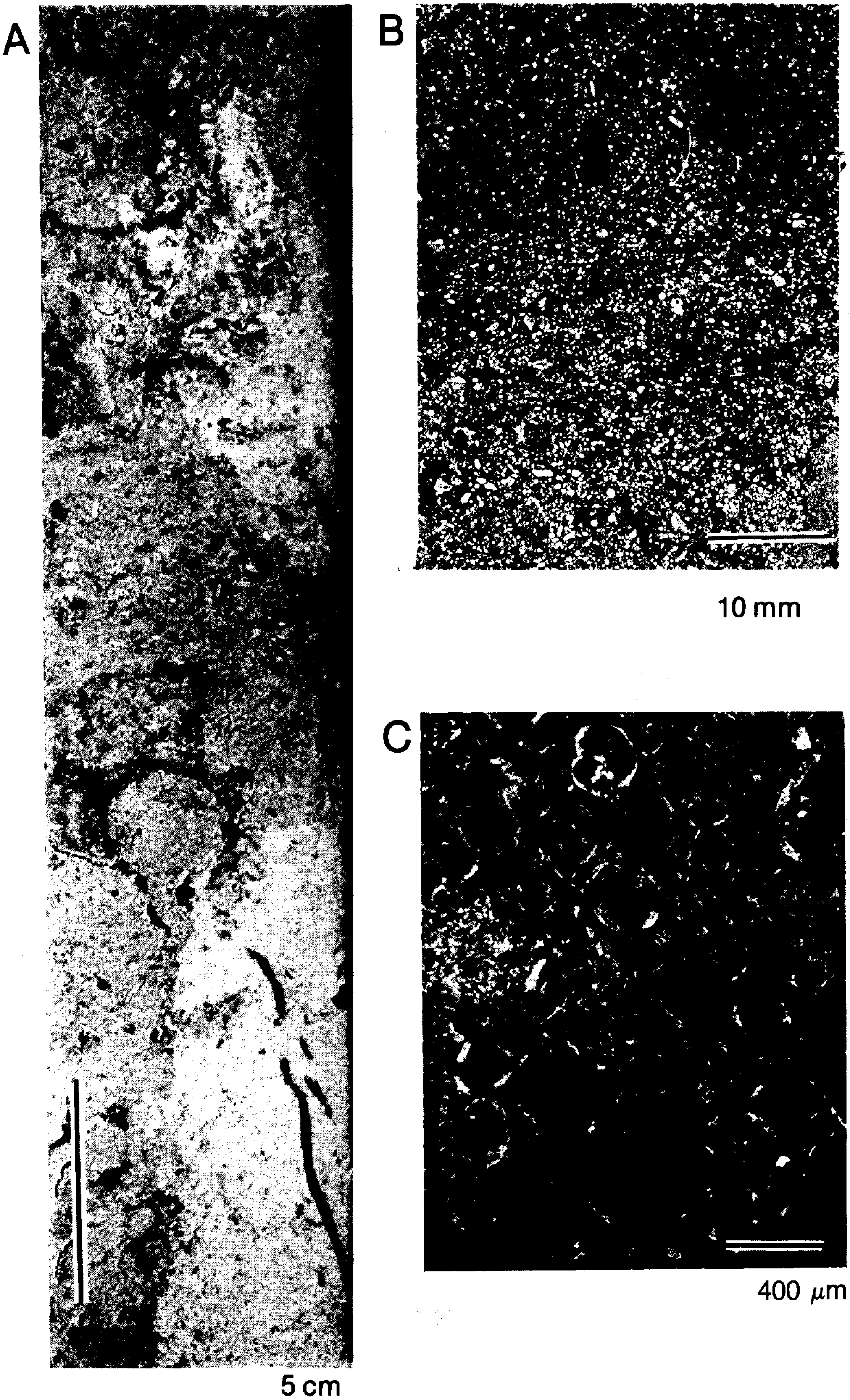


Figure 60.—Ooid packstone shown in core (A), entire thin section (B), and close-up of thin section (C) (from Harris 1979).

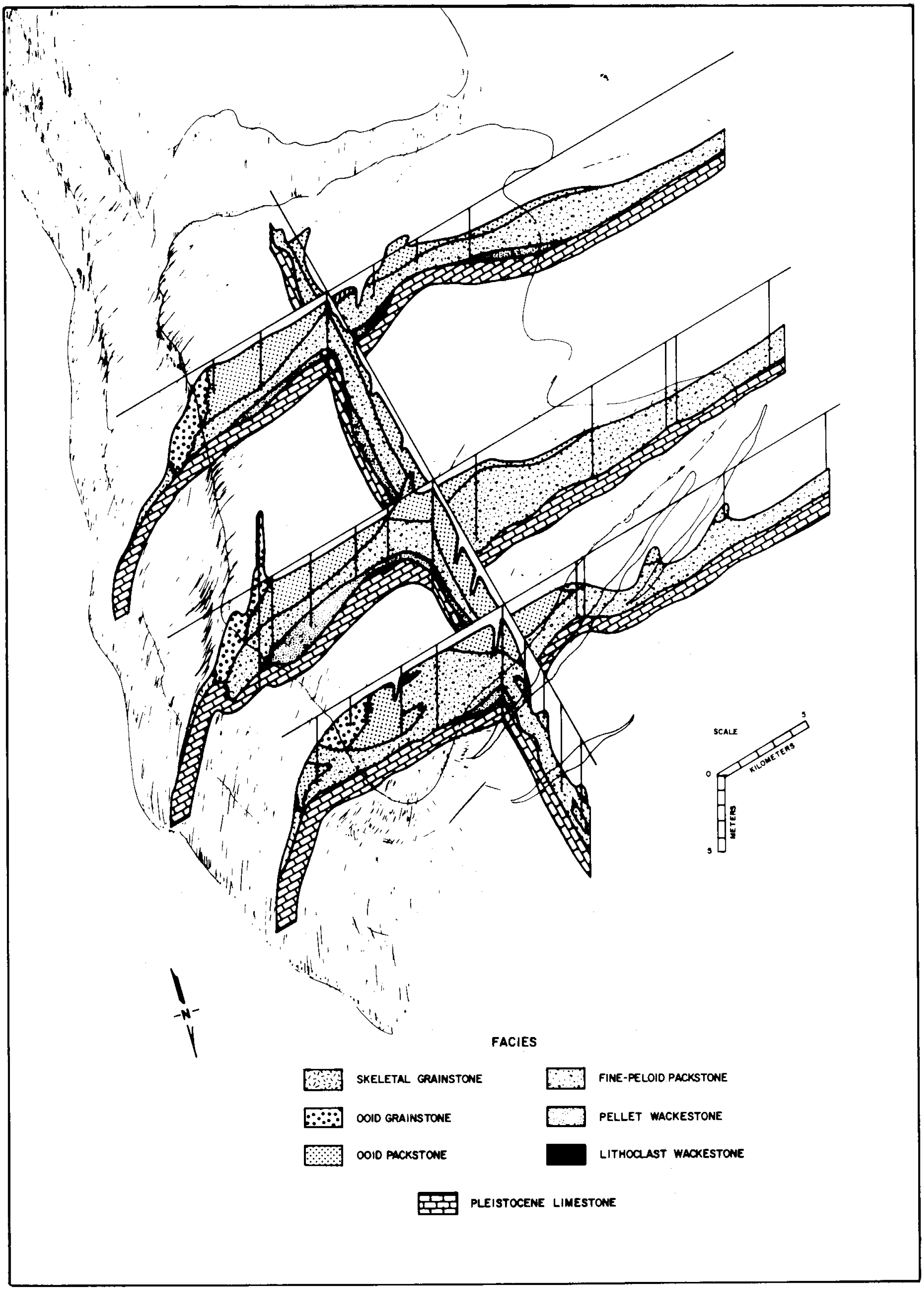


Figure 61.—Fence diagram showing facies relations documented by coring of Joulters ooid shoal (from Harris 1983).

