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JOACHIM REITNER
VOLKER THIEL
University of Göttingen
Germany

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Cross-references

Aerobic Metabolism
Anaerobic Transformation Processes, Microbiology
Biomarkers (Organic, Compound-Specific Isotopes)
Black Shales
Carbon (Organic, Cycling)
Deep Biosphere of Sediments
Extracellular Polymeric Substances (EPS)
Methane, Origin
Microbial Degradation
Nitrogen
Shales
Soils
Sulfate-Reducing Bacteria

CARBON CYCLE

The carbon cycle is a biogeochemical cycle that describes the movement of carbon between its different compartments (reservoirs) on the Earth, namely biosphere, atmosphere, hydrosphere, pedosphere, and geosphere. Please refer to entries "*Carbon (Organic, Cycling)*" and "*Carbon (Organic, Degradation)*."

CARBON ISOTOPES

Please refer to "*Isotopes and Geobiology*" and *"Biomarkers (Organic, Compound-Specific Isotopes)*."

CARBONATE ENVIRONMENTS

Eberhard Gischler
J.W. Goethe-Universität, Frankfurt am Main, Germany

Definition

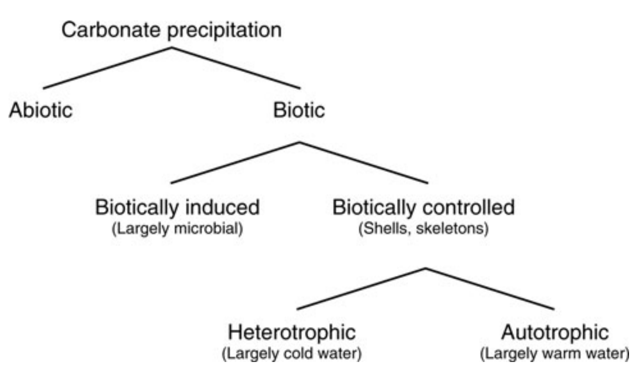
Carbonate environments occur both in the terrestrial and marine realms as well as in transitional zones between the land and the sea. The variety of environments spans from high-elevation continental lakes to the deep sea, from the equator to latitudes of about 60°, and they include dry and wet climate realms. The majority of carbonates are produced in the sea, currently in more or less similar proportions in neritic and pelagic settings. Carbonate production is to a large part a consequence of biological activity, either directly as in shell and skeleton formation or indirectly as a result of metabolic reactions, which trigger precipitation. The highest rates of carbonate production per unit time and space are probably reached in tropical coral reefs where several kilograms of calcium carbonate form on 1 m^2 during 1 year's time.

Introduction

Modern carbonate sediments are largely a product of biologic activity, either enzymatically controlled like in calcium carbonate ($CaCO_3$) shell and skeleton formation of aquatic organisms or indirectly by the metabolic activity of organisms, which change water chemistry so that carbonate precipitation is enhanced (Milliman, 1974; Scholle et al., 1983; Scoffin, 1987; Tucker and Wright, 1990; Flügel, 2004; Schlager, 2005). The former refers to common carbonate producers such as corals, mollusks, calcareous algae, brachiopods, bryozoa, echinoderms, sponges, foraminifera, or coccolithophorids. The latter includes microbes and algae, which are of great importance in the formation of microbialites (stromatolites and thrombolites), tufa, in the initiation of carbonate precipitation, e.g., by removing carbon dioxide (CO_2) during photosynthesis, or as photosymbionts in reef corals, larger benthic foraminifera, giant clams, and sponges (Figure 1).

Possible exceptions of biologic production of carbonate include some carbonate muds and non-skeletal carbonate particles such as ooids, peloids, and aggregate grains that are currently forming in the Bahama Banks, the Persian Gulf, restricted platforms, or certain lakes. Peloids include both cemented fecal pellets as well as micritized or recrystallized skeletal grains (Purdy, 1968). Recrystallization is a common pattern in carbonate sediments and rocks, which is a geologically fast process (Bathurst, 1971; Reid and Macintyre, 1998). Carbonate cements are another common phenomenon in which biogenic *versus* abiogenic precipitation is discussed (Schneidermann and Harris, 1985; Schlager, 2005). Classic examples of relatively fast and abundant carbonate cementation include tropical coral reef frameworks (Grammer et al., 1993) and beachrock (Gischler, 2007).

The majority of carbonate sediments by far are produced in the sea. Presently, about half of this carbonate originates in the pelagic realm and largely includes shells of planktonic and benthic foraminifera, pteropod mollusks, and coccolithophorids. The other half is produced in the neritic realm by corals, mollusks, calcareous algae, brachiopods, bryozoa, and echinoderms. It is estimated that this 50:50 ratio between pelagic and neritic carbonate sedimentation changed to 90:10 during Quaternary lowstands of sea level (Milliman and Droxler, 1996).



Carbonate Environments, Figure 1 Carbonate formation in terrestrial, transitional, and marine environments. (Modified from Schlager 2005.)

Today, carbonate minerals commonly formed in the sea are aragonite and high-magnesium calcite (calcite with 4–21% $MgCO_3$). After sodium (Na^+), magnesium (Mg^{2+}) (1,300 ppm) and calcium (Ca^{2+}) (410 ppm) are the second and third most common cations in the ocean, respectively. Carbonate as HCO_3^- is the third most common anion (140 ppm) after chloride (Cl^-) and sulphate (SO_4^{2-}). Calcium carbonate is precipitated after the formula:

$$Ca^{2+} + 2HCO_3^- \longrightarrow CaCO_3 + H_2O + CO_2 \tag{1}$$

In freshwater environments, for comparison, where ion concentrations in general and the magnesium/calcium ratio in particular are much lower, low-magnesium calcite typically is precipitated. Subaerial exposure surfaces of limestone are characterized by dissolution, and form typical karst features.

Dolomite ($MgCa(CO_3)_2$) may form in limited quantity both in the ocean and in lakes. Preconditions for dolomite precipitation are a high Mg/Ca-ratio and elevated salinity. Modern examples of dolomite formation include the deep sea, certain lakes, and sabkhas, which are intertidal carbonate areas in an arid climate. The majority of dolomite presently exposed on the earth's surface, however, was formed secondarily in the burial environment, by dolomitization of carbonate rocks from magnesium-rich solutions.

The analysis of non-skeletal grains such as ooids and cements throughout the marine geologic record has revealed that the mineralogy of precipitated carbonate has changed between aragonite/high-magnesium calcite and low-magnesium calcite (Sandberg, 1983). This observation, known as the Sandberg-Cycles, is a consequence of changing Mg/Ca-ratios in the ocean as controlled by plate tectonic activity. When lower spreading-rates, colder climate, and lower sea levels are experienced during the periods such as today's or the late Paleozoic icehouse worlds, aragonite predominates over low-magnesium calcite (high Mg/Ca). When spreading-rates and sea level were high and the climate was warm, such as during Devonian or Cretaceous greenhouse worlds, low-magnesium calcite precipitation was favored (low Mg/Ca). Meanwhile, additional fossil and aquarium studies have shown that quite a number of carbonate-producing organisms also change their skeleton or shell mineralogy according to Mg/Ca-ratios of seawater (Stanley and Hardie, 1998). Investigations in Precambrian carbonate rocks have supported the contention that carbonate concentrations in the ocean also fluctuated during Earth history, thereby influencing carbonate deposition (Knoll et al., 1993). In the Proterozoic, carbonate concentrations were so high that in places cements were physicochemically precipitated on the seafloor. Carbonate concentrations decreased in the Cambrian when shell and skeleton production exploded. Carbonate concentrations presumably sank again in the Mesozoic when calcareous plankton such as planctonic foraminifera and coccolithophorids evolved. The Soda Ocean model of Kempe and Degens (1985) speculates that calcium concentrations in the ocean also dramatically decreased with shell and skeleton evolution during the end of the Precambrian through the Cambrian. These authors interpreted shell and skeleton formation as calcium expelling as a consequence of cell poisoning.

In the following, highlights of typical carbonate environments are discussed in a cross section going from the terrestrial, coast, and shallow marine to deep marine settings (Figure 2).

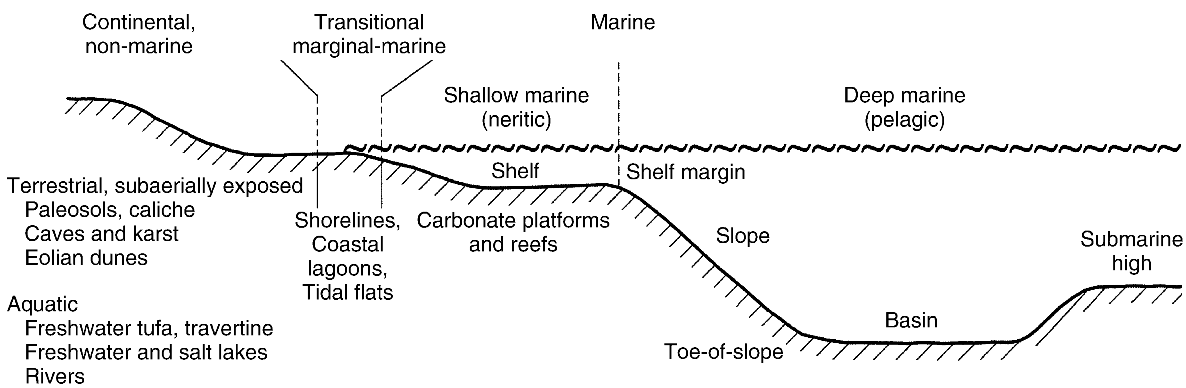
Terrestrial environments

Springs

Carbonate deposits are normally very rare in running freshwater because of the low ion contents of the water. Turbulence that causes CO_2 degassing may lead to low-magnesium calcite precipitation. Also, pressure release or cooling of hot waters at springs may bring about calcite precipitation. The limestone product is usually termed tufa, which forms under ambient temperatures and may contain calcified remains of bacteria, algae, and plants (Chafetz and Folk, 1984; Ford and Pedley, 1996). Travertine, which forms in hot water (hydrothermally), is dense, usually clearly laminated, and largely lacks algae and plant remains. Typical morphologies of tufa deposits include barrages, cascades, or terraces with pools. Ooids may form in pools that hold water. Tufa deposits are restricted to the late Cenozoic, which presumably is a consequence of limited preservation potential (Figure 3).

Lakes

Lacustrine carbonates occur in both freshwater and saline lake environments (Matter and Tucker, 1978; Hakanson



Carbonate Environments, Figure 2 Carbonate environments in a land-to-sea cross section. (Modified from Flügel, 2004.)

