

Chapter 5

LIMESTONES

1. INTRODUCTION

1.1 Something like about one-fifth of all sedimentary rocks are carbonate rocks. The two main kinds of carbonate rocks, limestones and dolostones, together with sandstones and shales, are what might be called the “big four” of sedimentary rock types. I’m reluctant to try to guess what percentage of all sedimentary rocks those “big four” account for, but the figure must be in the upper nineties. Moreover, carbonate rocks are economically important because together with sandstones they constitute reservoirs for most of the world’s petroleum and gas reserves (and let’s not forget that they are the source of all of the world’s portland cement—not a jazzy, exciting resource, but a very important one for our modern civilization).

1.2 Up until about fifty years ago, the petrologic study of carbonate rocks lagged far behind that of siliciclastic rocks. Since that time, however, there has been great progress, as it has become realized that to a great extent carbonate rocks can be treated as clastic deposits analogous to sandstones and shales (the main exception being reef limestones). Great progress has also been made in recent years on the geochemistry of carbonate precipitation, the role of organisms in carbonate deposition, and the diagenesis of carbonate sediments.

1.3 Carbonate sediments are often described as *chemically precipitated*. In one sense, that’s true: they are formed by precipitation of one or another carbonate mineral in various sedimentary environments. But don’t let the term “chemically precipitated” fool you: they don’t form in the same way that rock candy does from a sugar solution on your windowsill. Some carbonate sediments are indeed precipitated directly from seawater, in the form of fine crystals in the water column, which then settle to the seafloor, or as successive spherical shells deposited around a nucleus particle in a warm, shallow marine environment. Most, however, are *biochemically* precipitated, in the tissues of organisms, mainly marine invertebrates of various phyla.

1.4 The title of this chapter could have been “carbonate rocks”, but it seems somewhat more natural to restrict it to limestones. You will learn that dolomite is almost invariably a secondary rather than a primary sedimentary mineral, meaning that carbonate sediments don’t start their lives as dolomite. Much dolomite is precipitated very early, however, at shallow depths in the originally calcium carbonate sediment, although much is also precipitated late, after deep burial has produced solid limestone. Material on dolostone (a carbonate rock consisting mainly of the mineral dolomite) is postponed until the later chapter on diagenesis.

2. CARBONATE MINERALS

2.1 About sixty minerals have the carbonate ion in their composition. But there are only three really important carbonate minerals: *calcite*, *aragonite*, and *dolomite*. (In the parlance of mineralogy, the first two are said to be *polymorphs*.) And aragonite is unimportant in ancient rocks, because it reverts to calcite with time. Other sedimentary carbonates of non-negligible importance are *magnesite* (magnesium carbonate) and *siderite* (ferrous iron carbonate). Dolomites containing some percentage of Fe^{2+} are called *ferroan dolomite*. The middle member of the range between dolomite and hypothetical $(\text{Fe}, \text{Ca})(\text{CO}_3)_2$ is called *ankerite*. There are also a few significant carbonate evaporite minerals (trona, natron) we won't consider here.

2.2 Figure 5-1 is a composition triangle showing the range of carbonate minerals stable at the low temperatures at or near the Earth's surface. It's in terms of the three divalent positive ions, Ca^{2+} (0.99 Å), Mg^{2+} (0.66 Å), and Fe^{2+} (0.74 Å), that are abundant and of about the right size to fit into carbonate structures. (Remember that one angstrom is equal to 10^{-10} meters, or 0.1 nanometers.)

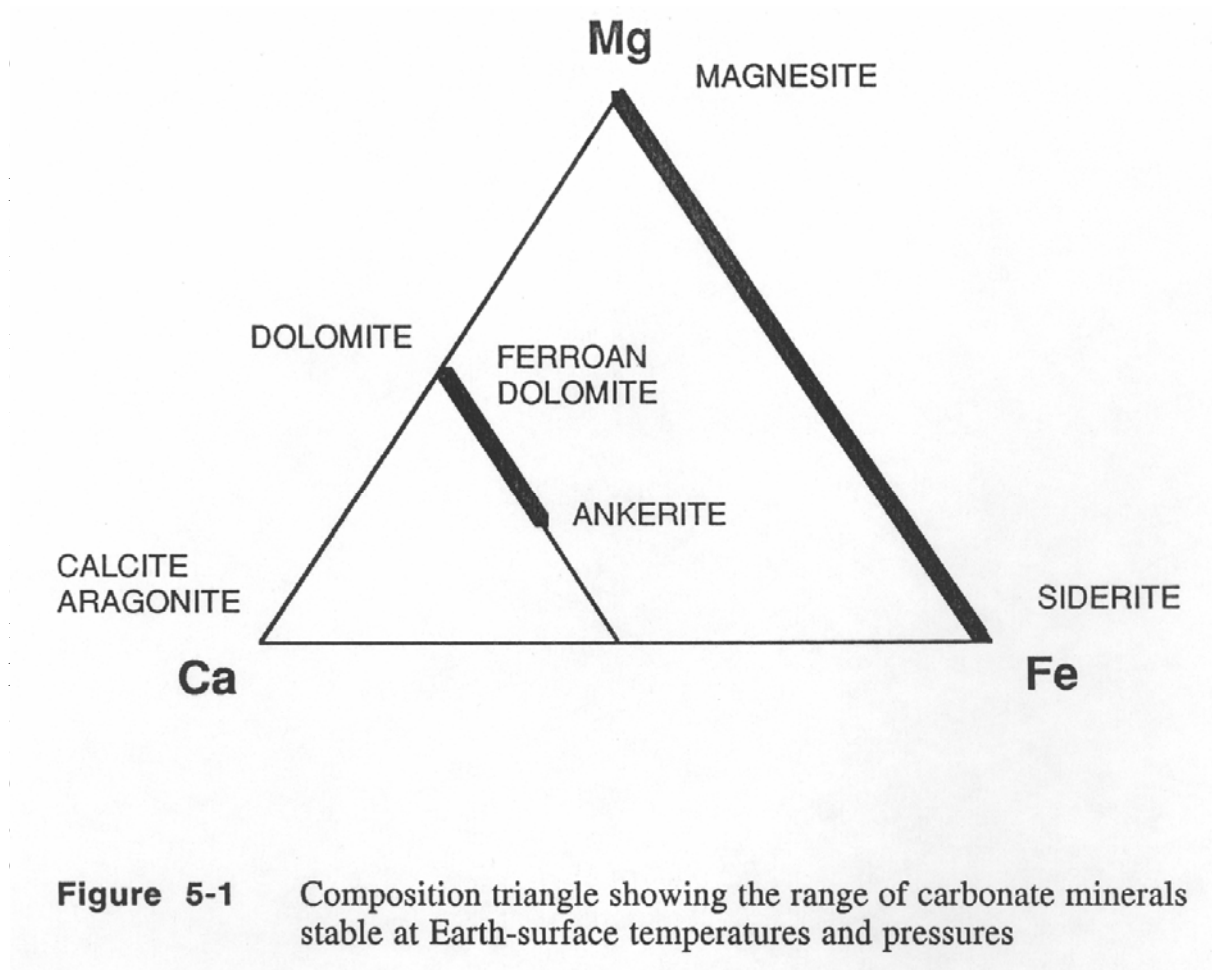


Figure 5-1 Composition triangle showing the range of carbonate minerals stable at Earth-surface temperatures and pressures

2.5 Most of the calcite precipitated by marine organisms contains a certain percentage of magnesium. Such calcite is called *magnesian calcite*; it's subdivided into *low-magnesium calcite* and *high-magnesium calcite* at 4% MgCO₃ content. Generally *the more advanced the organism, the less magnesium in the calcite*. In the case of red algae, an important sediment producer, the percentage is as much as 25%. Magnesian calcite is unstable, and it eventually expels its magnesium, but that takes a long time.

2.6 The effective ionic radii of Fe²⁺ and Mg²⁺ are almost the same, so these two ions substitute for each other in any proportion. There is *complete isomorphism* between magnesite and siderite, and also between dolomite and the iron equivalent, called *ferroan dolomite*. So you should expect most dolomite to contain at least some iron. This is why dolostones are typically tan-weathering, in contrast to usually gray-weathering limestone.