The Joulters sand shoal has a characteristic vertical succession of scattered lithoclast packstones and pellet wackestone at the base, fine-peloid packstone in the middle, and ooid packstone at the top, showing an upward increase in grain size, percentage of ooids, and grain-support fabric. This facies sequence thins to the south over a shallowing limestone surface, as well as to the north and west as overall sediment thickness decreases. Within the shoal, the thicknesses of the dominant facies are complementary; ooid packstone thins in a bankward direction as fine-peloid packstone thickens to form the thickest part of an interplatform sheet. Ooid grainstone directly overlies limestone along the seaward margin of the shoal and interfingers with the packstones bankward.

The interpreted growth through time of the Joulters ooid shoal suggests that the facies changes resulted from changing depositional patterns in response to rising sea level. The shoal grew in three stages: (1) an early bank-flooding stage during which muddy sands of fine-peloids and pellets accumulated in protected lows on the Pleistocene limestone floor (figure 62), (2) a shoal-forming stage during which ooid production began on bedrock highs where bottom agitation was focused (figure 62), and (3) shoal development in which the production and dispersal of ooid sands established the present size and physiography of the shoal. This changed the general nature of bank-margin sediments in the area from muddy peloidal sands to ooid sands (figure 63).

During stage 3, ooid sands were transported farther bankward as a belt of active bars broadened. Eventually the exchange of water between the seaward and bankward sides of the shoal was increasingly restricted by three mechanisms: widespread sediment buildup approaching sea level, restriction of tidal channel flow, and island formation along the shoal's ocean-facing margin. The series of bars and channels became an intertidal sand flat where the sediments are a mixture of burrowed ooid and peloid sands. These overlie and grade bankward into the muddier sands of the bank interior.

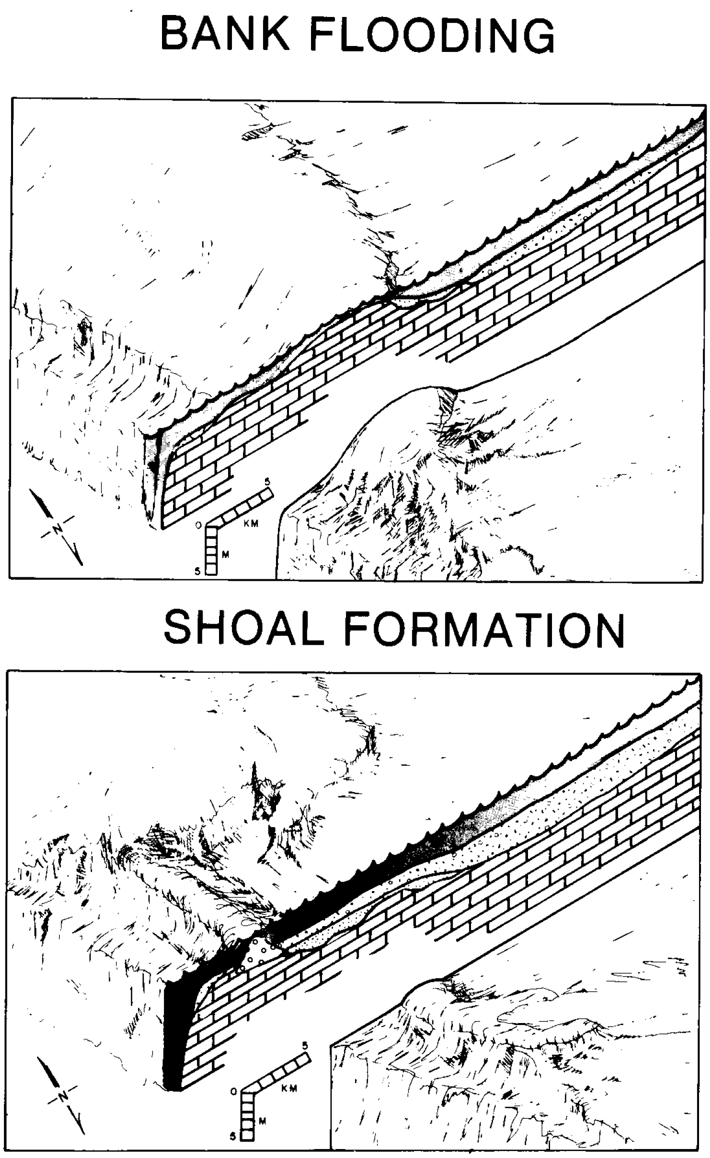


Figure 62.—Bank flooding and stages of shoal formation in growth of Joulters ooid shoal (from Harris 1983).

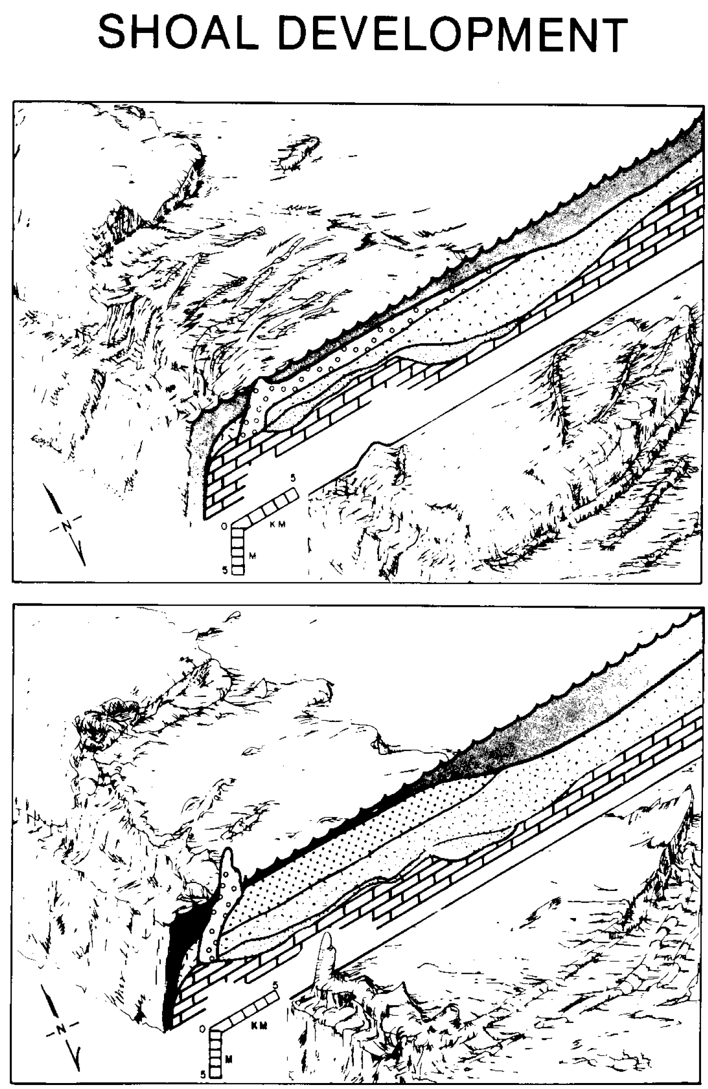


Figure 63.—Growth stages of Joulters ooid shoal (from Harris 1983).