Carbonate Environments, Figure 3 Photo of sliced Quaternary tufa deposit on Casino building of Goethe University, Frankfurt, Germany, with microbial lamination and calcified remains of algal thalli.

and Jansson, 1983; Anadon et al., 1991). Freshwater lakes in humid climates such as Lake Constance, Germany, Lake Zurich, Switzerland, or Green Lake in North America are usually perennial. Their size can be as small as <10 km^2 as in Green Lake to more than 500 km^2 as in the example of Lake Constance. They are characterized by fine-grained carbonate deposits, usually in an annually laminated fashion (Figure 4). Fine-grained siliciclastics are more common during winter. Low-magnesium calcite is precipitated preferentially during warm summer months, enhanced by photosynthetic activity of phytoplankton that withdraws CO_2. Other carbonate producers are usually ostracods, gastropods, pelecypods, and the charophytic alga *Chara*. The thalli of this alga may calcify and form characteristic tubes. Reproductive cells (oogonia) are calcified also. Diatoms occur as well. In the case of high carbonate content of the lake water, microbial activity may occur and produce stromatolites, thrombolites, and/or oncolites (Figure 5).

Carbonate Environments, Figure 4 Photo of annually layered Holocene lake sediment from core, Lake Zurich, Switzerland. Light layer is carbonate; dark layer is siliciclastics. Thicker gray bed is a turbidite. Scale in centimeters. (Photo courtesy of Flavio Anselmetti.)

Saline lakes in arid climates such as the Great Salt Lake in North America or the Dead Sea have considerable sizes of 4,400 and 600 km^2, respectively. Saline lakes are often ephemeral. They are also characterized by abundant carbonate deposition. Controlled by the Mg/Ca-ratio, either low-magnesium calcite, high-magnesium calcite, aragonite, magnesite ($MgCO_3$), or dolomite is precipitated. Under high evaporation rates, sulphates and chlorides such as gypsum ($CaSO_4 \cdot 2H_2O$) and halite (NaCl) may form. In the case of the Great Salt Lake, aragonitic ooids occur in abundance in shallow areas along shore. Also, microbial buildups and sheets are found, which are to large parts due to cyanobacterial activity. Along shore of saline lakes, caliche may be common due to fluctuating lake levels that expose larger areas. When springs are present, tufa or travertine is formed as sheets, mounds, and towers. The latter is characteristic of Mono Lake in California where spectacular towers of tufa are found.

Subaerial exposure

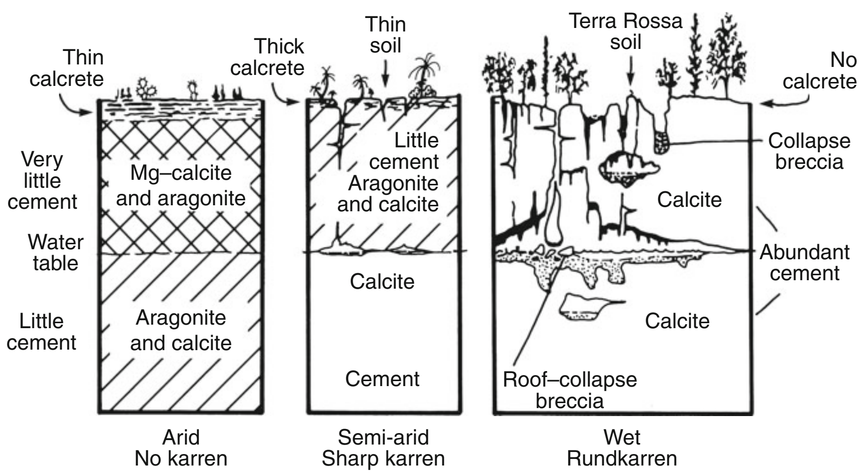
Subaerial exposure of limestone leads to dissolution by meteoric waters. Typical landforms that may have considerable relief are termed karst (James and Choquette, 1988). Landforms include dolines, towers, karren, or poljes. Dissolution may produce cavities and larger caves. Calcium carbonate is re-precipitated in limestone caves in the form of speleothems such as stalactites and stalagmites. Cave pearls or pisoids ("diagenetic ooids") may form in ponds. Residues of dissolution together with organic material may later become soils on limestone such as laterite or terra rossa. In warm and/or arid regions, carbonate-rich soils form deposits that are called caliche or calcrete (Wright and Tucker, 1991). A caliche is characterized by patterns such as brecciation, clotted textures, root holes or rhizoids, and low-magnesium calcite cements, sometimes in a needle fiber texture ("whiskers"). Laminated soilstone crusts are a special case of calcrete that can resemble stromatolitic formations. Soilstone crusts form by precipitation of calcium carbonate from evaporating pore water. The identification of subaerial exposure surfaces in the fossil record is of great importance, e.g., for the reconstruction of fluctuations in sea level and for sequence stratigraphy (Figures 6 and 7).

Transitional environments

Beaches and shores

Carbonate beaches are usually formed by sand-sized particles that originate from the adjacent shallow seabed (Inden and Moore, 1983). Shells and skeletons of organisms as well as non-skeletal grains occur. Beach zones include the shoreface, the foreshore, and the backshore. The shoreface lies below the lower tide level and is characterized by cross-bedding. The foreshore is situated between

Carbonate Environments, Figure 5 Photo of lacustrine stromatolites, Bacalar, Yucatan Peninsula, Mexico.



Carbonate Environments, Figure 6 Scheme of karst profiles in various climates. (From James and Choquette, 1988; with permission of Springer, NY.)

low- and high-tide levels. The backshore lies above the high-tide level. Sediments of the foreshore and backshore zones exhibit parallel lamination and beds dip gently toward the sea (Figure 8).

A striking feature in the tropical/subtropical realm, and common in latitudes up to 60°, is beachrock. Beachrock forms in the intertidal zone by rapid cementation of sedimentary particles by marine aragonite and high-magnesium calcite cements. The formation of beachrock is not fully understood, and is assigned to physicochemical and/or biological processes (Gischler, 2007). The strictly intertidal position of beachrock supports the contention that cements precipitate from evaporating pore water during low tide. Also, beachrock appears to be common in windward and rather exposed sites indicating that flushing of pores by water rich in calcium carbonate is of importance. Typical sediment structures of beachrock are seaward dipping beds. Keystone vugs are small cavities in beaches and beachrock that are produced when air escapes from the beach (Figure 9).

Behind the backshore, in the landward direction, dunes may occur that are a product of eolian deposition (Abegg et al., 2001). Carbonate eolianites are common behind carbonate beaches, and are found in abundance along the eastern coast of the Yucatan Peninsula, on the islands of the Bahamas and Bermuda, along the Trucial Coast (southern Persian Gulf), and on the Balearic Islands in the Mediterranean Sea. High angle cross-bedding as well as rhizoids and caliche, indications of the meteoric realm, are characteristic. *Microcodium* are small spherical bodies

Carbonate Environments, Figure 7 Hand specimen of caliche with laminated crust. Miami Limestone from the Everglades, Florida.

of radiating calcite that forms among rhizoids in eolianites. Carbonate cements include meteoric types composed of low-magnesium calcite.

Tidal flats

As indicated by the name, tidal flats are broad and low-relief areas, which are usually flooded during high tide and subaerially exposed during low tide (Ginsburg, 1975). Classic modern carbonate-dominated examples include the humid tidal flats west of Andros Island on Great Bahama Bank (Shinn et al., 1969) and the arid examples on the southwestern side of the Persian Gulf (Kendall and Skipwith, 1969). Tidal flats are dissected by tidal channels, which are lined by levees. Levees usually enclose low-relief ponds. Beach ridges form the transition to the adjacent subtidal realm. In the case of humid tidal flats, algae such as *Batophora* and microbial mats as well as mangroves thrive along the channels and further inland in the so-called marsh. In the Bahamas, microbial mats are largely formed by the genera *Scytonema* and *Schizothrix*. Gastropods of the genus *Batillaria* occur in great abundance. Benthic foraminifera are present. In arid tidal flats or sabkhas, which transition into the desert on the landward side, microbial mats occur along channels. Also, burrows of crabs may be abundant along the channels. Nebkhas are small mounds that are produced by halophile plants that baffle sediment around their roots (Figure 10).

Sediments on tidal flats are usually poorly sorted muds, silts, and sands. They are horizontally layered and are largely derived during storms from the adjacent seabed, from precipitation, or from eolian transport. Fine-grained sediment is usually found in ponds. Mud cracks and wrinkled, dried-out, and curled microbial mats are a typical phenomenon. Sands make up levees and beach ridges. A characteristic sedimentary structure is the bird's eyes structure, millimeter-sized fenestral pores that result from gas-bubble formation. In the fossil record, these structures are good indicators of the intertidal and shallow subtidal realm. As a consequence of evaporation on tidal flats, cemented carbonate crusts are formed that may contain dolomite. In the arid realm, gypsum and/or anhydrite is precipitated within the sediment and forms nodules, enterolithid folds, discs, and chicken-wire textures (Figure 11).

Marine environments

Carbonate shelves and platforms (Figure 12)

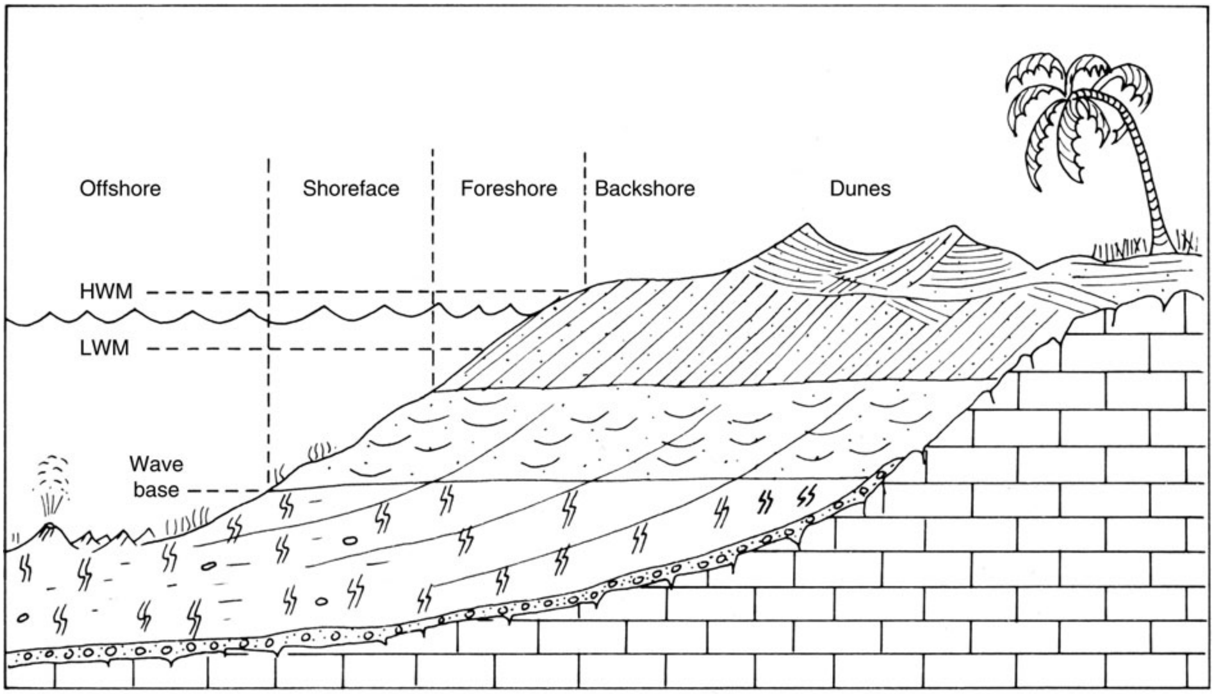
Restricted shelves and platforms

These carbonate environments are shallow (0–20 m) and characterized by partial or complete enclosure by islands, reefs, or shoals. Modern examples include Florida Bay (Ginsburg, 1956; Stockman et al., 1967), the Gulf of Batabano (Hoskins, 1964), Cuba, or Shark Bay, western Australia (Logan et al., 1974). The partial or complete enclosure of these platforms results in restricted circulation and in elevated water temperature and salinity. This in turn leads to low diversity biota of mollusks, benthic foraminifera, crustaceans, and echinoderms. Burrowing crustaceans (*Callianassa*) are responsible for abundant bioturbation (Figure 13). Tolerant corals such as *Solenastrea* and *Siderastrea* are occasionally found in Florida Bay. Sea grasses (*Thalassia*, *Syringodium*, *Posidonia*, and *Cymodocea*) are common, which act as bafflers of sediment. In Florida Bay, the codiacean algae *Penicillus* and *Halimeda* occur in great abundance. Dasycladacean algae such as *Acetabularia* are common. In Shark Bay, stromatolites thrive in intertidal and subtidal areas. Their abundance is classically explained by the exclusion of grazers due to the high salinities in the bay.