2.7 The other calcium carbonate mineral, *aragonite*, has an entirely different structure with *orthorhombic symmetry*. Aragonite can't tolerate even a few percent Mg²⁺ or Fe²⁺, but it can take some Sr²⁺ and Ba²⁺, which have much larger effective sizes. Also, some SO₄²⁻ sulfate ions can replace the carbonate ions.

2.8 Aragonite is the high-pressure form of calcium carbonate, and calcite is the theoretically stable form under all sedimentary conditions. But *in most sedimentary environments aragonite is precipitated rather than calcite*. The exception is purely fresh-water inorganic precipitation, which is of minor sedimentological importance. *Supersaturation, warm water, and the presence of sulfate and magnesium ions* tend to promote precipitation of aragonite instead of calcite. Aragonite eventually reverts to calcite, but it sometimes takes a long time; where sealed in sulfate-rich impermeable rocks, aragonite has in some cases been found in rocks even as old as the late Paleozoic.

3. THE CARBONATE BUDGET OF THE OCEANS

3.1 Calcite is more stable in pure water than its polymorph aragonite, so if the carbonate system in the oceans obeyed thermodynamics, calcite rather than aragonite should be precipitated. But it's observed that *aragonite is precipitated instead*. In the case of inorganic precipitation of aragonite, probably this has to do with the presence of other kinds of ions, like sulfate and magnesium, in sea water. These ions act as *kinetic inhibitors* and serve to impede the growth of calcite. In the case of biogenic precipitation, *organisms also seem to disregard the laws of thermodynamics*, and deposit either calcite or aragonite, or both at the same time, or one at one time and one at another. Some details follow.

calcareous algae: *Halimeda* and other green algae, aragonite; *Lithothamnium* and other reef-making red algae, calcite

modern corals (hexacorals): aragonite

- Paleozoic corals** (rugose corals, tabulate corals): probably calcite, by good preservation of structures
- brachiopods, bryozoans, foraminifera**: calcite then and now
- echinoderms**: calcite then and now; large single-crystal skeletal components, very durable
- trilobites**: probably calcite, by good preservation of structures
- mollusks**: mostly aragonite, sometimes partly or entirely calcite

3.2 At present, *most of the open surface waters of the oceans, except at high latitudes, are about saturated or even supersaturated with respect to CaCO₃*, so there must be at least a broad inorganic control on precipitation of CaCO₃. Figure 5-2 shows the degree of saturation of aragonite and calcite with depth for both the Atlantic Ocean and the Pacific Ocean. It's not known how long things have been this way, but judging from the overall similarity of present deposits and past deposits, the situation has probably been the same at least back into the Precambrian. In the Precambrian, the saturation of seawater with calcium carbonate may have been the highest of all time, because of much higher levels of CO₂ in the atmosphere.

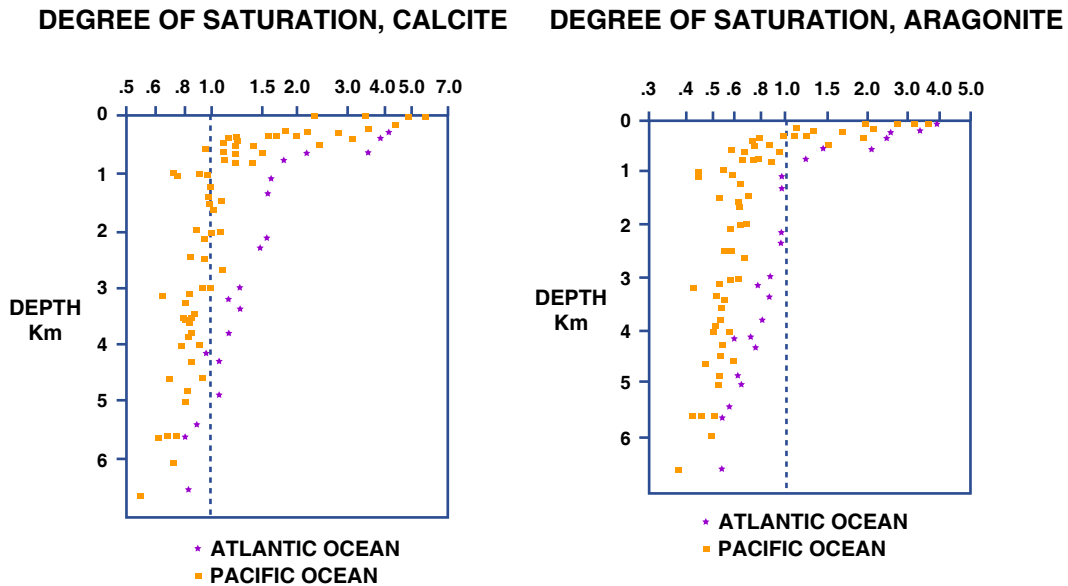


Figure by MIT OCW.

Figure 5-2: Degree of saturation with respect to calcite and aragonite with depth in the Atlantic and Pacific Oceans

3.3 By river contributions of dissolved load from rock weathering on the continents, *the doubling time of Ca²⁺ in the oceans should be about a million years*. (By the doubling time is meant the time it would take for the concentration of a given substance to increase by a factor of two if no other process is at work to remove the substance at the same time.) Obviously something must be taking it out at about the same rate. The same is true for Mg²⁺ ion, but *the doubling time is*

about an order of magnitude greater, something like twenty million years. In the case of calcium ion, the obvious process by which it's removed from solution is precipitation of carbonate minerals. For magnesium ion, however, the removal process is less obvious—because, as you will see presently, dolomite is being precipitated as a primary sedimentary mineral almost not at all in the modern oceans. It's known that Mg is removed from seawater very efficiently by reaction of the seawater with freshly crystallized basalt at and near mid-ocean ridges. Also, in shallow-water carbonate-producing environments the magnesium in solution is buried with the calcium carbonate sediments, where, with time, during shallow-burial diagenesis, it dolomitizes the calcium carbonate minerals—about which you will learn more in the later chapter on diagenesis.

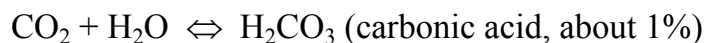
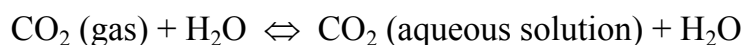
3.4 As for the entire ocean bottom, *about one gram of CaCO₃ is deposited per square centimeter of ocean bottom per 1000 years on the average*, on the basis of balance considerations. Over large areas, however, no carbonate is being deposited at all, whereas in other places the rates of accumulation are far greater than the average.

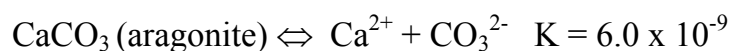
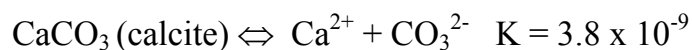
3.5 The *rate* of deposition of CaCO₃ is much greater in shallow ocean than in the deep ocean, but the *volume* of newly deposited shallow-water CaCO₃ is far smaller than that of deep-sea CaCO₃. But this is true only for the present time: planktonic carbonate-secreting organisms evolved late in geologic history, in the Mesozoic, so *the volume of shallow-water CaCO₃ must have been much greater in the Precambrian and the Paleozoic than now* (on the reasonable assumption that the rate of supply of dissolved calcium ion was about the same).

3.6 Another point worth noting here is that, from the standpoint of long-term geologic history, *deep-sea storage of carbonate is only temporary*: sea-floor spreading and subduction of oceanic lithosphere at continental margins provides a very satisfying way of reincorporating deep-sea carbonate deposits into the geological record of the continents, albeit usually in unrecognizable form. Before the plate-tectonics revolution in the 1960s, geologists believed that what was put into the deep ocean stayed there forever!

4. GEOCHEMISTRY OF CARBONATE PRECIPITATION

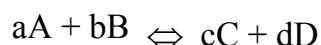
4.1 Precipitation of carbonate in natural waters is more complicated than that of, say, halides or sulfates, because of the *dissolution of carbon dioxide in natural waters*. Here are the reactions that are relevant to carbonate precipitation:





BACKGROUND: EQUILIBRIUM CONSTANTS

You probably can recall from some earlier chemistry course that for a chemical reaction



where, if the reaction goes to the right, A and B are called the *reactants* and C and D are called the *products* (whereas if the reaction goes to the left, A and B are the products and C and D are the reactants), and the coefficients a, b, c, and d are the numbers of atoms or ions involved in the reaction, the equilibrium constant K is

$$([A]^a[B]^b)/([C]^c[D]^d)$$

where the square brackets signify the concentrations (technically, the activities, but in dilute solutions the two are not greatly different) of the various substances.