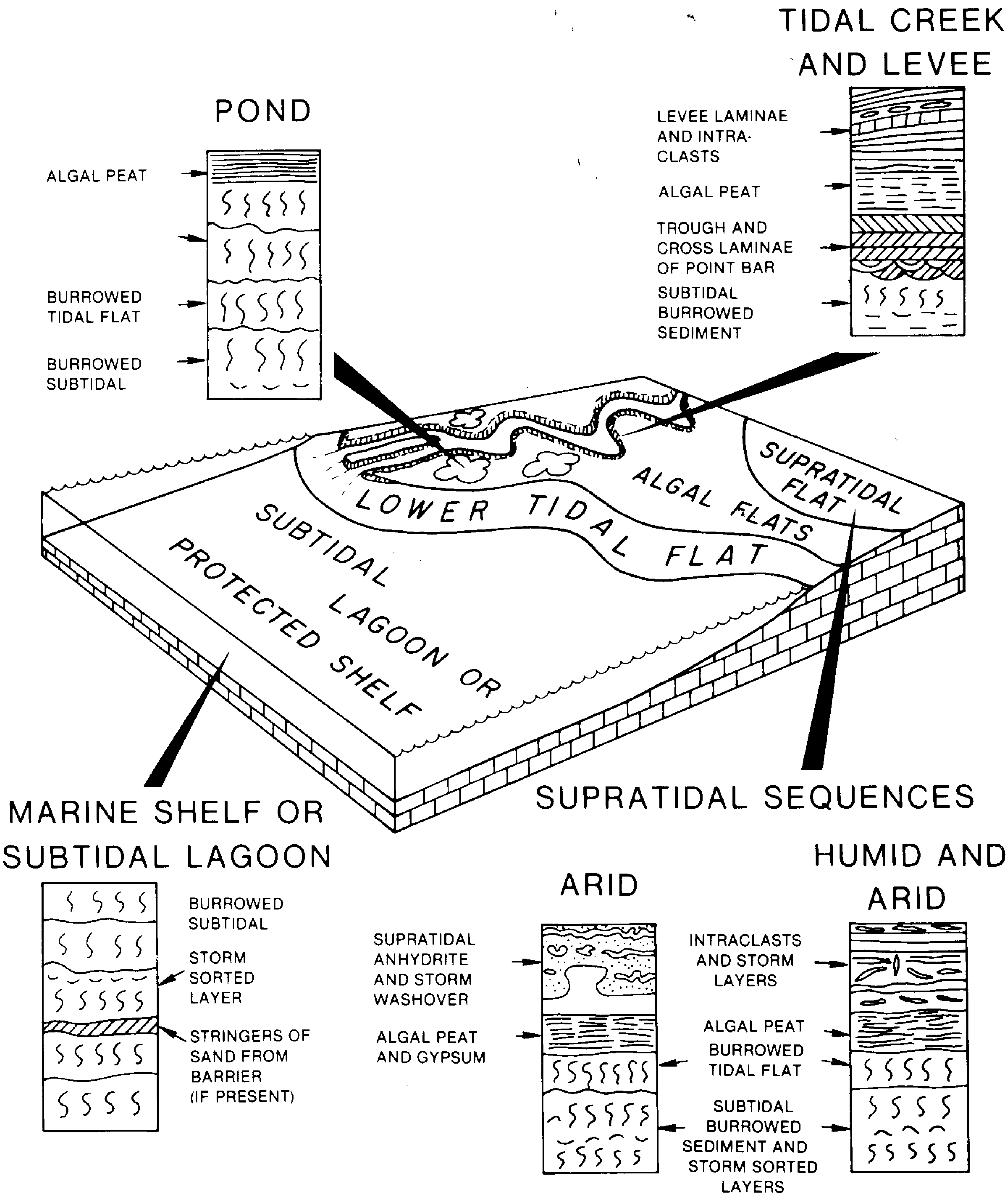


Figure 64.—Generalized view of lagoon and tidal flat environments and facies.

LAGOONS AND TIDAL FLATS

Lagoons protected by a wide, shallow sea, by reefs, or by mobile carbonate sand barriers are characterized by continuous, wide sheets of poorly sorted sediments that are commonly extensively burrowed (figure 64). The sediments either formed in situ or were transported from a seaward barrier by the winnowing action of waves and currents.

In normal marine settings, faunal remains are abundant but not diverse. However, the setting may be several tens of kilometers from the open sea. As in epeiric seas, elevated salinity causes faunas to steadily decrease in species diversity. At the landward margins of such lagoons, where salinities are frequently at their highest, the only biota may be subtidal blue-green algal stromatolite heads and mats.

The principal facies in shallow-water lagoons are clean carbonate sands, muddy skeletal sands, and lime muds. In deeper water, marls and shales are common. The sands in shallow water form on stable flats where current energy is sufficient to winnow lime mud but not grains. Grains may be oolitic, but are more commonly pellets, grapestones, or oncolites.

Lime muds form in areas with restricted circulation. Along the landward margins of lagoons or epeiric seas, the muddy sediments are usually dolomitic and stromatolitic, well bedded, and widely distributed laterally. In subtidal areas, faunal abundance is low, but diversity is high. In intertidal areas, the opposite can be true.

Tidal-flat sediments include those in the intertidal zone (flooded by daily tides) and the supratidal zone (flooded by wind and spring tides) (figure 64). Sediments range from carbonate sands to muds and commonly contain algal stromatolites. Tidal-flat sediments occur as widespread sheets that are often dissected by channels. Bedding is thin and even, and contacts are sharp. Evaporites, however, show irregular bedding and may be nodular. Collapse breccias of angular fragments tend to parallel depositional strike and are local.

Traced landward, the principal facies belts are sandy tidal flats, muddy tidal flats, mangrove/algal flats, and supratidal flats. The sandy flats, commonly cross-bedded and winnowed, reflect storm wave and current movement. The muddy tidal flats are burrowed and homogenized; bedding planes may be irregular, in part a result of the burrowing. Mangroves, algae, and other plants bind and trap transported lime mud and may also aid its precipitation. Algae produce a variety of structures in response to desiccation and erosion; the most significant are lamellar fenestral fabrics and dome heads.

Tidal creeks rework the tidal-flat sediments, developing sandy or muddy point bars. The channel margins may be marked by levees. If these are made of lime mud, they generally exhibit a variety of laminations, mud cracks, and intraclasts. The levees may pond water in the overbank areas, where sediments are highly burrowed.

Supratidal-flat sediments vary according to their climatic settings. For instance, with high salinities and high magnesium concentrations, dolomite cement replaces calcium carbonate. In arid regions gypsum and anhydrite can precipitate directly within the sediment. Simultaneously halite may precipitate locally on the sediment surface, but is usually removed by wind or marine flooding.