Sediments in Florida Bay are largely of skeletal origin. Dominant carbonate producers are mollusks, foraminifera, and the alga *Penicillus*, which is the most important source of aragonite mud (Stockman et al., 1967). Mud banks, elongated hillock-type accumulations of mud, are densely covered by sea grass (Bosence, 1995). Both skeletal (mollusk, foraminifera) and non-skeletal (peloidal) grain types are found in the Gulf of Batabano. Likewise, Shark Bay sediments include skeletal grains of mollusks and foraminifera as well as mud from the breakdown of *Penicillus*. Non-skeletal grains include ooids that are formed in shallow areas, peloids, and reworked lithoclasts from underlying Pleistocene limestone (Figures 14–16).

Open-rimmed shelves and platforms

Intensively studied examples of rimmed carbonate shelves and platforms include the Florida Reef Tract, the Belize shelf and barrier reef, Central America, and the Queensland shelf and Great Barrier Reef, NE Australia. These environments are largely characterized by normal marine



Zone	Grain size coarse fine	Sorting poor well	Lithology	Core	Sedimentary structures	Processes
Island dunes			Grainstone		Parallel laminations high angle cross bedding with rhizoids and caliche	Wind blown
Foreshore			Grainstone		Parallel laminations small scale avalanche cross bedding fine graded laminations vertical burrows	Wave swash
Shoreface			Grainstone to Packstone		Small to medium tabular festoon cross bedding	Tidal and long shore currents
Offshore			Packstone to Wackestone		Horizontal branching burrows mottled texture	Bioturbation
Transgressive lag gravel			Intraclast conglomerate		Coarse chaotic mixture	(surf action)
					Bored limestone surface	

Carbonate Environments, Figure 8 Scheme of beach zonation and sediments. (From Inden and Moore, 1983; with permission of American Association of Petroleum Geologists.) HWM: high water mark; LWM: low water mark.

conditions. Water depths range from 0 to 200 m. The areas covered are several 1,000 km^2 to several 100,000 km^2 in the case of the Queensland shelf. In between the coast and platform margins, low and high islands as well as patch reefs and faroes (small circular lagoon reefs) of coral can be found.

Carbonate sediments on these rimmed platforms are almost exclusively of skeletal origin and facies belts more or less parallel to the coast (Ginsburg, 1956; Maxwell, 1968; Purdy and Gischler, 2003; Purdy et al., 2003). On the Belize and Queensland shelves, siliciclastics derived from the hinterland mix with carbonate sediment. As

Carbonate Environments, Figure 9 Tropical beach with beachrock outcrop. Loggerhead Key, Dry Tortugas, South Florida.

a consequence, facies are siliciclastics-dominated near the coast. Marls are found in the middle shelf. Largely carbonate sediments occur at the outer platform and platform margins. Central shelf marls are rich in fragments of mollusks and foraminifera. Mud content is high. In Belize and NE Australia, pelagic biota such as pteropods and planktonic foraminifera, respectively, are found in these platform areas due to the high depths of the shelves. Shelf reefs and outer platforms and margins are predominated by mostly coarse-grained sediments rich in coral, coralline red algae, *Halimeda*, and benthic foraminifera (Figure 17).

Platform margins are several kilometers wide and water depths are between 0 and 50 m. Coral reefs are widespread with the common genera *Acropora*, *Porites*, *Montastraea*, *Diploria*, *Platygyra*, *Pocillopora*, and *Millepora*. The reef crest proper is at sea level. The shallow forereef slope is often characterized by spur-and-groove systems (Shinn et al., 1981) (Figure 18). In the case of Belize, a steep sometimes vertical slope occurs in the deep forereef. This bypass margin was investigated in detail by James and Ginsburg (1979). The Florida and Queensland margins with lower angles are examples of depositional margins (Read, 1985).

Isolated platforms

The largest known isolated carbonate platforms include the Bahamas in the Atlantic and Great Chagos Bank in the Indian Ocean, which each cover 10,000 to 100,000 km^2, respectively. In the Bahamas archipelago, Great Bahama Bank (Purdy, 1963; Ginsburg, 2001) is probably the best known example (Figure 19). Great Bahama Bank is a shallow, low-relief submarine bank, which is surrounded by deep water. Water depths on the bank only rarely exceed 6 m. The bank has very high and steep bypass margins, especially on the eastern side. The steepness is presumably the result of both biological construction and erosion. On the eastern side, a well-developed coral reef rim with abundant *Acropora* exists in shallow water. Otherwise, the margins of Great Bahama Bank are largely predominated by ooid sand shoals. These shoals are very shallow, high-energy environments where ooids form and are moving in the form of up to several kilometer-long rippled sand waves, more or less parallel to the bank margins. Joulters Cays ooid shoal in the north-eastern corner of the bank is intensively studied (Harris, 1979). The western bank margin also has long and wide ooid shoals. Near the southern end of Tongue of the Ocean, ooid sand waves are oriented perpendicular to the bank margin. The eastern interior part of Great Bahama Bank is covered by the Pleistocene island of Andros, which gets as high as 30 m above present sea level. Most of the platform interior is covered by shallow water as mentioned above. Benthic marine animals such as gastropods (*Strombus*) and echinoids (*Oreaster*) are rather rare. Codiacean algae such as *Penicillus* and sea grass (*Thalassia*) are common. Salinities are above normal marine on the bank and may reach 45% close to Andros Island. Sediments are mostly non-skeletal and include peloidal muds as well as aggregate grain ("grapestone") and peloidal sands (Purdy, 1963). Skeletal sands only occur around the eastern reefal margin and in a small band on the western margin.

The origin of carbonate mud on Great Bahama Bank is a matter of debate since several decades [see discussion in

a

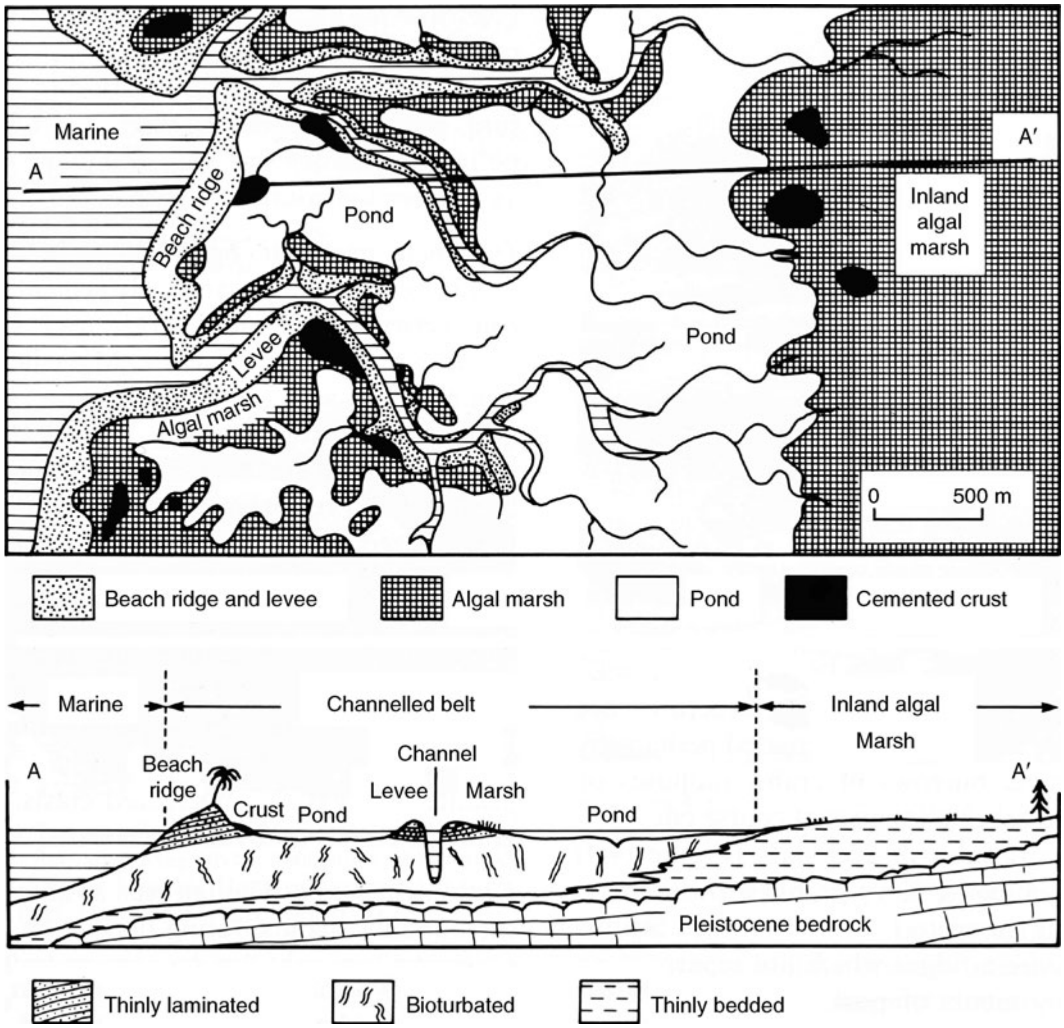
b

Carbonate Environments, Figure 10 Photos from tidal flats. (**a**) Aerial view of western part of Andros Island, Bahamas. (Photo courtesy of Robert Ginsburg.) (**b**) Microbial mat on sabkha in Khiran, southern Kuwait.

Gischler and Zingeler (2002)]. Mud largely consists of micron-sized aragonite needles. It potentially originates from codiacean algae such as *Penicillus*, from physico-chemical precipitation in the warm and saline surface waters of the bank, or from biologically (bacterially) induced precipitation of calcium carbonate. Commonly occurring clouds of suspensions of carbonate mud and water on the bank, known as "whitings," have been described as expressions of spontaneous precipitation phenomena. However, there is also evidence that whitings are simply a consequence of bottom sediment stirring either by fish or by currents.