If the equilibrium constant for a given reaction is small, that means that at equilibrium (that is, the reaction is proceeding just as fast to the left as to the right, so the concentrations of the various substances remain the same through time) the concentrations of the substances on the right side of the reaction are smaller than the concentrations of the substances on the left side of the reaction.

4.2 There are various ways of adding these reactions, but the way that's most relevant to carbonate precipitation in the oceans is



So

*anything that **increases** the concentration of dissolved CO₂ tends to cause **dissolution** of calcium carbonate,*

and

*anything that **decreases** the concentration of dissolved CO_2 tends to cause **precipitation** of calcium carbonate.*

The two most important effects are:

temperature: as the water temperature increases, the equilibrium solubility of CO_2 decreases, so as sea water is warmed there is a tendency for CO_2 to be released back into the atmosphere and for CaCO_3 to be precipitated.

photosynthesis: in photosynthesis, plants take up CO_2 from the environment and fix it in organic compounds in their tissues, thereby releasing oxygen.

4.3 Another important factor is this: carbonate-secreting marine invertebrates live in greatest numbers in the warm shallow parts of the oceans. These organisms can secrete CaCO_3 even from water that is not saturated in CaCO_3 , but they do it best and most abundantly where the water is saturated.

4.4 So it makes sense that *most of the CaCO_3 precipitated in the oceans today is in warm, shallow water*, where the water is warmed, so that the concentration of dissolved CO_2 is lowered and saturation with respect to CaCO_3 is thus enhanced, and where aquatic plants (largely algae) flourish.

4.5 In summary, *there is a broad inorganic control on carbonate precipitation in the oceans, but the specific controls have to do with local water temperature and photosynthesis.*

4.6 So far in this section we have addressed only the precipitation or dissolution of the calcium carbonate minerals. How about the mineral dolomite? To deal with precipitation of dolomite, we have to think about *undersaturation* and *oversaturation* (also called *supersaturation*). Suppose you put a piece of calcite in a beaker of distilled water, and for the sake of simplicity you arrange that no carbon dioxide is dissolved in the water. You know what will happen: the calcite dissolves slowly, and, as it does, the concentrations of dissolved Ca^{2+} ions and CO_3^{2-} ions increase. The solution is said to be undersaturated with respect to calcite. Eventually (and it would take months), the reaction reaches equilibrium, whereupon the concentrations of the Ca^{2+} ions and the CO_3^{2-} ions reach constant values. If, however, we somehow pump Ca^{2+} ions into the solution, we drive the reaction toward precipitation of calcite. The solution is said to be oversaturated.

4.7 We can think separately about the saturation state of calcite (or aragonite) in seawater and about the saturation state of dolomite in seawater. The conventional way of doing that is to define a quantity variously called the *solution quotient* or the *mass action quotient* or the *ion activity product* (the last designated IAP), which has exactly the same form as the equilibrium constant but the concentrations are those that exist at a given time in the solution, whether the

solution is in equilibrium (the state of saturation) or undersaturated or oversaturated. If the solution is oversaturated, then the IAP is greater than the equilibrium constant K ; if the solution is undersaturated, then the IAP is less than the equilibrium constant K .

4.8 It turns out that in warm, shallow seawater (the part of the oceans that is most relevant to precipitation of carbonate minerals, as you saw above), *all three of the important carbonate minerals—calcite, aragonite, and dolomite—are in a state of oversaturation*: that is, the situation should be conducive to precipitation. You have seen that calcite or aragonite are indeed precipitated, but dolomite is not. Yet, perhaps surprisingly, the ratio IAP/ K for the three minerals is as follows:

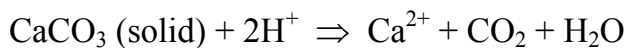
Aragonite: 2.1

Calcite: 3.3

Dolomite: 55

4.9 *Seawater is far more oversaturated with respect to dolomite than it is with respect to aragonite and calcite!* How, then, can we account for this seemingly paradoxical situation? We need to think not just about the *equilibrium* of chemical reactions but also about the *kinetics* of chemical reactions. Ionic crystals are built ion by ion, at some rate. Building of crystals of minerals with a greater degree of ionic ordering is inherently slower—more difficult—than that of crystals with a lesser degree of ordering. Dolomite has a much more highly ordered structure than does calcite or aragonite, inasmuch as in dolomite, in contrast to the two calcium carbonate minerals, the calciums and the magnesiums need to be in a regularly alternating array. *Reaction kinetics is the basic reason why dolomite, despite its great oversaturation, is not precipitated from seawater: it loses out in the kinetic competition.* And this must have been true in the geologic past as well as at present.

4.10 Another way of looking at dissolution of calcite is to write a reaction that pertains to rapid dissolution in the presence of hydrogen ions:



What's going on here is that the calcite dissolves in acidic waters to release calcium ions and carbon dioxide gas. That's what happens when we add ground-up limestone to our lawns and gardens to decrease the acidity of the soil. It's also what happens when we use our little bottle of dilute hydrochloric acid in the field to test whether a given rock is a limestone. When I was an undergrad, I worked in my advisor's paleontology laboratory, where we dissolved large blocks of limestone containing silicified fossils by painting the bottom surface of the block with a plastic resin and then putting the block in a tub of dilute hydrochloric acid in a fume hood. After the entire block was dissolved, we rinsed the residue and picked the silicified fossils with tweezers under a microscope. The topic of silicification will appear in a later chapter. (As a final note, the above reaction can also be used to account for the gradual dissolution of marble monuments and gravestones by acidic rainwater.)

5. MODERN MARINE CARBONATE SEDIMENTS

5.1 Carbonate sediments are forming today on many parts of the ocean bottom, according to definite but complex controls. Presumably this has been so throughout much of geologic history, but because organisms do most of (or, at least, are ultimately responsible for) the deposition, the controls must have changed through time as the organisms evolved.

5.2 Carbonate deposits are most abundant between about 30° N and 30° S. (They are not restricted to that zone, though: the topic of cold-water carbonates is an active area of research nowadays.) The low-latitude regions are where the surface waters tend to be saturated with respect to CaCO₃, and where the water is warm enough the year round for carbonate-secreting organisms to flourish. But this picture is obviously complicated in detail, by patterns of ocean currents, nutrient supply, and dilution by siliciclastic sediments.

5.3 There are *three main classes of modern marine carbonate sediments*:

- calcareous oozes in the deep ocean
- carbonate buildups
- calcareous sand and mud on platforms (shelf and ramp carbonates)

6. CALCAREOUS OOZE

6.1 More than a third of the present deep ocean bottom is covered with sediment containing more than 30% CaCO₃. The carbonate in these sediments is in the form of tiny shells or tests of various *carbonate-secreting planktonic organisms* that live in the warm shallow waters above. Such deposits are called *ooze*; there is *calcareous ooze* and *siliceous ooze*, depending upon what the organisms secrete.

6.2 The most abundant kind of calcareous ooze is *foraminiferal ooze*. Foraminifera are single-celled protozoans. In the oceans of today there are about thirty species of foraminifera (forams, for short) belonging to two families, Globigerinidae and Globorotaliidae. These are the only two of more than fifty foram families that are adapted to a *planktonic* rather than a *benthic* mode of life. But these planktonic forams grow in enormous numbers in the warm shallow waters of the ocean. At subdividing time the protoplasm of each foram subdivides into zoospores that swarm out to develop into new forams, leaving the empty test to sink to the bottom. So *very little organic matter goes down with the tests*. Figure 5-3: (left) Brasier, M.D., 1980, *Microfossils*: George Allen & Unwin, 193 p. (Figure 13.27g, p. 116). (center) Brasier, M.D., 1980, *Microfossils*: George Allen & Unwin, 193 p. (Figure 13.27e, p. 116) (right) Boersma, A., 1978, *Foraminifera*, in Haq, B.U., and Boersma, A., *Introduction to Marine Micropaleontology*: Elsevier, 376 p. (Appendix I, p. 69-75) shows what some planktonic forams look like.

6.3 Because of the greater solubility of CaCO₃ in the colder deeper waters, which are derived from cold polar regions and thus contain more dissolved CO₂, *below a certain depth the foram tests are completely dissolved before they have a chance to be covered by more tests*. Figure 5-4 is a graph of CaCO₃ content of bottom sediments vs. depth.