Tidal-flat muds and pelleted sands have low porosities due to dewatering and compaction. However, dolomitization of these deposits forms reservoirs by creating good porosity and permeability. Tidal-flat carbonates are commonly associated with evaporites that act as seals to the reservoirs. Examples of production from tidal-flat sequences include the Ordovician Ellenburger Formation, the Ordovician Red River of the Williston Basin, the Permian Basin carbonates of Texas, and the Cretaceous offshore of West Africa.

Tidal-flat carbonates have abundant algal organic matter mixed into them. There may be ample opportunity for the organic matter to be oxidized and come in contact with fresh waters. However, some tidal-flat sequences were evidently deposited quickly enough to retain a relatively high percentage of organic matter. In addition, there is a growing belief that evaporites can have sufficient organic content to serve as hydrocarbon sources.

Florida-Bahama Examples

Florida Bay. Florida Bay, a triangular-shaped lagoon or restricted shelf between the southern end of the Florida Peninsula and the Upper Florida Keys, has long been a focus of geological study because it illustrates a broad spectrum of mud deposition (Ginsburg 1972). The bay is segmented into numerous depressions or "lakes" by long, linear mud banks and small islands (figure 65).

The mud banks are ridge-like, flat-topped depositional buildups that are nearly exposed at low tides. The sediments are commonly burrowed and support a covering of sea grasses and algae. Bare mud patches, typically scattered on the leeward sides of the banks, include both depressions (blowouts) and low-relief mounds. Winnowed lags of mollusk shell fragments form on the windward sides of the banks. Islands are scattered along the mud banks.

Supratidal deposition on the islands includes trapping and binding of sediment by blue-green algae during flooding by winter storms or hurricanes. The islands are also colonized by peat-forming mangroves and hardwoods, and coarse-grained storm ridges form around portions of the island perimeters (figure 66). Algal stromatolites, mud cracks, fenestral porosity, and other distinctive structures characterize the supratidal flats. The storm ridges and beaches of mollusk and foram sands are best developed on the windward side of the island, whereas fringes of peat-forming mangroves typify the leeward side.

The eastern half of Florida Bay is largely subcircular to irregular-shaped lakes; much less area is covered by mud banks and islands. The longer history of deposition in the western portions has meant that mud banks have grown at the expense of the lakes. As a result, the lakes are not significant to the west. The lake bottom varies from bare Pleistocene limestone with scattered rock-encrusting organisms to a thin veneer of coarse gravel of mollusk fragments and lithoclasts, with a varying amount of sea-grass cover.

Although a varied flora and fauna is encountered in Florida

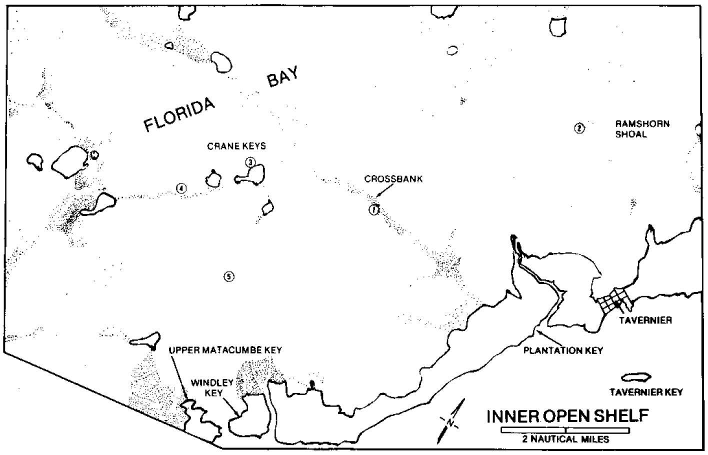


Figure 65.—Map of Florida Bay showing mudbanks (1, 2, and 4), island (3), and "lake" bottom (5).

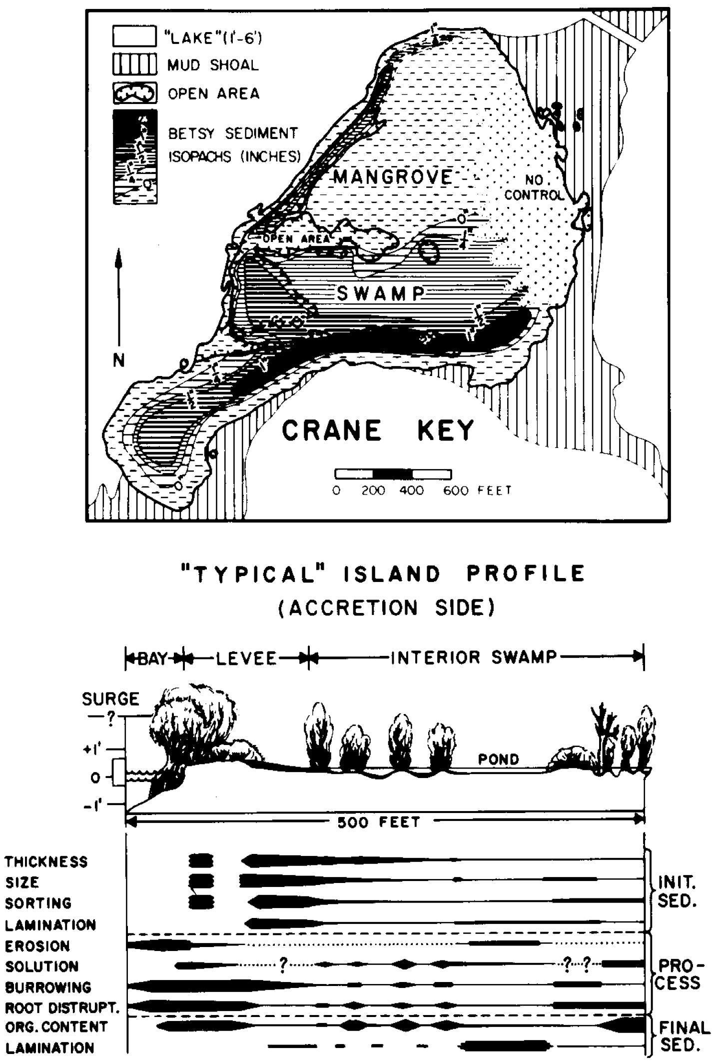


Figure 66.—Detailed map and cross section of Crane Key (see figure 65) (from Pray 1966). Sediment is deposited on the islands during storms such as Hurricane Betsy.

Bay, a few organisms are especially important geologically and deserve special attention.

Thallassia testudinum, or turtle grass, covers the mud banks and parts of the lake floor in Florida Bay, and also occurs on the open shelf seaward of the Keys. The grass is important geologically in several ways: its blades serve as a baffle to trap fine-grained sediment from suspension and are coated with mucous-bound sediment and numerous attached organisms that become part of the sediment. A dense network of rhizomes and roots stabilizes the muddy bottom and at the same time penetrates the sediment to destroy physical layering.

Calcareous green algae, primarily of the family Codiaceae, are important sediment producers in Florida Bay and on the open shelf seaward of the Florida Keys. Heavily calcified forms such as *Halimeda* contribute their whole or broken segments as sand-size sediment, and the lightly calcified forms such as *Penicillus*, *Rhipocephalus*, and *Udotea* (resembling shaving brushes or small fans) disintegrate to produce mud-sized debris.