More than 2 m high subtidal stromatolites were discovered by Dill et al. (1986) on the Bahamas in normal marine seawater near the Exuma islands. Previously, only small microbial buildups were known from the Bahamas. These stromatolites develop within very strong currents and significant bottom sediment movement. Apparently sediment movement largely excludes grazers and triggers carbonate accumulation via trapping and binding.

Seismic profiling and deep drilling has revealed that the geologic history of Great Bahama Bank was not simply a process of aggradation (Ginsburg, 2001). A complex interplay of slow subsidence and sea-level change has



Carbonate Environments, Figure 11 Tidal flat scheme from Andros Island, Bahamas. (From Scoffin, 1987; with permission of Blackie and Son Ltd.)

modified the bank during the Neogene in that shallow water banks repeatedly aggraded, backstepped, and prograded, thereby filling deep seaways and moving the western margin of Great Bahama Bank by 30 km to the west into the Straits of Florida.

Carbonate ramps

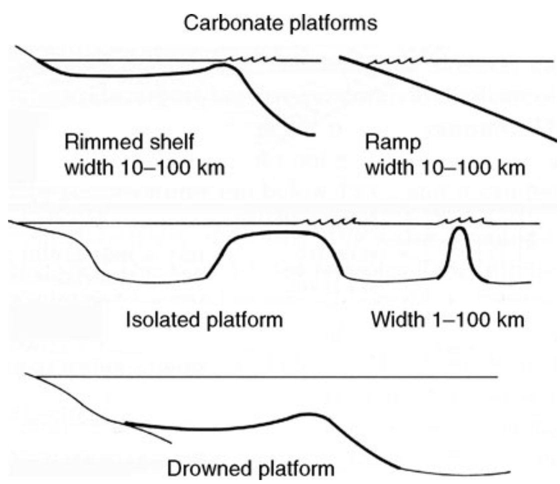
The term ramp was introduced to the carbonate literature for unrimmed carbonate shelves with a low-angle dip ($<1°$) toward the open sea (Burchette and Wright, 1992; Wright and Burchette, 1998). The classical modern example is the southwestern Persian Gulf. Other modern ramp examples include the western Florida margin or Campeche Bank, Mexico. Geometrically, there are homoclinal and distally steepened ramps (Figure 20). Depositional environments are usually divided into inner, middle, and outer ramps. The inner ramp environment is above fair-weather wave base (FWWB); the mid-ramp between FWWB, and the storm wave base (SWB); the outer ramp below SWB. In the southern Gulf, inner ramp environments consist of shoals and barrier facies with ooid-skeletal sands. Smaller reefs are found just seaward of the barriers. Back barrier sediments are usually fine-grained peloidal and mollusk muds and sands. Mollusk-rich silts and sands are common on the mid-ramp. Outer ramps are characterized by carbonate and argillaceous muds ("marl") with abundant mollusks. In the northern Gulf, barriers are largely lacking. In contrast to the southern Gulf, which is tide-dominated, the northern Persian Gulf ramp is largely wave-dominated. On the inner ramp, quartz grains derived from the adjacent mainland desert may be present in addition to ooid shoals. Small coral patch reefs may also be found in the inner ramp zone. Mid-ramp and outer ramp deposits are also dominated by mollusk-rich silts and sand and argillaceous muds, respectively. Pinnacle-type patch reefs of coral occur in the outer ramp.

Characteristic storm beds or tempestites in mid-ramp settings were identified by Aigner (1982) in the European Triassic Muschelkalk beds. Tempestites exhibit characteristics such as sharp bases, sole marks, graded bedding, hummocky cross-stratification, and ripple marks. Because there are no larger outcrops in modern ramps and shelves showing complete sedimentary bodies or large-scale sedimentary structures, our knowledge of storm beds is largely based on the fossil record.

Temperate- and cool-water shelves and platforms

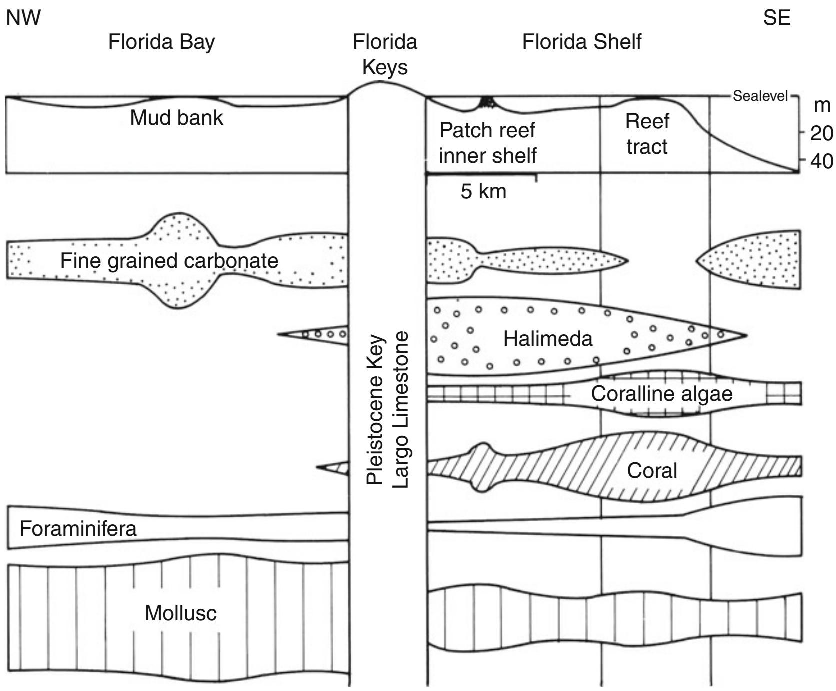
Apart from the abundant typical warm water and low latitude deposits, about one third of the modern carbonate shelves and platforms are characterized by temperate and cool-water carbonate sedimentation (Lees and Buller, 1972; James and Clarke, 1997). Prominent and well-studied modern examples are the Great Australian Bight or the Mediterranean realm. Temperatures in the cool-water realm are below 20°C and heterotrophic carbonate producers predominate. They include mollusks, echinoderms, bryozoa, foraminifera, coralline algae, barnacles, and serpulids ("heterozoan association"). Constituent carbonate particles of the same taxa are also present in warm-water carbonates; however, there, they are swamped by carbonate grains from rapidly growing producers that have photosymbionts such as corals and codiacean algae as well as by mud and non-skeletal carbonate formation as in ooids, peloids, and cements ("photozoan association"). There are also cases where heterozoan carbonates occur in warm-water settings, e.g., in western Florida or to some degree on Campeche Bank, Mexico, as well as in several fossil examples. Elevated nutrient contents and primary production are usually responsible, and hamper abundant photozoan carbonate production (Figure 21).

Temperate and cool-water carbonates usually form unrimmed shelves or platforms, and they are strongly influenced by hydrodynamic energy. Submarine cementation is rather weak. In modern examples, kelp (brown algae) forests occur in the constantly wave-dominated inner ramp. Siliciclastic sediment may derive from local sources from the mainland. Sand waves and dunes are found on the mid-ramp, which is prone to storm-reworking. Sediments on inner and mid-ramps are mostly coarse-grained and frequently redeposited. The outer ramp environment may contain fine-grained sediment with bryozoan, sponge, or non-zooxanthellate coral mounds.

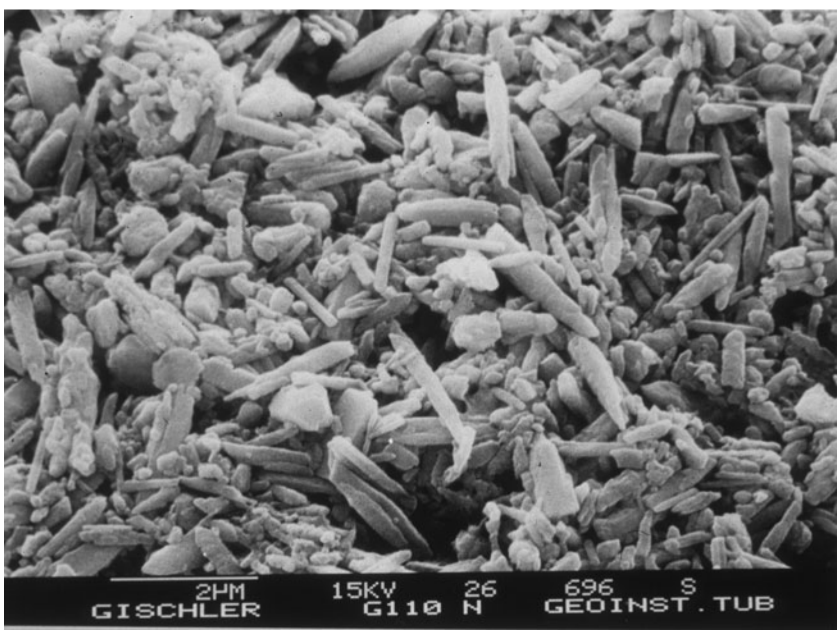


Carbonate Environments, Figure 12 Scheme of carbonate platform types and morphologies. (Modified from Tucker and Wright, 1990.)

Carbonate Environments, Figure 13 Underwater photograph of *Callianassa* mounds, Florida Bay.



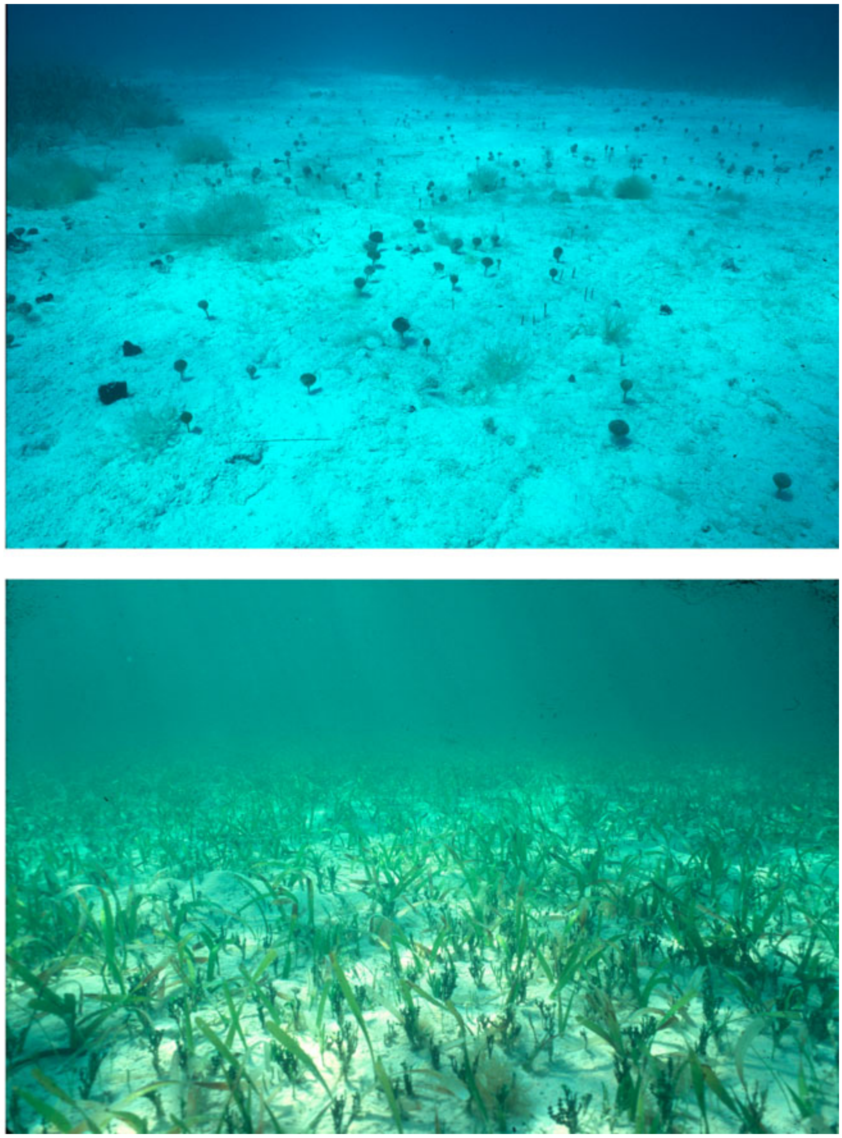
Carbonate Environments, Figure 14 Sediment distribution in Florida Bay and the reef tract. (Modified from Ginsburg, 1956.)