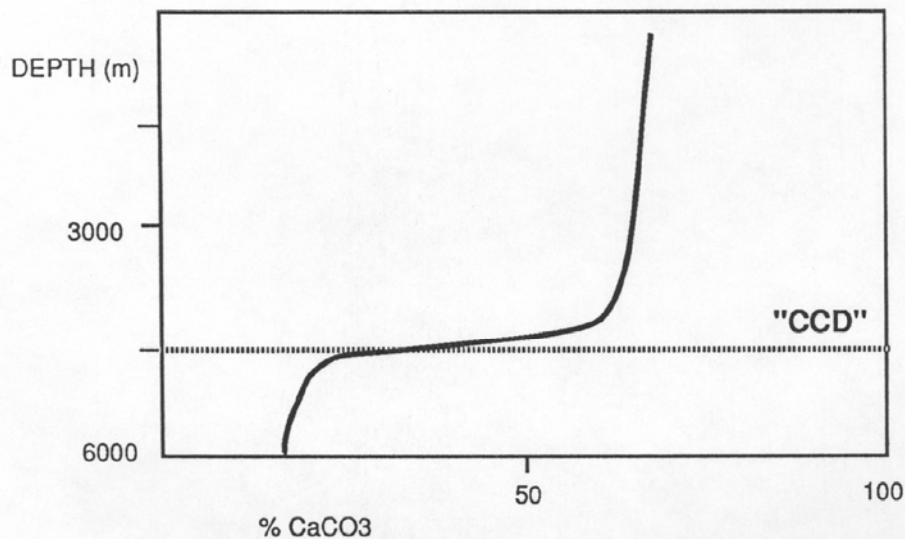


Figure 5-4 CaCO₃ content of marine bottom sediments vs. depth to ocean bottom

6.4 Calcareous oozes are abundant down to about 4000 m, but below that they become much less abundant; there is very little carbonate below about 4500 m. Oceanographers call *the depth between 4000 and 4500 m where carbonate becomes less abundant the **carbonate compensation depth**, or CCD*; it's actually a narrow zone rather than a single depth.

6.5 The oceans are undersaturated with respect to CaCO₃ below about 500 m everywhere; the reason carbonate is present down to thousands of meters is basically a *time-lag effect*. Presumably the CCD has to do with *what happens on the bottom, not what happens on the way down*, because it doesn't take long for the tests to get to the bottom, but they sit there exposed to the water for a long time.

6.6 The distribution of calcareous ooze is very irregular. Calcareous ooze is present mainly on high areas in low latitudes, and it's present to greater depths where surface productivity is high. There's not much under areas of low productivity, where nutrient concentrations in the surface waters are low, or at high latitudes, where surface waters are unfavorable.

6.7 Forams are the major constituent of calcareous ooze, but not the only one. Two other kinds of carbonate-secreting planktonic organisms also contribute to calcareous ooze:

- a minor group of tiny gastropods, called ***pteropods*** and ***heteropods*** (mostly the former); these are larger than forams but they secrete aragonite, so they are not found below about 3500 m;
- two kinds of calcareous algae, called ***coccoliths*** and ***rhabdoliths*** (mostly the former).

6.8 In a few areas pteropods or coccoliths predominate in the ooze (then the sediment is called *pteropod ooze* or *coccolith ooze*), but usually they are just a minor admixture in foram ooze. Figure 5-5: (left) Brasier, M.D., 1980, *Microfossils*: George Allen & Unwin, 193 p. (Figure 8.3a, p. 48). (upper center) Brasier, M.D., 1980, *Microfossils*: George Allen & Unwin, 193 p. (Figure 8.3c, p. 48). (lower center) Haq, B.U., 1978, *Calcareous nannoplankton*, in Haq, B.U., and Boersma, A., *Introduction to Marine Micropaleontology*: Elsevier, 376 p. (Fig. 13, p. 91). (right) Herman, Y., 1978, *Pteropods*, in Haq, B.U., and Boersma, A., *Introduction to Marine Micropaleontology*: Elsevier, 376 p. (Fig. 3, p. 154) shows what pteropods and coccoliths look like.

6.9 Calcareous ooze is a tan-colored to cream-colored sediment, gritty and clean-feeling; it's not at all as repulsive as its name suggests. In the deep ocean it grades laterally into the noncalcareous brown clay of the deepest ocean.

7. PLATFORM CARBONATES

7.1 Introduction

7.1.1 In many places in the world today, *carbonate sediments are accumulating on platforms along continental margins where siliciclastic deposits are absent*. The two most important factors that control this accumulation are

- lack of siliciclastic input, and
- high biogenic carbonate productivity.

Unfortunately *the range of sedimentary environments represented by such areas is less extensive than must have prevailed at many times in the geologic past*. That makes it more difficult for carbonate sedimentologists to interpret ancient carbonate rocks than it is for siliciclastic sedimentologists to interpret ancient siliciclastic rocks—because they don't have a wide a range of modern sedimentary processes to use as a basis for interpreting the ancient rocks.

7.1.2 Because the precipitation of carbonate is easiest in warm, shallow seawater, *most carbonate production takes place on tropical platforms within a fairly restricted range of shallow subtidal water depths*. These environments have come to be called, picturesquely, by carbonate sedimentologists *the carbonate factory*. Although the most of the sediments that are produced in the carbonate factory remain in the source area, some are transported landward and some are transported basinward (Figure 5-6). Thus, there are three zones of accumulation:

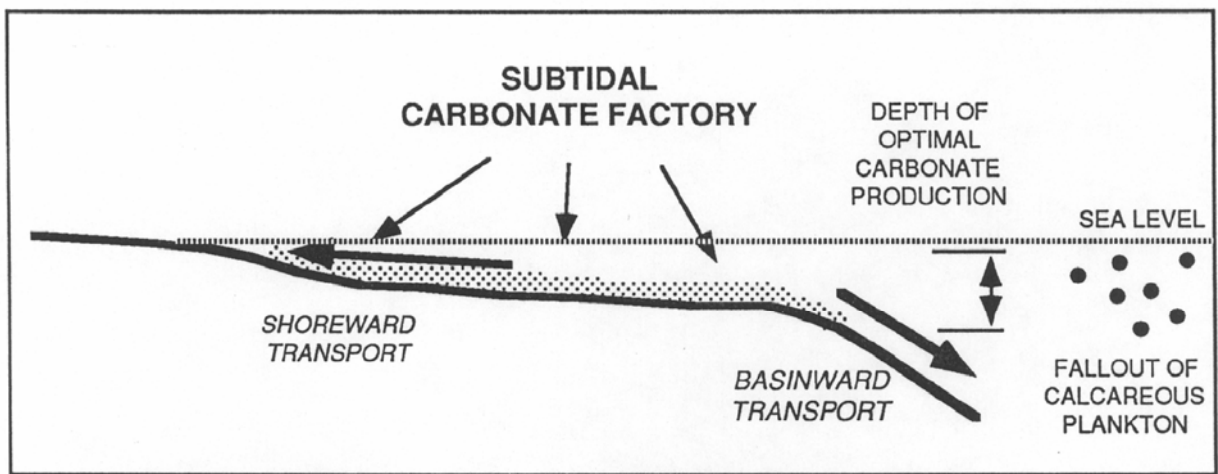


Figure 5-6 Carbonate production on shallow tropical banks

- the *subtidal open shelf and shelf margin*, characterized by in-place accumulations of carbonate sands, carbonate muds, and reefs;
- the *shoreline*, where sediments are transported from the open shelf onto beaches and tidal flats; and
- the *slope and basin*, where shelf-edge sediments are transported seaward, often by mass movements, and redeposited at depth.

7.1.3 *The shallow-water carbonate factory is very sensitive to sea-level change.* At various times in Earth history, sea level has fluctuated, on time scales of tens of thousands to hundreds of thousands of years, and with magnitudes ranging from meters to hundreds of meters. Most of the major carbonate-secreting organisms flourish when the water is shallow. If sea-level rise is slow, and the concomitant increase in water depth is slow, the carbonate factory can keep up production; this is called **keep-up** mode. But if sea level increases fast enough, and water depth thereby increases fast enough, the carbonate factory has a strong tendency to shut down. This is called **give-up** mode.