Islands in Florida Bay are formed by colonization of mud banks by mangrove trees. The characteristic prop-roots of the red mangrove *Rhizophora mangle* form a dense network that baffles the current during storms to catch fine-grained sediment and prevent erosion. Both red and black mangroves are land-formers, stabilizing carbonate mud and forming peat.

The marshes or swamps in the interior of islands like Crane Key are covered with mats of blue-green algae. The algae trap and bind storm deposits and in the process create distinctly laminated, often stromatolitic, structures. The mat dries and cracks into irregular polygons a few inches to a foot across. Filamentous blue-green algae also stabilize the subtidal sediment surface and form oncolites.

Radiocarbon age determinations have been made on buried mangrove peats just overlying the Pleistocene rock floor in cores taken on islands and banks. Results indicate that sea level was 3 to 4 m below present level approximately 4000 years B.P. The mud banks and islands of Florida Bay formed as sea level rose to its present position.

Sediment cores from Cross Bank reveal a mixture of pelleted lime mud and skeletal debris, predominantly mollusks and foraminifera. The sediment is burrowed by crustaceans and worms and penetrated by vertical grass roots. Irregular-shaped concentrations of shell debris and *Halimeda* plates are filled burrows. Cross sections through Cross Bank (figure 67) illustrate that most of the buildup is burrowed wackestone, but that vertical and lateral lithological changes to mudstone and packstone are numerous.

The typical sedimentary record of a Florida Bay island consists of a flooding event overlain by a shallowing-upward sequence (Enos and Perkins 1979). Such sequences reflect bank and island sedimentation rates in excess of the rate of sea-level rise. In ascending order, the sequence as shown on figure 68 is (1) white calcite mud, deposited in freshwater ponds; (2) mangrove peat, recording the encroachment of salt water to produce mangrove swamps; (3) gray, very shelly, open marine lime packstone, representing the maximum transgression; (4) gray lime wackestone with scattered shells from a more restricted environment, representing mud-bank development; and (5) tan or cream, laminated or crumbly lime wackestone

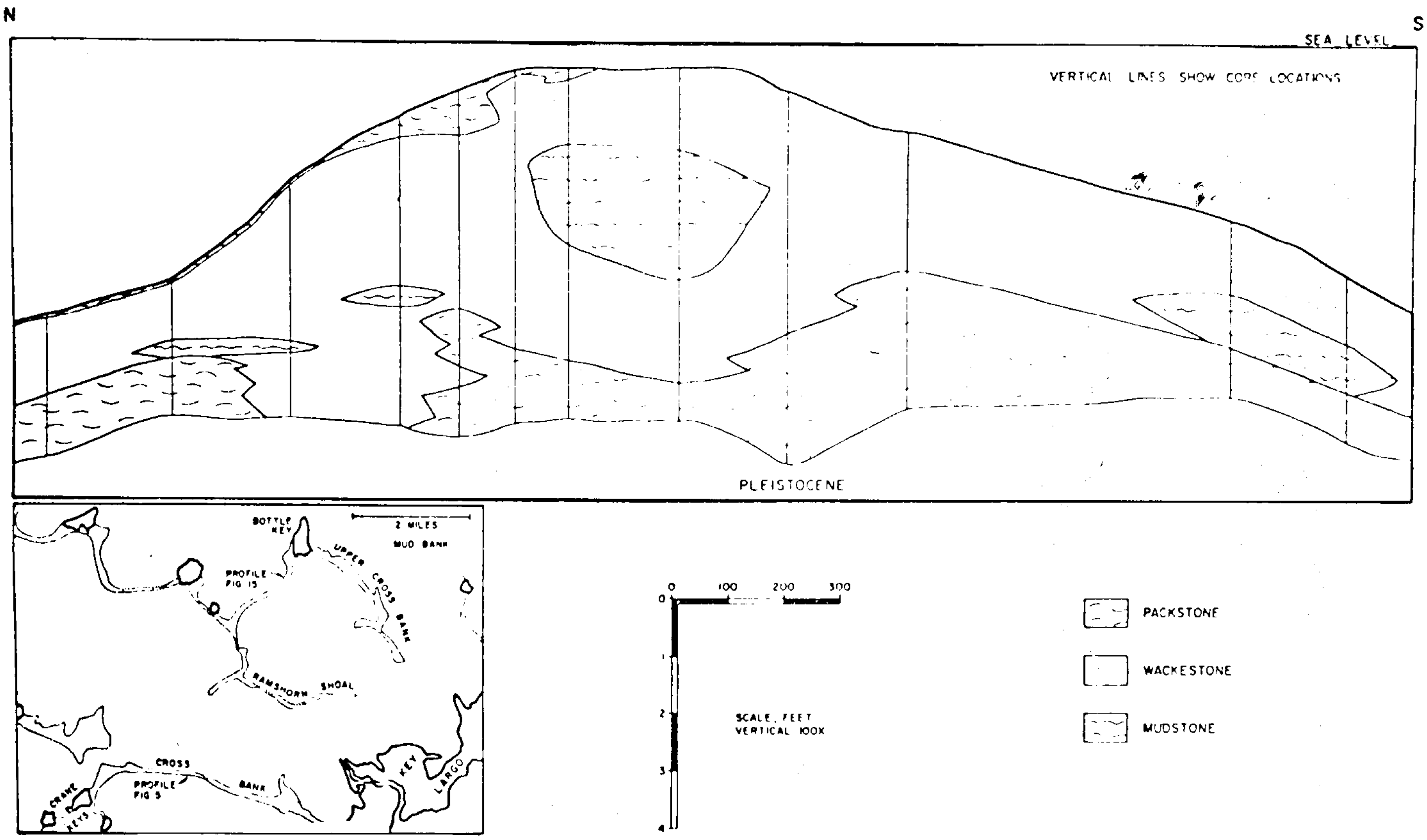


Figure 67.—Dip cross section through Cross Bank (from Enos and Perkins 1979).

and mudstone with peat lenses, representing island sedimentation in the intertidal and supratidal zones.

A core from the supratidal deposits is distinctly laminated and composed of dark organic laminae of grass fragments or algal mats and lighter sediment laminae of fine sand- and silt-sized skeletal debris. Mud cracks and fenestral voids are present. Relative thicknesses of the individual parts of the entire sequence vary across an island, as a result of both upward and lateral island growth (figure 69).

The shallowing-upward sequence that is formed by the mud banks and islands of Florida Bay has characteristics that are commonly recognized in ancient carbonate-shelf deposits. The sequence relates reservoir-bearing, dolomitized subtidal deposits and overlying intertidal-supratidal deposits with sealing potential. In thick subsurface units with multiple porosity zones, commonly the supratidal caps of separate cycles effectively seal and isolate the reservoirs.

Andros tidal flats. The tidal flats of Andros Island provide an ideal opportunity to examine sedimentary structures and facies relationships in peritidal sequences. Such deposits are significant exploration targets where porosity has developed through dolomitization or freshwater leaching. This results in the interfingering of potential reservoir, seal, and source facies.

Two parallel facies belts characterize the Andros tidal flats (figure 70). One is a channeled belt with tidal channels cutting perpendicular to the strike of the tidal flat, channel levees, ponds localized between adjacent levees, and beach ridges fronting the belt and separating it from the offshore marine. The other is an inland algal marsh lying essentially updip of the channels and covered with a stromatolite algal cover.