Carbonate Environments, Figure 15 SEM photo of carbonate mud from Belize with abundant aragonite needles and nanograins, which originate from codiacean algae.

Reefs

Reefs are classical objects of geobiological research because they are rock formations built by the biological activity of organisms. Reefs normally form topographic highs above the surrounding sea floor. Reef organisms comprise large numbers of taxa, which either construct a framework or baffle and bind sediment, which is rapidly consolidated. In addition, there are reef dwellers, which do not necessarily contribute to reef growth, and, there are reef-destroying organisms, which bore into, dissolve, or



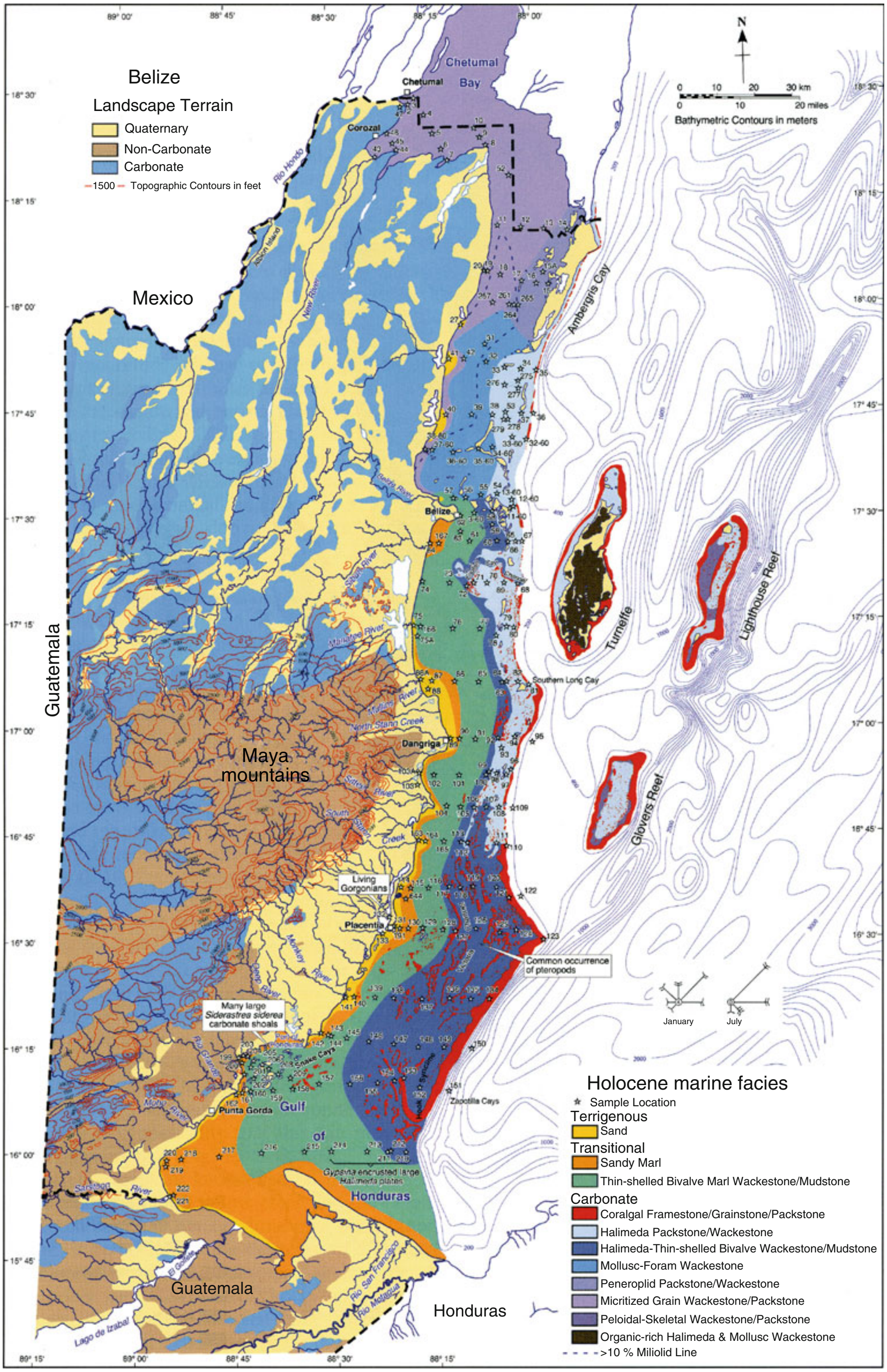
a b

Carbonate Environments, Figure 16 Underwater photographs of calcareous codiacean algae (**a**) *Penicillus* and (**b**) *Halimeda*, Belize.

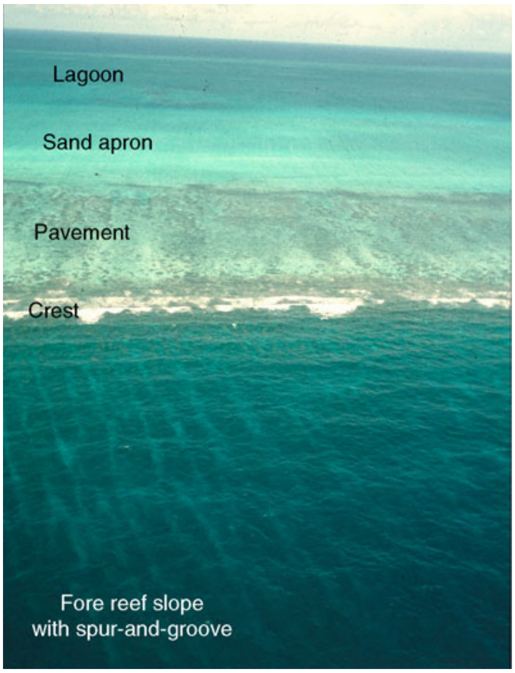
erode reef limestone. This ecological guild concept of constructors, bafflers, binders, dwellers, and destroyers was developed for reef organisms by Fagerstrom (1987), who compiled a geological history of reef building. Like Wilson (1975) before, he stressed that reef building is a geologically old phenomenon that reaches back into the Archaean, however, reef-building organisms have repeatedly changed with time. Recently, Wood (1999), Stanley (2001), and Kiessling et al. (2002) have provided additional monographs and editions on the long history of reef building. The most abundant groups of reef builders during earth history have been microbes, corals, sponges, bryozoa, and rudist bivalves. Reef destroyers include boring sponges, bivalves, worms, and microborers as well as grazing gastropods, echinoderms, and fish. Reef building is a complex interplay of reef construction, destruction, sedimentation, and cementation, processes which occur at the same time and eventually produce what geoscientists would term reef limestone (Scoffin, 1992) (Figure 22).

Tropical reefs

Modern tropical reefs cover an area of about 600,000 km^2 in tropical-subtropical latitudes around the globe. Reefs thrive in warm waters between 18 and 30°C, normal marine salinity, and clear and oligotrophic waters. They represent the most diverse marine ecosystem with several tens of thousand of species including taxa from all known animal phyla. Scleractinian corals and red coralline algae are the dominant builders (Figure 23). Other calcareous



Carbonate Environments, Figure 17 Carbonate sediment distribution on the Belize shelf and platforms. (With permission of Purdy and Gischler, 2003.)



Carbonate Environments, Figure 18 Aerial photograph of zoned margin in Lighthouse Reef, Belize.

algae such as codiaceans, mollusks, benthic foraminifera, and echinoderms are important sediment producers on tropical reefs. Reef-building Scleractinian corals live in symbiosis with dinoflagellates ("zooxanthellae"), which live in their tissues and photosynthesize. The symbionts are both important for the coral metabolism by providing food and also enhance skeletal growth of the aragonite coral skeleton because they consume carbon dioxide.