7.2 Regional Geometry

7.2.1 The term **platform carbonate** is in common use for *all accumulations of carbonate sediments in tectonically stable shallow-water environments*. This includes reefs, but I'll deal with them later. Because in areas of high productivity carbonate sedimentation can be so rapid, *carbonates tend to mold their own environment*, even on the scale of entire shelves. So *the regional bathymetry of carbonate areas tends to be more varied than that of siliciclastic shelves*, which are more familiar to most geologists. Below is an outline of the regional geometry of shallow-water carbonate bodies. Refer to Figures 5-7, 5-8, 5-9, and 5-10.

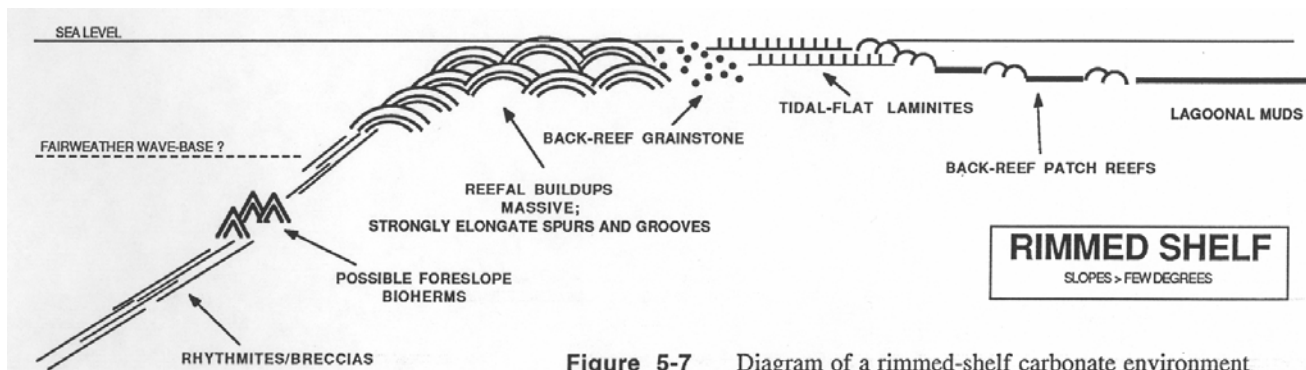


Figure 5-7 Diagram of a rimmed-shelf carbonate environment

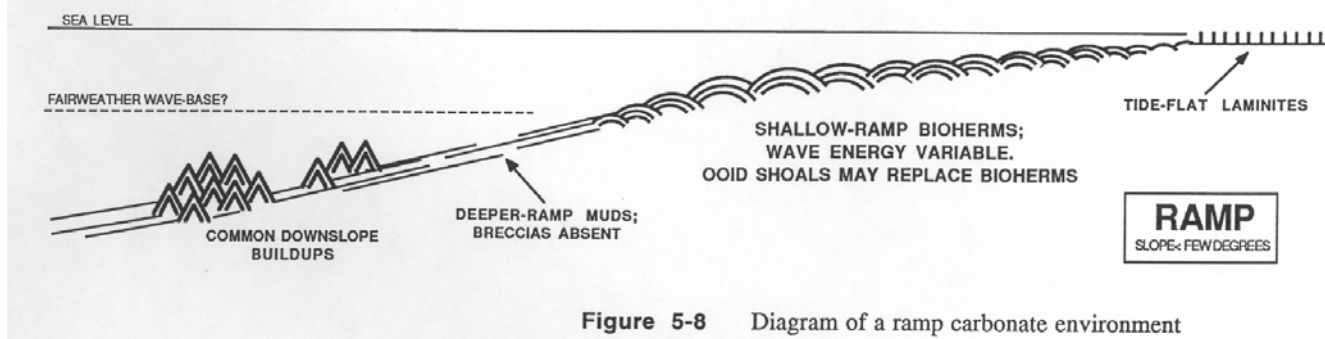


Figure 5-8 Diagram of a ramp carbonate environment

ROCKNEST PLATFORM

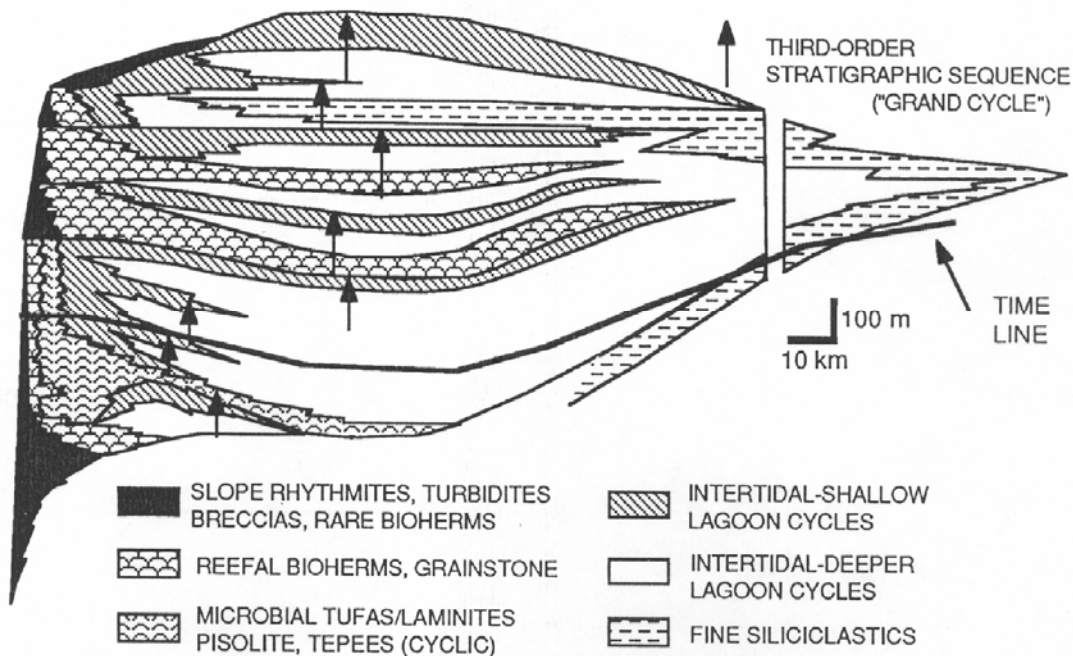


Figure 5-9 Regional stratigraphic cross section showing facies of the Rocknest Platform

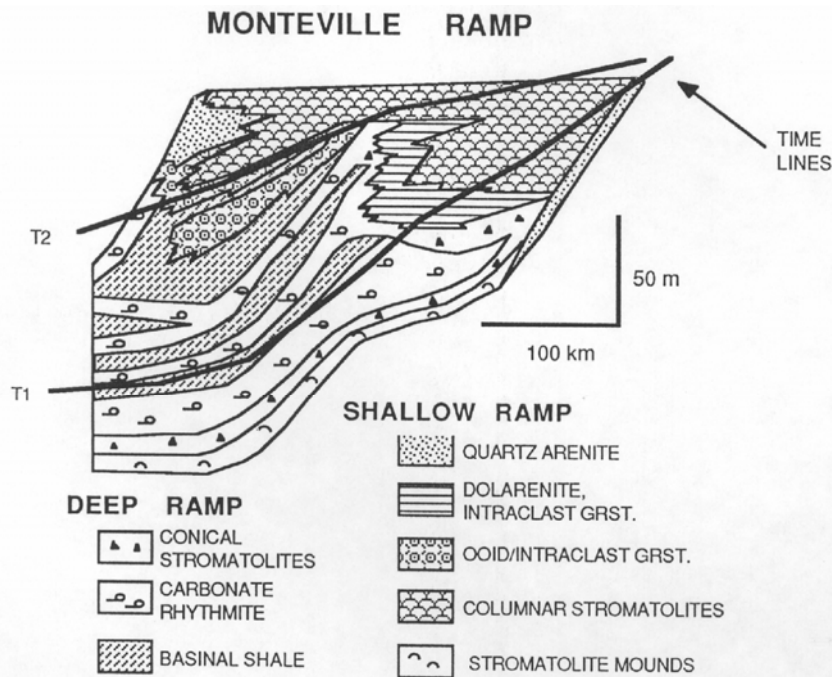


Figure 5-10 Regional stratigraphic cross section showing facies of the Monteville Ramp

7.2.2 Carbonate ramps are major carbonate bodies built far outward from land to have gentle regional sea-floor slopes. **Carbonate shelves** are large flat-topped carbonate bodies extending seaward from land areas. Ramps tend to evolve into platforms where carbonate productivity is high. **Isolated platforms** (or offshore carbonate banks) are large carbonate accumulations built up over local high areas offshore of continental land areas; they typically develop approximately flat tops and steep margins. Carbonate workers use the term **shelf** a little differently from others to be the flat upper surface of carbonate platforms or banks. Carbonate buildups are smaller-scale or more local carbonate bodies showing topographic relief; they are commonly present at shelf margins, but they are formed in other places as well. They usually result from the rapid growth of a community of calcareous organisms and include the well-known group of coral reefs.