The bottoms of channels vary within the channeled belt, but commonly have exposed Pleistocene rock in their axes and are mud covered along their flanks. The muds are stabilized by sea grasses or occur as mound-like, algal-coated buildups. The levees and algal marsh are characterized by laminated muds and fine sands, whereas the ponds contain bioturbated, fossiliferous muds. These are only subtly different from burrowed, pelleted muds that occur in the marine environment offshore of the tidal flats. Blocks of algal-coated, cohesive mud are reworked from the levees as they are undercut by the channel. The crests of the levees are mud cracked, as is the mat of blue-green algae on the supratidal algal marsh.

The Holocene sediment record of the west coast of Andros Island is of bioturbated, unlayered, pelleted mud and silt (Shinn and others 1965; Hardie and Ginsburg 1977). Layered sediment exists only as a thin cap over bioturbated sediment in the channeled belt and as thin-bedded inland algal marsh sediment (figure 71).

Homogenization of layered sediment by marine organisms is a major sedimentary process on subtidal and intertidal parts of the tidal flats. Layering is preserved where burrowing organisms are excluded by prolonged exposure above sea level (Ginsburg and others 1977). The crest and backslope of the levees are exposed over 90 percent of the time, the ponded areas only 10 to 60 percent of the time. The duration of expo-

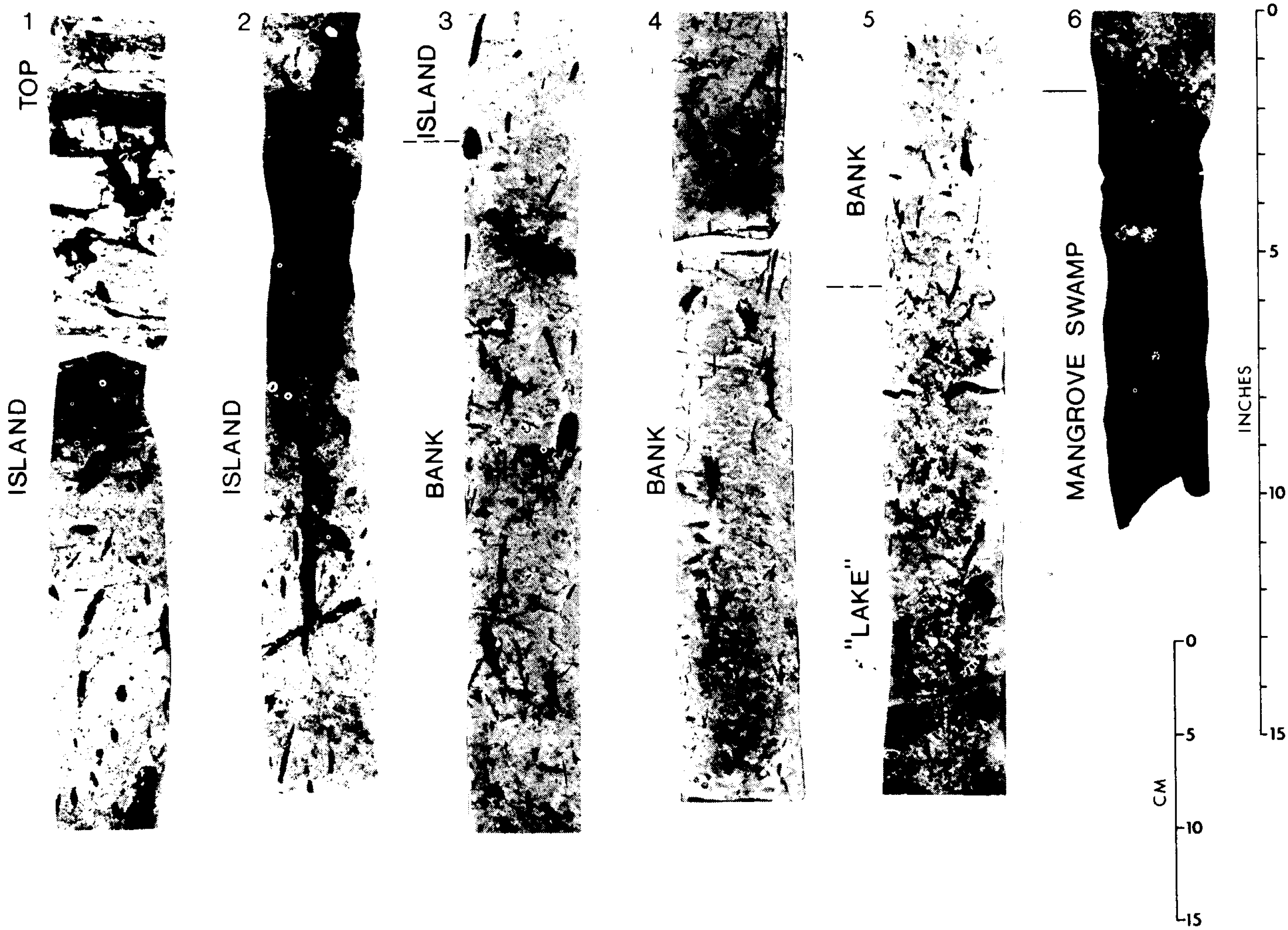


Figure 68.—Predominant facies shown in stratigraphic sequence of continuous cores from Crane Key (from Enos and Perkins 1979). Same facies are shown in cross sections on figure 69.

sure largely controls (1) variations in the algal mats, (2) physical layering of sediments, (3) formation of intraclasts and mud clasts by desiccation, and (4) burrowing by crustaceans, worms, and insects.

The Holocene record of the tidal flats begins with a freshwater marsh accumulation unconformably overlying Pleistocene limestone bedrock (figure 72). Flooding seas significantly eroded the marsh sediments underlying most of the present-day tidal flat, except beneath the inland algal marsh, which has remained a freshwater marsh without interruption. Accumulation was mainly by vertical accretion in a complex of environments (channels, channel bars, levees, ponds, and marshes) behind a protecting barrier beach ridge. The distribution of tidal environments behind the barrier is complex. The basic stratigraphic record consists of a basal unit of bioturbated subtidal and intertidal sediment overlain by a thinner, well-layered unit deposited by the most severe onshore storms.

Sediment from a lower intertidal channel-margin is a bioturbated, pelleted mud with a minor admixture of high-spired gastropod shells and foraminifera tests. The rock equivalent would be wackestone or mudstone. The sediment is penetrated by numerous roots and worm burrows, and has vague, discontinuous laminations.

The sediment from a supratidal levee would also be a wackestone or mudstone, but would contain distinctive structures. Such sediments are of laminated, pelleted mud with numerous cracks and fenestral voids. Scattered roots penetrate the sediment but do not significantly disrupt the laminations. Subtle color changes between laminae reflect variation in the amount of oxidation.

Sediment from the algal marsh is most distinctive. In cores, thin beds of pelleted lime mud and fine skeletal sand alternate with darker, algae-rich layers. The rock equivalent would be wackestone, or sometimes packstone. The sediment layers represent deposition during storms. Sediment is washed and blown onto the algal marsh from the offshore and channeled-belt environments.