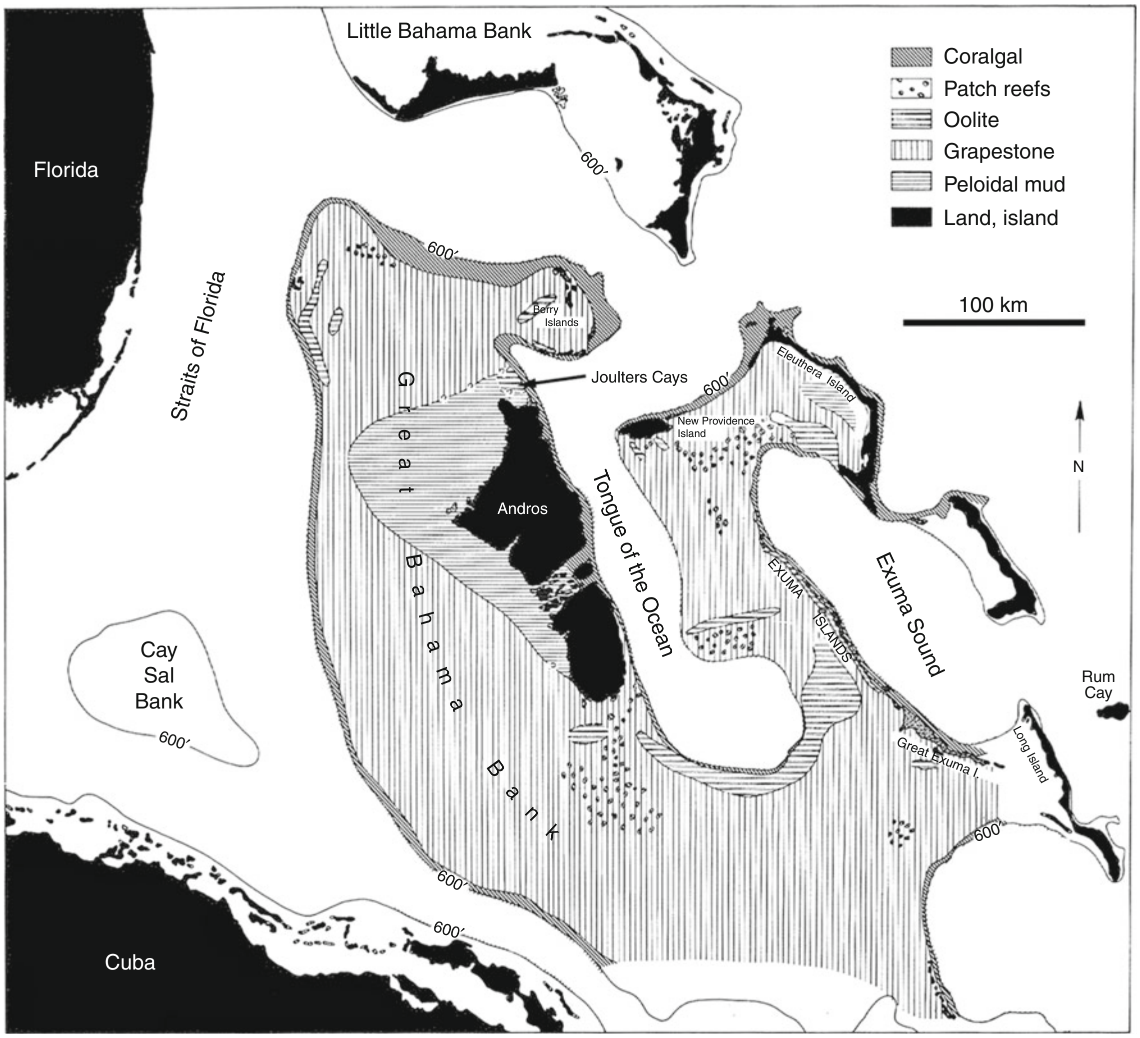
$$CO_2 + H_2O \longrightarrow [CH_2O]_n + O_2 \tag{2}$$

Because of the symbiosis, the tops of tropical reefs are at or close to sea level. Fossil coral reefs are therefore used as gauges for past sea levels. The annual banding of coral skeletons and the longevity of colonial reef corals makes them ideal archives of historical climate change ("sclerochronology").

Since Darwin (1842), tropical reefs are morphologically subdivided into fringing reefs, barrier reefs, and atolls. Fringing reefs occur close to the coast with a shallow moat or lagoon in between reef and coastline. These reefs are abundant in the Caribbean Sea and the Indonesian archipelago. Barrier reefs are larger structures that are separated from the coast by a deep channel, like in northeastern Australia, New Caledonia, or Belize. The Great Barrier Reef is the longest barrier reef with more than 2,000 km in length. Barrier reefs in the sense of Darwin surround volcanic islands like, e.g., Bora Bora in the Tuamotu archipelago. Atolls are circular reefs, which enclose a deep lagoon. There are about 400 atolls worldwide. They are most common in the Pacific and Indian Oceans, and have a volcanic basement. According to Darwin's (1842) elegant subsidence theory, fringing reef, barrier reef, and atoll are genetically connected by the simple process of subsidence of a volcanic island. Tayama (1952) carried the theory even further and explained the formation of small circular reefs with shallow or filled lagoons, so-called table reefs, as the final stage in reef development before drowning and seamount (guyot) formation. Even so, the subsidence theory does not explain large barrier reefs that are attached to large continental land masses, and, it neglects the significance of eustatic sea-level change, which was first considered in Daly's (1910) glacial control theory. Indeed, drilling on Enewetak Atoll in the 1950s has demonstrated about 1.3 km of shallow-water reef limestone on top of an Eocene volcano, but numerous hiatuses were found, which are the result of repeated subaerial exposure. The antecedent karst model of Purdy (1974) takes into account the importance of limestone dissolution by meteoric waters during eustatic lowstands of sea level, which are even able to produce fringing, barrier, and atoll reef morphologies as well as spur-and-groove patterns in subaerially exposed limestone and limestone-volcanic islands. Recent studies have shown that modern barrier reefs are comparably young features that only came into existence about 400–600 ky ago when high amplitude, eccentricity-driven eustatic sea-level changes became more and more dominant (International Consortium of Great Barrier Reef Drilling, 2001; Multer et al., 2002).

Tropical reefs are characterized by ecological zonation, with zones running more or less parallel to bathymetry (see Figure 18). Modern reefs also exhibit provincialism between Atlantic and Indo-Pacific realms. For high sea-level greenhouse climates in earth history, like in the Devonian or Jurassic and Cretaceous, however, no evidence of provincialism was found. Common reef zones include from outboard to inboard the fore reef, reef crest, back reef, sand apron, and lagoon. Surface sediments are coarse sands of coral and calcareous algae in reefal environments and more fine-grained, mollusk- and foraminifer-rich sediments in lagoonal areas. Several shallow lagoons of Caribbean reefs are rich in non-skeletal grains. Milliman (1974) explained this difference to Indo-Pacific reefs, which he termed the "ooid problem," by the shallower and smaller lagoons, in which water exchange is much faster. Carbonate-rich water circulates more rapidly and thereby preferentially produces non-skeletal grains such as ooids, peloids, and aggregates by submarine cementation (Figure 24). However, the discovery of abundant modern non-skeletal grains such as ooids and cemented fecal pellets in the shallow lagoons of Aitutaki Atoll (Cook Islands) and Bora Bora (Tuamotus) in the south Pacific opposes this widely held concept (Rankey and Reeder 2009; Gischler 2010).

Drilling in the barrier reef around the island of Tahiti (Montaggioni and Camoin, 1993) and investigations in cavities of the Great Barrier Reef (Reitner, 1993) have shown the importance of abundant microbial activity in



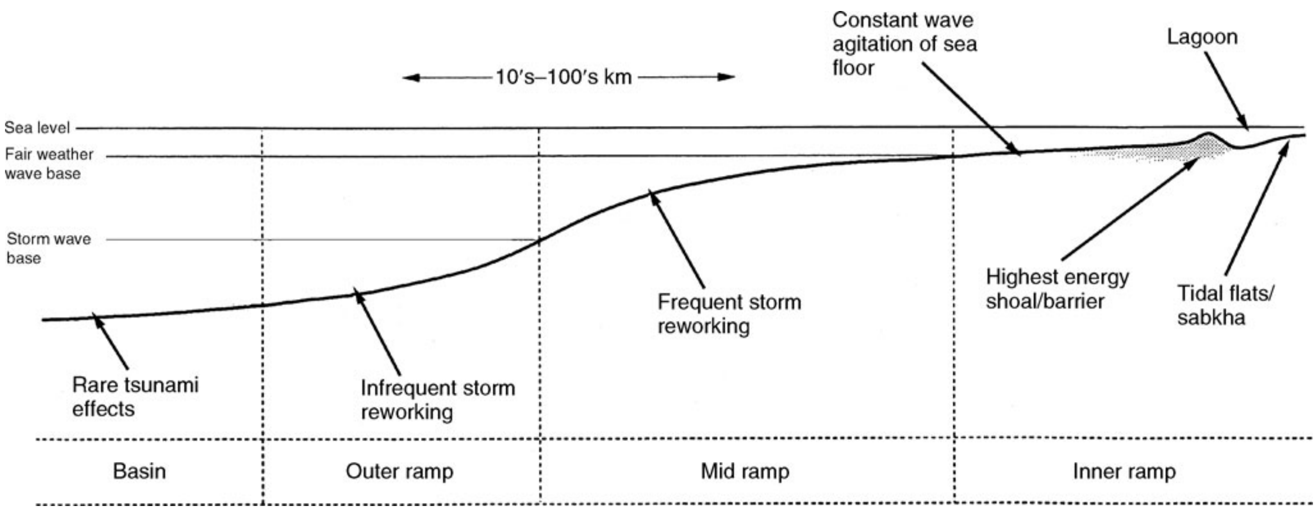
Carbonate Environments, Figure 19 Sediment map of the Great Bahama Bank. (With permission of Traverse and Ginsburg, 1966.) Bank margin is defined by 600 ft (180 m) depth contour.

tropical coral reef cavities. Especially in Tahiti, stromatolite-type formations are quite abundant, even though microbial activity was apparently most widespread during the early Holocene. In the Great Barrier Reef example, sponges and sponge tissue decay are additional agents besides microbes that trigger carbonate formation.

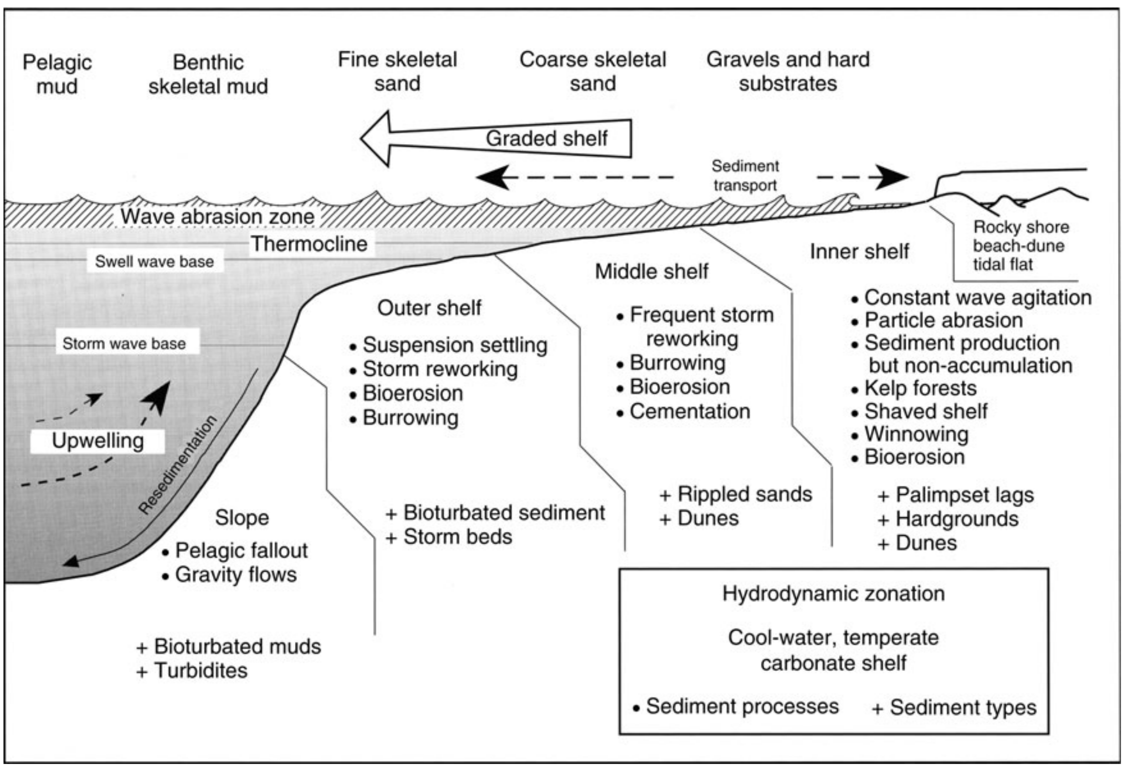
Deep- and cold-water reefs and mounds

After the first discovery of abundant coral reefs in deep and cold water off the Scandinavian coast (Teichert, 1958), it took several decades before these reef structures were investigated in great detail (Freiwald and Roberts, 2005; Roberts et al., 2006). These reefs are largely built by non-zooxanthelate branched corals such as *Lophelia*, *Madrepora*, *or Cladocora* in water depths below 200 m and temperatures between 4 and 2°C. Reef thickness may reach more than 30 m and individual reef structures have lengths of more than 10 km. Species richness in the north Atlantic examples is around 1,300 taxa. Deep-water reef foundations are usually submarine topographic highs such as, e.g., iceberg plough marks. Meanwhile, it has become clear that deep- and cold-water reefs in the northeastern Atlantic are not restricted to the Scandinavian area but are also abundant off the Atlantic shelf margin as far south as the Mediterranean and in the Caribbean.