7.3 The Bahamas

7.3.1 Few places in the world today are very representative of the vast areas of shelf carbonate deposition at various times in the geologic past. Some of the carbonate-producing regions are compared in Figure 5-11. An excellent place to study shelf carbonates today is on the Bahama Banks. The Bahamas are an extensive and outstanding example of warm, shallow, extensive carbonate-depositing seas. The following is a brief account of the Bahama Banks.

COMPARISON OF MODERN ENVIRONMENTS

	BAHAMAS	FLORIDA	SHARK BAY	PERSIAN GULF
MORPH.	Oceanic Platform Flat Top, Steep Margin	Pericratonic Platform Flat Top, Steep Margin	Pericratonic Platform Carbonate Ramp, Gently Sloping into Deeper Water	Intracratonic Platform (Foreland Basin) Carbonate Ramp, Gently Sloping into Deeper Water
TIDE (m)	0.8 to 0.0	0.7 to 0.15	1.7 to 0.6	1.5 to 1.0
WATER TEMP (Deg. C)	22 to 31	15 to 33 (Reef) 15 to 40 (Bay)	21 to 25 (Open Areas) 15 to 30 (Shallows)	20 (Winter) 36 (Summer)
RAINFALL (cm/yr)	100 to 150 Wet, Humid	Wet, Humid	20 Semi - Arid	20 (North) 5 (South) Arid
EVAP. (cm/yr)			200	350 to 800
CLASTIC INFLUX	None	Minor	Small Deltas	Major Deltas in North
SALINITY (o/oo)	36 to 38 46 (Near Andros)	32 to 38 (Reefs) 10 to 50 (Bay)	36 to 60 Minor Evaporites	36 to 70 Evaporites
BUILDUPS	Platform Edge and Bryozoan Reefs	Platform Edge Reefs and Mud Mounds in Bay	Barrier and Fringing Banks No Reefs	Shallow to Deeper Water Banks Minor Reefs

Figure 5-11 Comparison table of modern carbonate environments

7.3.2 The Bahama Islands are located on several submerged shallow carbonate platforms that lie just off the North American continental shelf. The banks cover 60,000 square miles, but the land area of the islands is only about 4400 square miles. *Water depths over most of the banks are less than ten meters!* There are several large islands and thousands of very small islands called cays. The exposed land surface is very pure Pleistocene limestone with little soil development. In places elevations are over a hundred meters; these are fossil subaerial dune ridges of oolitic sediment. Keep in mind that *the entire surface area of the banks was exposed during the low stands of sea level during the Pleistocene.*

7.3.3 The largest of the Bahama Banks, Great Bahama Bank, is split by two troughs, Tongue of the Ocean and Exuma Sound, which extend down to deep ocean depths, and each of the banks is separated from its neighbors by deep-water channels. The shelf margins are well defined. The upper few hundred meters of the slope is very steep, greater than the angle of repose; this is probably controlled by reef development in the past.

7.3.4 The Bahamas lie in the belt of northeast trade winds. This governs the position of islands on the windward margins of the platforms. Hurricanes are common, and they do important geological work. Throughgoing currents are warmed as they pass over the banks, leading to supersaturation with respect to calcium carbonate.

7.3.5 Bahama *reefs* play a minor part in the carbonate sediment picture today, presumably because of the Pleistocene history of fluctuating sea level. But deep borings show that they were much more important in the past. Reef construction is most important along the outer edges of the banks facing deep ocean water, and they are much better developed on the windward sides than on the leeward sides.

7.3.6 Great deposits of pure calcium carbonate sand and mud are forming and accumulating on the banks. With stable sea level, the carbonate sediment is transported to deep water as fast as it is produced, but *slow crustal subsidence has caused about 5 km of carbonate sediment to have been deposited since at least the early Cretaceous*, so the banks date from not long after the opening of the Atlantic.

7.3.7 All of the various kinds of shelf-carbonate constituents described in the next section are found on the Bahama Banks, most of them in abundance. Areas of ooids and of aragonite muds in particular are shown on the sketch map in Figure 5-12.

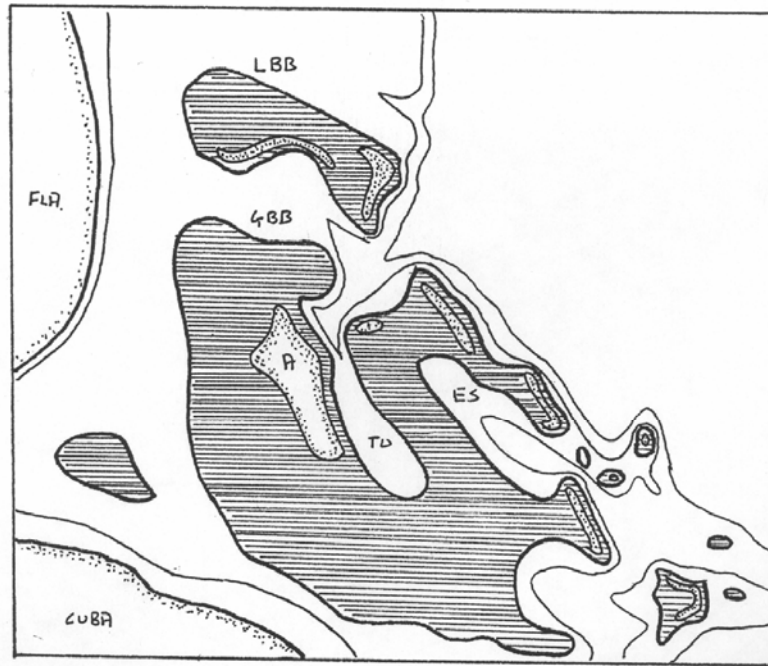


Figure 5-12 Sketch map of the Bahama Banks. Horizontal ruling shows banks; stippled areas show major islands.

7.4 Constituents

7.4.1 A great variety of sedimentary carbonate particles are produced in carbonate depositional environments. Here's a summary. I'll leave the broader patterns of depositional facies aside for now.

Skeletal Grains

Many kinds of marine invertebrates precipitate calcium carbonate to form their skeletons. There is an enormous range in size, shape, internal structure, and composition. Crystal size within the skeletal material ranges from microscopic to large single crystals. Some is aragonite, and some is calcite; the calcite itself ranges from magnesium-free to as much as 25% Mg (in the case of red algae). Grain shape depends on the skeletal geometry of the particular species, and, in the case of colonial organisms, on the style of colonial development as well. In the words of Dunham (1962), "Carbonate grains are shaped more like twigs or potato chips than marbles". Some of this skeletal material is buried in place; some is transported, and can undergo various degrees of breakage and wear. I won't even attempt to illustrate the enormous range of geometry and structure of skeletal grains; you will be seeing many in both hand specimen and thin section.

Ooids

Ooids are nearly spherical grains consisting of a grain of calcareous or noncalcareous material serving as a nucleus around which successive layers or shells of calcium carbonate are precipitated or accreted while the particle is moved in flowing water that is supersaturated with respect to calcium carbonate Figure 5-13: (left): Tucker, M., 1991, *Sedimentary Petrology; An Introduction to the Origin of Sedimentary Rocks*: Blackwell, 260 p. (Figure 4.2, p. 110). (right): Williams, H., Turner, F.J., and Gilbert, C.M., 1982, *Petrography; An Introduction to the Study of Rocks in Thin Section*: W.H., Freeman, 626 p. (Figure 14-8, p. 385). Size ranges mostly from a fraction of a millimeter to about 2 mm—although, especially in the Neoproterozoic, they can be on a centimeter scale. If the coating is thin relative to the size of the nucleus, the ooids are called ***superficial ooids*** or ***coated grains***. Commonly there are two kinds of concentric spherical shells in the ooid structure: tangentially oriented aragonite needles, and non-oriented cryptocrystalline aragonite. Ooids tend to contain organic matter in the form of algal mucilaginous matter; this is seen as dark brown areas in thin section. Ooids can also show radial magnesian calcite. This seems to result from purely inorganic precipitation. Ooids are common in the ancient and are known but not common in modern carbonate sediments. The processes of ooid growth are still a subject of discussion.