SUMMARY

The preceding brief discussion of depositional environments

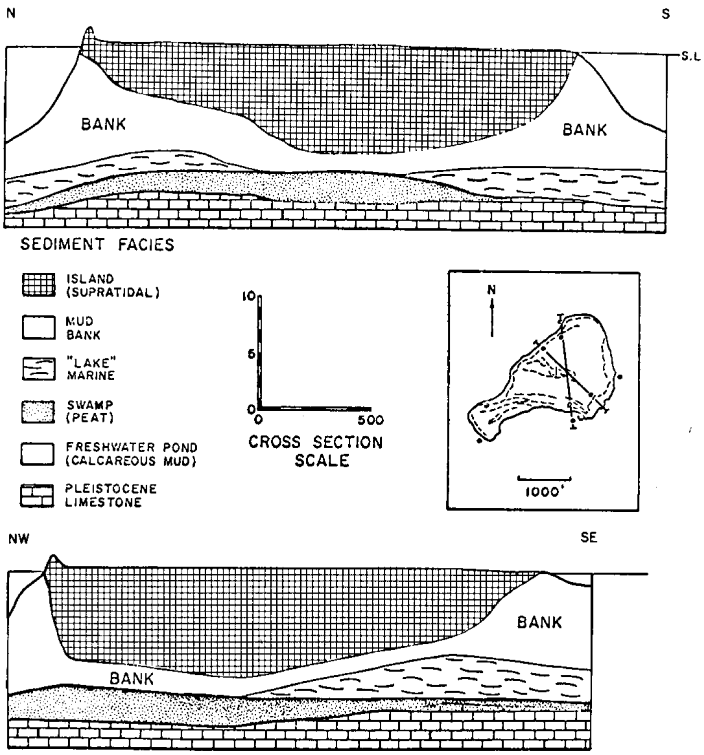


Figure 69.—Intersecting cross sections from Crane Key (from Enos and Perkins 1979).

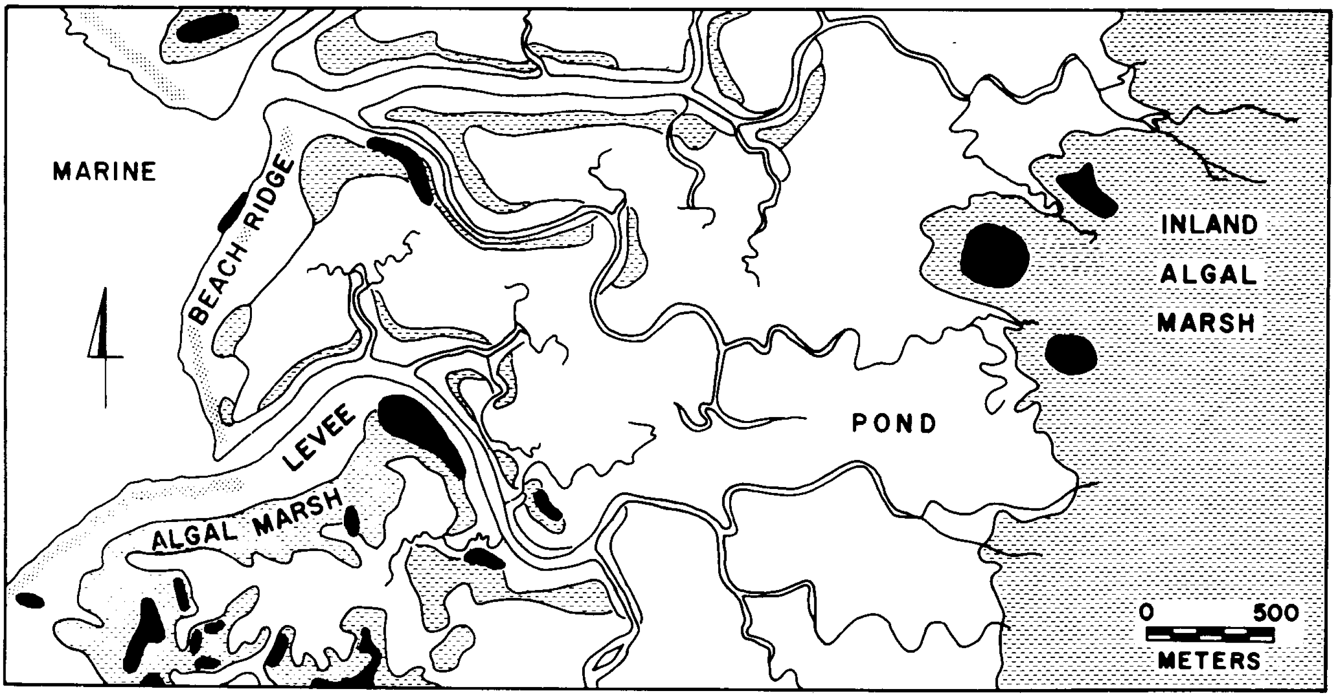


Figure 70.—Major environments on northern portion of Andros tidal flats showing the channeled belt and inland algal marsh (from Hardie and Garrett 1977). Tidal flats occur along the western side of Andros Island (see figure 42).