Somewhat similar deep- and cold-water reefs, termed lithoherms, were discovered in strong currents at the bottom of the Straits of Florida in 600–700 m of water (Neumann et al., 1977). Lithoherms are several 100 m long and up to 50 m high. Their shape is ellipsoidal with the long axis parallel to the current. The faunal elements mainly include non-zooxanthellate corals, crinoids, sponges, and bryozoa. Interstitial muddy to sandy detrital carbonate sediment is abundant and largely a product of baffling. The mounds are cemented at the surface in that concentric, hard carbonate crusts occur. Stromatactis-type fenestral



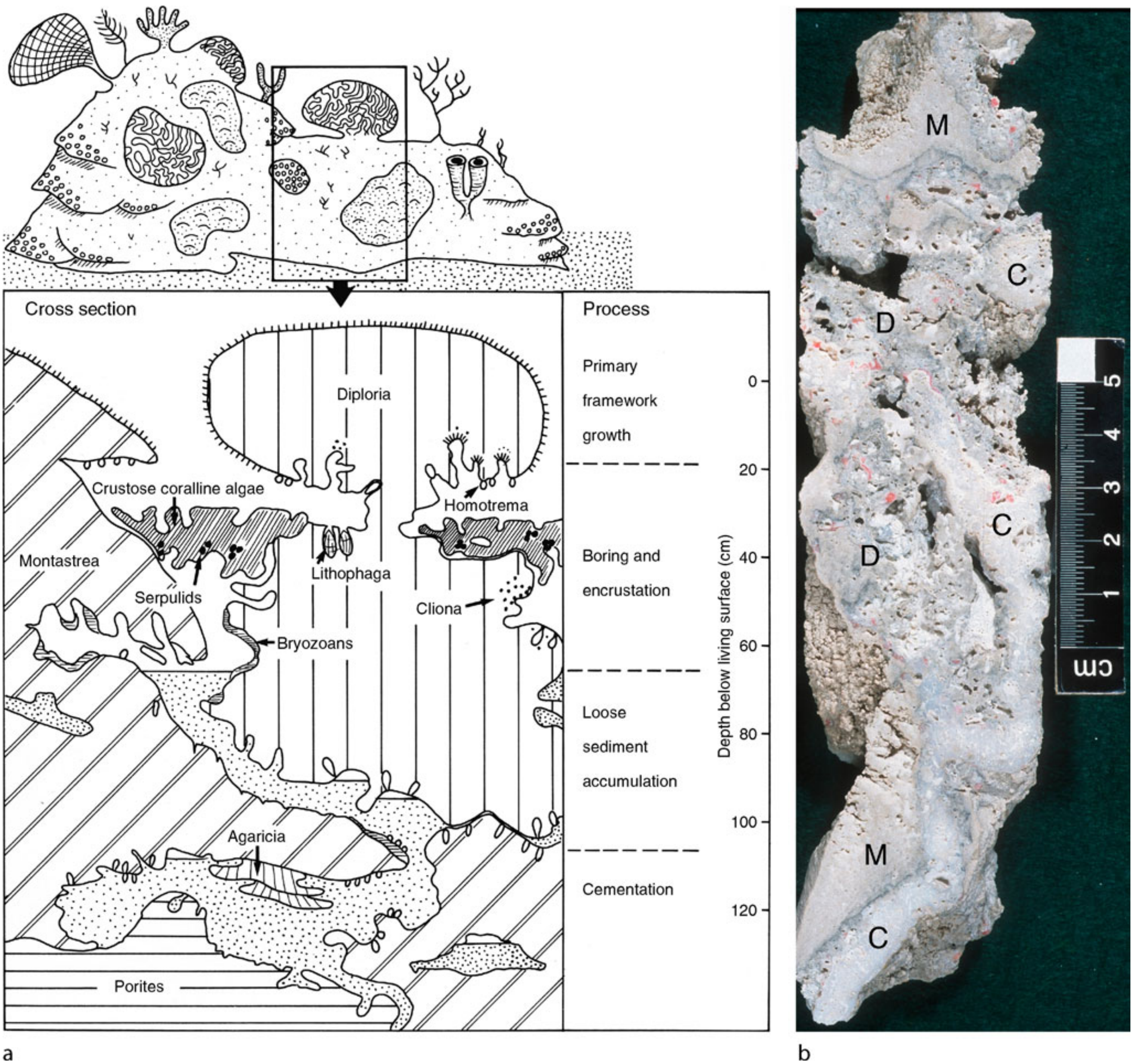
Carbonate Environments, Figure 20 Scheme of a ramp. (Modified from Burchette and Wright, 1992.)



Carbonate Environments, Figure 21 Scheme of cool-water shelves. (Modified from James and Clarke, 1997.)

cavities were discovered and attributed largely to burrowing and subsequent sediment infill.

It is tempting to interpret the Florida lithoherms as modern analogues of fossil deep-water mud mounds. Mud mounds are common in the fossil record, especially in the Devonian and Carboniferous (Monty et al., 1995). The most prominent examples are probably the Carboniferous ("Waulsortian") mud mounds. Waulsortian mounds or banks also formed in deep water and contain biota with crinoids, sponges, bryozoa, brachiopods, and mollusks. Corals are rare. Stromatactis, large cement and sediment-filled fenestral cavities with flat bases and irregular tops, is very common in Paleozoic mud mounds (Krause et al., 2004). Even so, major differences between



Carbonate Environments, Figure 22 (**a**) Scheme of reef limestone formation with synchronous processes. (From Scoffin, 1992; with permission of Springer: Berlin, Heidelberg.) (**b**) Quaternary reef limestone from Belize Barrier Reef. *C* coral; *M* microbialite, *D* detritus.

the lithoherms and the Waulsortian mud mounds appear to be the abundance of carbonate mud in the fossil examples. Lithoherms have higher quantities of coarse sediment. Also, lithoherms are largely biodetrital mounds whereas the Waulsortian ones are microbial mounds in which in situ mud production prevailed (Monty et al., 1995).

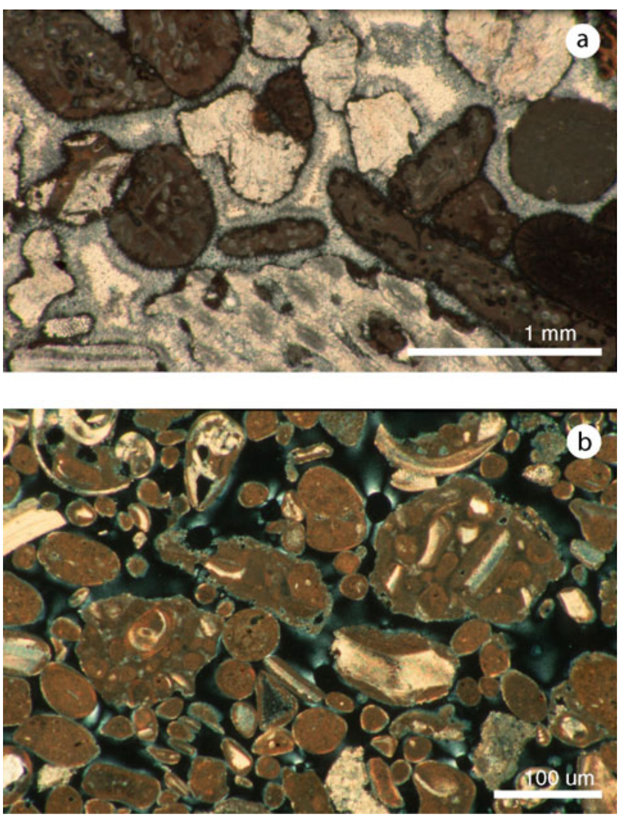
Also, mud mounds and deep-water carbonate mounds have been found to be associated with cold seeps or hot vents (Beauchamp and von Bitter, 1992; Mounji et al., 1998). Around locations of seepage of methane (CH_4) and hydrogen sulfide (H_2S) on the seafloor, authigenic carbonate precipitation may be triggered by bacteria via chemosynthetic processes that produce bicarbonate:

$$CH_4 + SO_4{}^{2-} \longrightarrow HCO_3{}^{-} + HS^{-} + H_2O \qquad (3)$$

$$SO_4{}^{2-} + 2(CH_2O)_n \longrightarrow 2HCO_3{}^{-} + H_2S \qquad (4)$$

Usually, metazoans such as certain bivalve mollusks (*Caplyptogena*, *Bathymodiolus*, lucinids), vestimentiferan worms (*Riftia*), and thyasirid shrimps, which have

Carbonate Environments, Figure 23 Underwater photograph of a reef with abundant acroporids and milleporids. Glovers Reef Atoll, Belize.