Pellets

Pellets are *rounded grains of very fine-grained aragonite and calcite*, a few tenths of a millimeter to about a millimeter in size. They are usually elliptical or ovoid in shape, but they may be broken to form beehive-shaped grains. Some are clearly *fecal pellets* excreted by such organisms as worms, gastropods, mollusks, and crustaceans. These are soft and friable at first, but they soon become well cemented. Other pellets seem to be formed by cementation and rounding of friable irregular aggregates of aragonite silt. Because it is usually difficult or impossible to distinguish among the various possible processes that form such objects, the term ***peloid*** is in common use.

Intraclasts

Intraclasts are *fragments of carbonate sediment, usually fine-grained, that was deposited and then later ripped up by strong currents to be redeposited with other carbonate sediment*. The derivation of the word implies that the ripping up took place within the environment of carbonate deposition, geologically soon after the depositional of the sediment; don't confuse these with carbonate rock fragments in a largely siliciclastic conglomerate. The stage of cementation varies considerably. Commonly the intraclasts are tabular, reflecting breakage of semi-consolidated sediment along stratification planes.

Carbonate Mud

Carbonate mud consisting mostly of needle-shaped aragonite crystals is common in areas with weak currents. Some of this carbonate mud is produced by abrasion of larger grains, but most seems to be precipitated directly from seawater. The nature of this precipitation has been controversial: *is it purely inorganic, or is it caused by algae?* (You can imagine precipitation of aragonite next to the bodies of photosynthesizing algae, because the CO₂ content of the water right next to their bodies is decreased, which favors carbonate precipitation.) The answer seems to be that both processes operate, but that *algal precipitation is generally much more important than inorganic precipitation.*

8. REEFS

8.1 Introduction

8.1.1 A reef, rising above the sea floor, is an entity of its own making—a sedimentary system within itself. The numerous, large calcium-carbonate-secreting organisms stand upon the remains of their ancestors and are surrounded and often buried by the skeletal remains of the many small organisms that once lived on, beneath, and between them (Figure 5-14).

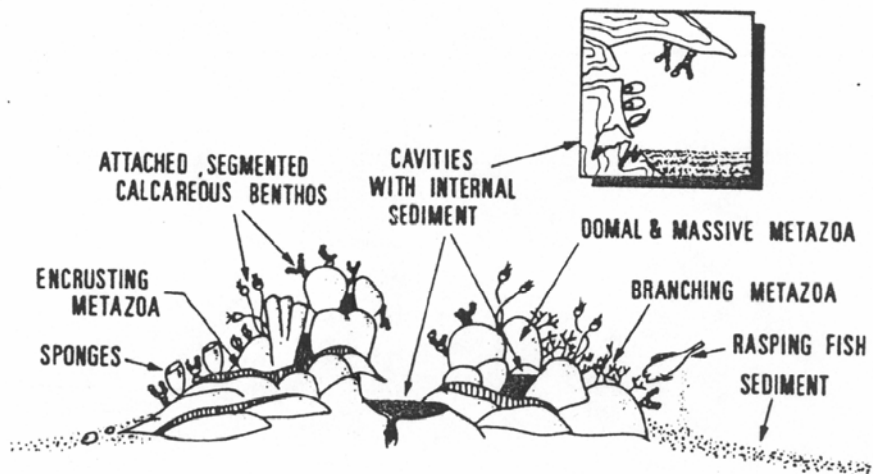


Figure 5-14 Detail of the arrangement of carbonate-secreting organisms on a typical reef

8.1.2 At present, far more shallow-water carbonate sediment is produced in, or in connection with, reefs than by any other means. Judging from the presence of reefs in the stratigraphic record, *reefs have been important sites of carbonate sedimentation from the Archean.* Throughout geologic history, they have

experienced various periods of importance and decline in terms of their absolute abundance (Figure 5-15). Furthermore, as organisms have evolved through time the record of this is reflected in the history of reefs that have been dominated by different communities at different times. However, in most cases, these organisms have had very similar functional morphologies.

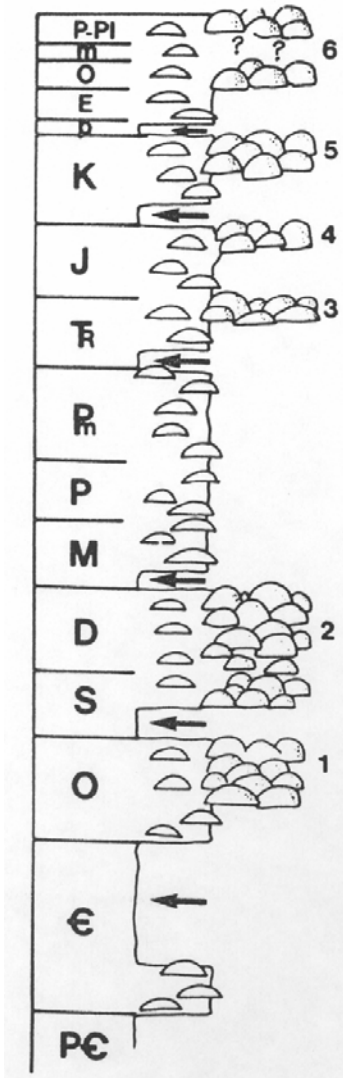


Figure 5-15 Abundance of reefs through geologic time

8.2 Terminology

8.2.1 In the most common sedimentological usage, the term *reef* is used for a marine structure built by organisms and with a framework strong enough to withstand attack by waves. But the usage is confused; some use the term for all kinds of carbonate buildups, the above kind being just one of many possibilities. And the nonscientific meaning of the term reef is just a shoal of any kind on which a ship can go aground. So it's best to preface the term with a modifier: *organic-*

framework reef is probably the best. And there are several kinds of such reefs, depending upon geographic setting.

8.3 General Stuff

8.3.1 Reefs are produced by growth of colonial carbonate-secreting organisms. But there's more to it than that. Marine life is abundant in warm shallow seas, but generally it has to adapt itself to the physical environment. But under certain conditions, certain kinds of organisms can create their own environment by building major structures that alter the local marine environment fundamentally.

8.3.2 Reefs need the strong waves and currents that act upon them: *without the strong water movements to clean them now and then of loose sediment and to provide them with nutrients from the open sea, reefs would not develop in the first place.* The geological significance of this is that fragmentation by the destructive forces of storms is a natural concomitant of reef growth.

8.3.3 Reefs have been of special interest to petroleum geologists because *reef rocks in the stratigraphic record have often ended up being petroleum reservoir rocks*: they tend to have high porosity, and they usually become encased in finer and much less permeable sediment, so they make good stratigraphic traps.

8.3.4 You probably think of reefs as being formed by corals; to many, the term "reef" is synonymous with "coral reef". Corals are indeed the chief reef-building organisms today, but calcareous algae are major contributors, and in the past algae, bryozoans, and archeocyathids were at times the major reef builders. It's better to think in terms of organic reefs than of coral reefs.

8.3.5 The *coral polyp*, the coelenterate that sits in and on a cup-shaped calcareous skeleton, feeds by filtering plankton. The polyps of reef-forming corals contain algae called *zooxanthellae*, mainly in the cells of the covering of the polyp. The polyps and the algae are *symbiotic*: the algae receive nutrients and CO₂ from the polyp, and the polyp receives oxygen from the algae. The polyps feed at night, and during the day they are partly closed to put the algae in the best position for photosynthesis.

8.3.6 Colonies of polyps develop by budding or fission. All of the polyps stay in contact with one another, so the colony is like a quasi-animal. The growth forms of coral colonies vary widely, even with the same genus: dome-shaped, mushroom-shaped, and branching in all sorts of ways.