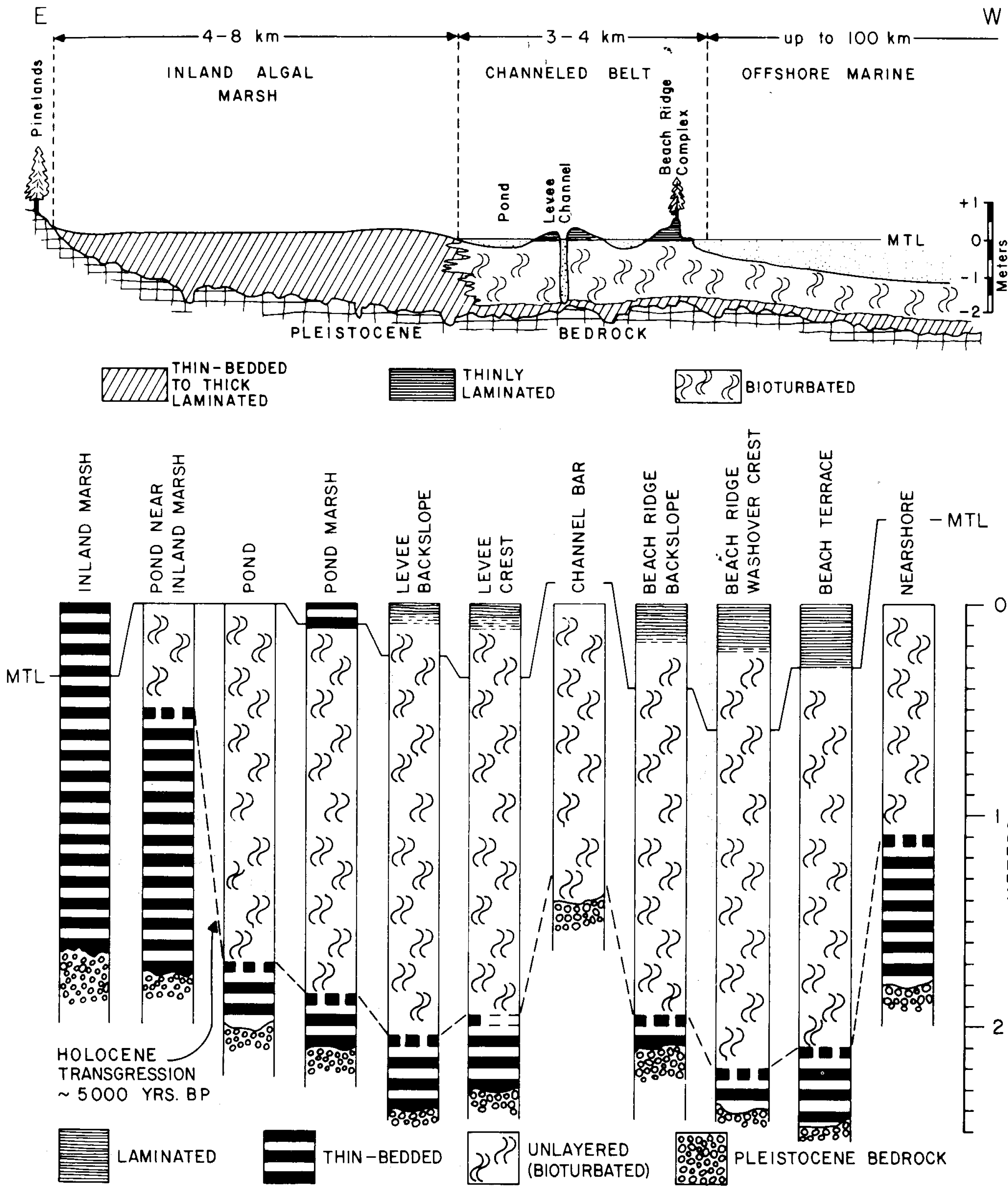


Figure 71.—East-west cross section and details of typical cores from northern portion of Andros tidal flats showing variation in layering (from Hardie and Ginsburg 1977).

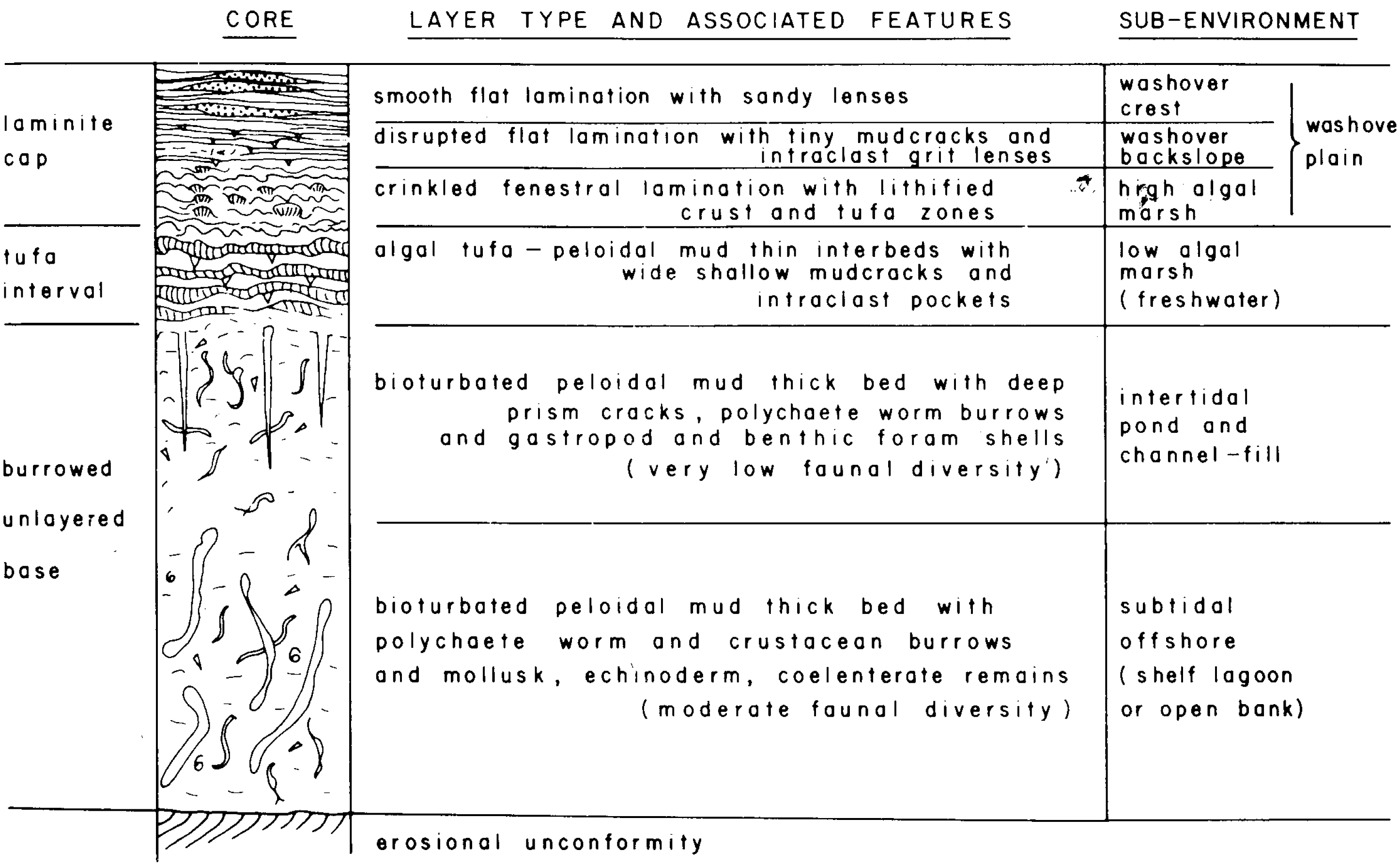


Figure 72.—Shallowing-upward sequence in tidal flat core (from Hardie and Ginsburg 1977).

of carbonate platforms outlines the general patterns of sedimentation and simplified vertical sequences in modern reefs, buildups, sand shoals, lagoons, and tidal flats. The specific localities, chosen from the Florida - Bahamas area, are well-documented sites of carbonate deposition. Over the years, the studies have proven to be valuable analogs for calibrating core studies and formulating depositional models for ancient subsurface examples. Certainly, the areas emphasized here, and the Florida - Bahamas region in general, lack some features that may limit their usefulness as universal analogs, that is, evaporite-dominated coastlines and "pinnacle" reef morphologies. These shortcomings are offset, however, by the area long serving as an accessible field laboratory where the organisms, sediments, structures, sequences, morphologies and processes of the fundamental building blocks of platform carbonates have been scrutinized.

An understanding of carbonate facies, depositional systems and models, based on the studies of modern environments, is critical when attempting to unravel ancient counterparts. It is also equally important to understand the diagenesis that has affected the sediments; there are numerous instances where the diagenetic overprint is equal to or more important than the depositional facies patterns in controlling the final porosity distribution in an area. The cementation and solution that occur during the earliest stages of diagenesis have not been emphasized for the various environments discussed here. That topic will be discussed in a subsequent part of the CSM carbonate series. Hopefully, the brief discussion of carbonate environments presented here will serve as a starting point in our understanding of platform carbonates and especially subsurface examples where the stratigraphic sequence provides the basic data.

Acknowledgments

The text draws heavily from seminar and field-course notes prepared for Gulf Research and Development Company. A more thorough introduction to carbonates is in preparation with co-authors C.G. Kendall and S.H. Frost.

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