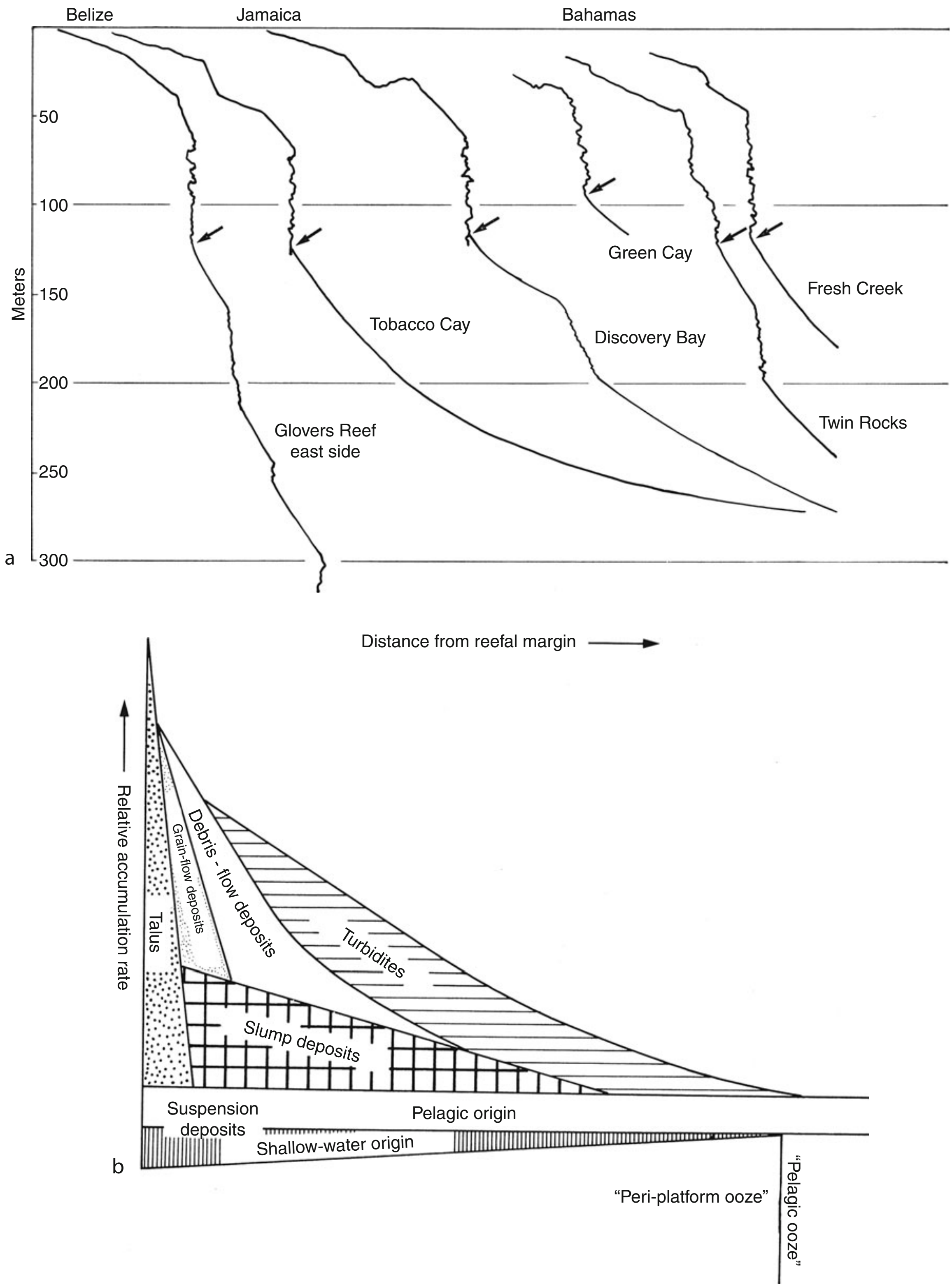


Carbonate Environments, Figure 24 Thin-section photographs of typical (**a**) skeletal carbonate sediment in beachrock with corals (light) and *Halimeda* (dark) and aragonite needle cement; (**b**) non-skeletal carbonate sediment with aggregate grains, ooids, and peloids. Kuwait, northern Persian Gulf.

chemosynthetic bacteria in their gills and tissue thrive at these seeps. Interestingly, fossil seeps often have abundant brachiopods (dimerelloids, e.g., *Peregrinella*), which do not have modern representatives that harbor chemosynthetic bacteria (Campbell and Bottjer, 1995). Meanwhile, fossil seep carbonates have been reported from rocks as old as Silurian (Little et al., 1997). Apart from their characteristic fauna and texture, and their patchy distribution, methane seep carbonates may be identified rather easily in the fossil record by geochemical means. Carbon isotopes of these carbonates are typically very strongly depleted (Beauchamp and von Bitter, 1992).

Slope

Carbonate slopes may have lower angles as in depositional margins or higher angles in bypass margins. In general, carbonate and siliciclastic slopes have similar heights, however, carbonate slopes are usually steeper (Schlager, 2005). James and Ginsburg (1979) have compiled slope morphologies of several western Atlantic carbonate margins, which usually exhibit a threefold division (Figure 25). There is a low-angle (20–30°) fore reef slope, a steep slope to vertical wall between ca. 50–150 m and a 50° slope below the wall, which is dominated by carbonate debris and blocks transported down from shallower depths. Foliaceous corals, *Halimeda*, crinoids, and sponges may be found colonizing the walls. In general, the slope is characterized by sediment redeposition as seen, e.g., in the occurrence of talus blocks, debris-flows, and slump deposits, as well as breccias (James and Ginsburg, 1979; Schlager, 2005). Limestone turbidites (Meischner, 1964) originating from platform tops occur on lower slopes and in the adjacent

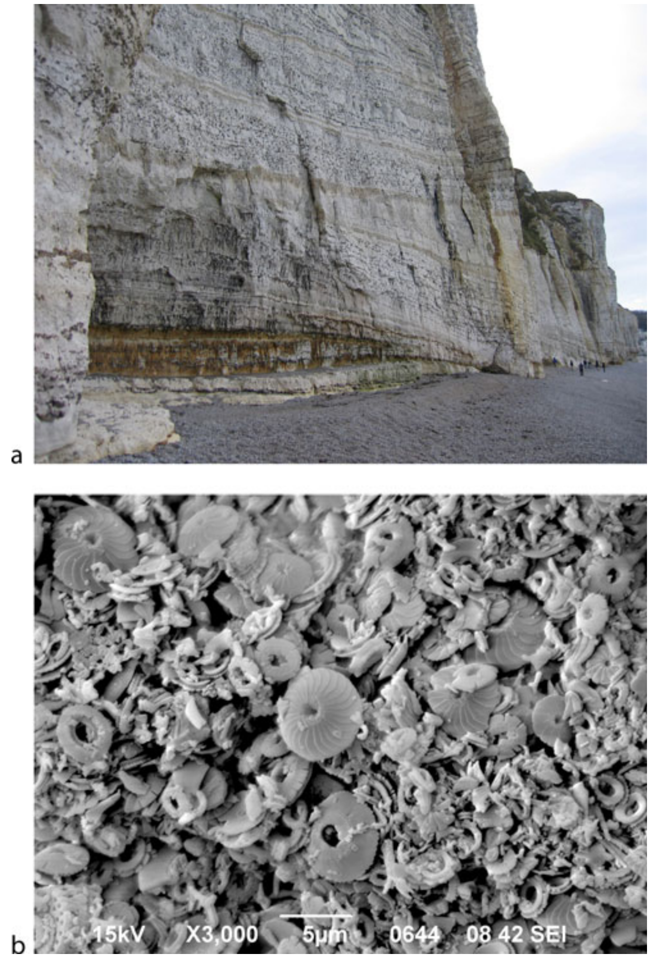


Carbonate Environments, Figure 25 (**a**) Schemes of fore reef slopes. (Modified from James and Ginsburg, 1979.) (**b**) Sedimentary processes on slopes. (From Enos and Moore, 1983; with permission of American Association of Petroleum Geologists.)

toe-of-slope among pelitic deposits. Limestone turbidites have similar characteristics as siliciclastic turbidites, with sharp bases, graded beds of debris, and lenticular geometries of individual turbidite bodies. Highstand shedding describes the fact that limestone turbidites in the Quaternary are more common during sea-level highstands when platforms are flooded and abundant carbonate produced as opposed to lowstands of sea level when platforms are subaerially exposed (Droxler and Schlager, 1985). Recent investigations in fossil slope deposits have revealed the formation of so-called auto-micrites as a result of microbial activity on carbonate slopes (Keim and Schlager, 2001). Automicrite facies is apparently a significant agent for platform margin and slope stabilization in these Triassic examples.

Basin

Basinal carbonate environments are in depths between about 200 and 5,000 m (Hsü and Jenkyns, 1974; Crevello and Harris, 1985). The basins of the Atlantic, Indian Ocean, and large parts of southern Pacific Ocean are covered by very fine-grained carbonate sediments, which are termed ooze. Consolidated ooze is called chalk. Carbonate deposition and accumulation is controlled by carbonate supply from the pelagic realm and largely includes tests of planktonic foraminifera, pteropod gastropods, and coccoliths (Figure 26). These particles are transported to the sea floor as "marine snow," which is largely composed of fecal pellet aggregates. Fecal pellets derive from arthropods such as krill and copepods, and sinking velocities of pellets are much faster as compared to those of individual tests. Other minor contributors of calcium carbonate to the deep sea include shells of ostracods, shells of pseudoplanktonic organisms such as arthropods and bivalve mollusks, and cephalopod shells from *Nautilus*, *Sepia*, and *Spirula*. Carbonate accumulation is also dependent upon the position of the aragonite and calcite compensation depths (ACD, CCD), respectively. The ACD lies at about 2,000–2,500 m and the CCD at about 5,000–5,500 m depth. Below the CCD all calcium carbonate is dissolved. Biota in basinal environments include benthic foraminifera, echinoderms, sponges, and crustaceans. Ichnospecies are *Planolites*, *Chondrites*, and *Zoophycos*. *Thallassinoides*-type bioturbation, often filled by chert, is typical in Cretaceous deep-water limestone. Except for fine lamination, sedimentary structures are usually rare. In Quaternary deposits, carbonate cycles were discovered, which are caused by changing carbonate saturation states during glacials and interglacials. During glacials, lower sea-level positions result in higher carbonate saturation in the pelagic realm, because the neritic carbonate factory is largely shut off. During high sea levels, carbonate production is more or less equally distributed between the neritic and pelagic realm, leading to lower carbonate content in basinal deposits. The genesis of deep-water carbonates during the Paleozoic is somewhat enigmatic, because calcareous plankton only evolved in the Mesozoic. Potential Paleozoic contributors include cephalopods, ostracods, tentaculites, calcispheres, and possibly very fine-grained detritus from the neritic realm. Cephalopod limestone is a typical Paleozoic deposit of shallow basins or deeper submarine rises with no modern counterpart (Wendt and Aigner, 1985).



Carbonate Environments, Figure 26 (**a**) Outcrop of Cretaceous chalk in Etretat, northern France. Dark spots are silicified *Thallassinoides* burrows. Erosional surfaces probably caused by currents and/or slumping/sliding. (**b**) SEM photograph of Pleistocene chalk with abundant coccoliths, Angola Basin, Atlantic.

Summary

Carbonate depositional environments were described in an idealized cross section from terrestrial to transitional, shallow marine, and to deep marine realms. Environments discussed include springs, lakes, subaerial exposure horizons, beaches and shores, tidal flats, restricted, open-rimmed, isolated shelves and platforms, ramps, tropical and cold deep-water reefs and mounds, slopes, and deep sea basins. The great majority of carbonates forms in the marine realm, and, carbonate formation is largely a product of biological activity. Biological control is either direct, e.g., during growth of shells and skeletons of invertebrates such as corals, mollusks, brachiopods, echinoderms, and sponges, tests of foraminifera or carbonate particles produced by calcareous algae or coccolithophorids. Indirect biological control relates to metabolic processes, e.g., photosynthesis of microbes, which triggers carbonate precipitation. Abiotic formation of carbonates is the exception and includes the formation of some carbonate muds, cements, certain dolomites, and non-skeletal grains such as grapestones, lumps, or ooids. Elevated temperature and salinity, high concentrations of calcium, magnesium, and carbonate ions, low pressure, and CO_2-removal are major environmental parameters that potentially trigger abiotic carbonate formation.

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