8.3.7 Important: *not more than a few tens of percent of the reef volume is the rigid coral framework.* There are abundant tunnels and cavities in the framework that are partly filled with other calcareous biogenic sediment. Most reef sediment is produced by the post-mortem disintegration of organisms that are segmented (crinoids, calcareous green algae) or non-segmented (bivalves,

brachiopods, foraminiferans) and that grow in the many nooks and crannies between the larger skeletal metazoans. The rest of the sediment is produced by the various taxa that erode the reef: boring organisms (bivalves, worms, sponges) produce lime mud; rasping organisms that graze the surface of the reef (echinoids, fish) produce copious quantities of lime sand and silt.

8.4 Controls on Reef Growth

8.4.1 Reef growth is mainly a function of the factors that determine whether the coral colonies can flourish and increase rapidly in volume:

Water Temperature. Winter water temperatures are rarely below 18° C over reefs; corals don't grow in profusion until the temperature is 25–30° C. Temperatures above 30° C are above the optimum for coral growth, and temperatures above about 35° C kill corals. But temperatures above about 30° C are rare in the open ocean, so *the chief temperature factor is absence of cold water*. Reefs are rare outside the tropics, and they are more common in the western than in the eastern parts of all three major oceans, because of cold equatorward-flowing currents in the eastern parts.

Water Depth. The depth to which reef-forming corals can thrive is a function of *light penetration*, because the symbiotic algae need light for photosynthesis. Therefore *it's not depth alone that limits coral growth, but light penetration*. In exceptionally clear water a few species of reef-forming corals can live at 100 m, but for most the limit is 70–80 m. With the usual turbidity around coral reefs, reef corals don't flourish deeper than 50 m, usually less. Probably the depth at which the greatest volume of new coral sediment or rock is added per year is only a few meters below mean sea level. There are deep-sea corals that are independent of light, but these don't develop colonies and build reefs. The distribution of reef building also relates to water depth; platy forms develop at greater depths, massive and branching forms at shallower depths, and encrusting forms face the reef front.

Nutrients. The coral colony needs a steady supply of food. Plankton are most abundant where nutrients in solution are carried in by currents from the open ocean. Algae produce oxygen during the day, but corals need oxygen at night too, and that has to be transported in by currents as well. Currents are thus favorable to reef growth.

Salinity. More or less normal salinity is required for reef growth. Reef corals live within the salinity range 27 to 40‰. Reefs can be killed by great floods of fresh water sweeping over them from land.

Fine Sediment. Fine sediment restricts penetration of light, and it also hampers growth when it settles on the colonies. Probably the effect of light is more important than smothering, because corals can readily clean themselves off.

Also, *corals need a suitable substrate on which to develop*: coral colonies can't develop on a uniformly muddy floor.

8.5 Kinds of reefs

Fringing Reefs: these grow directly against a rocky coast. Width varies up to a few hundred meters. See Figure 5-16.

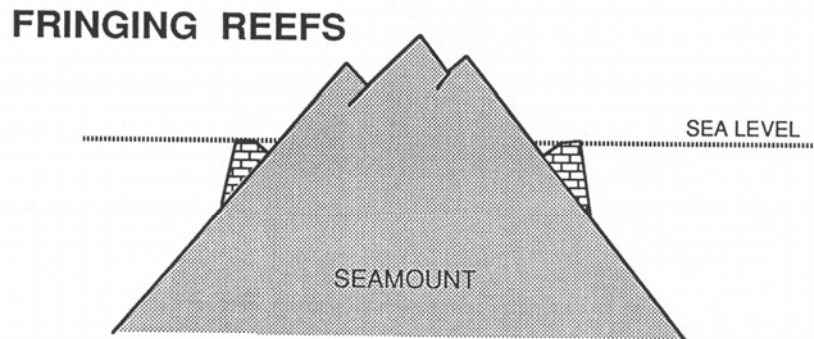


Figure 5-16 Cross-section sketch of a fringing reef

Atolls: ring-shaped reefs, unconnected with land. These presumably developed as fringing reefs around oceanic islands that later subsided while reef growth kept pace with subsidence. See Figure 5-17.

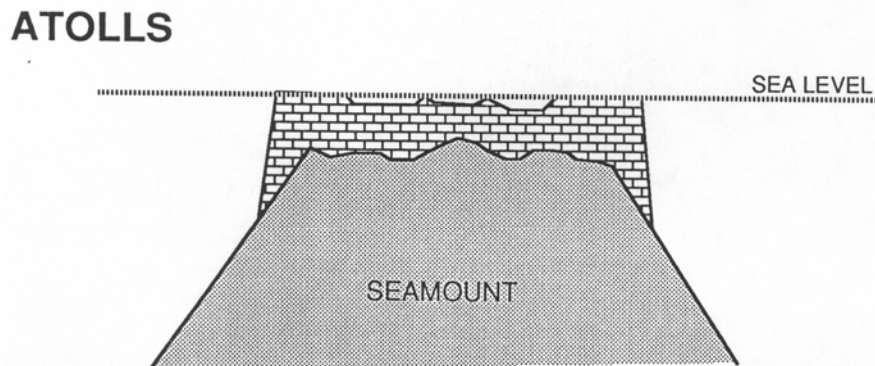


Figure 5-17 Cross-section sketch of an atoll

Barrier Reefs: these, the largest and most important, are reefs separated from the continental coast by a lagoon that is too deep for coral growth and up to a hundred kilometers wide, floored at least in part by siliciclastic sediment. (The Great Barrier Reef of Australia, following the coast of Queensland for 1500 miles and as much as 150 miles offshore, is of this kind.) See Figure 5-18.

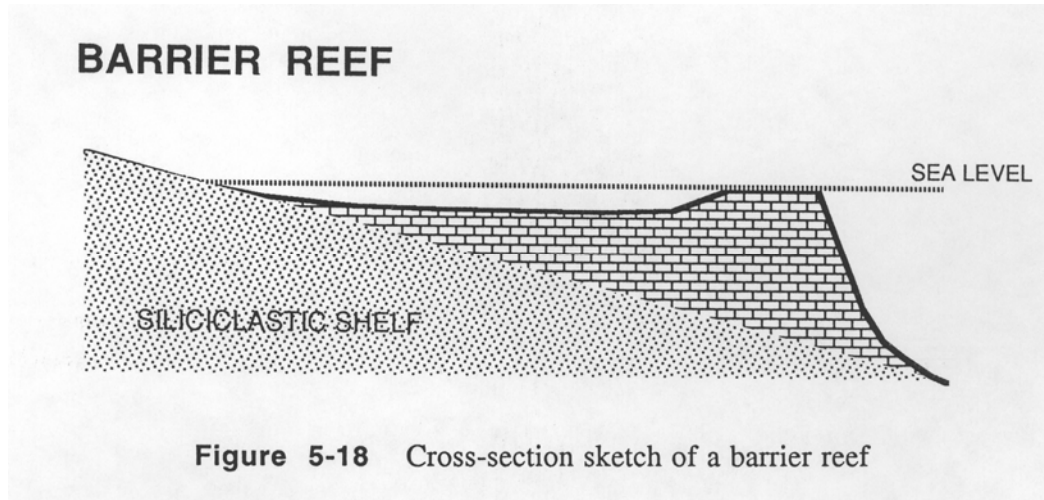


Figure 5-18 Cross-section sketch of a barrier reef

8.6 Morphology and Sedimentology

8.6.1 Figure 5-19 shows a typical cross section through a reef, normal to the reef front.

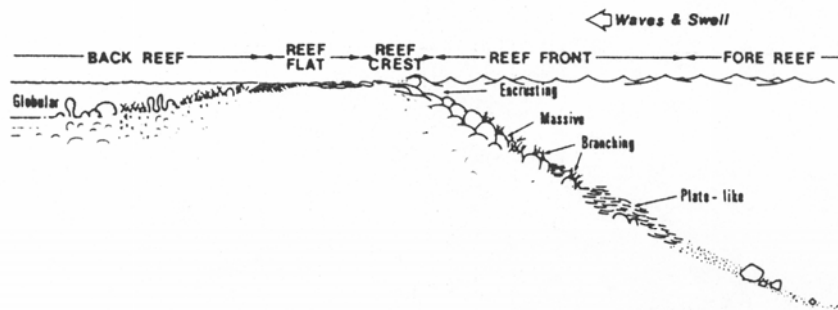


Figure 5-19 Typical cross section through a reef, normal to the reef front

Lagoon: The lagoon is *the relatively protected and shallow area behind the reef*. Smaller lagoons are floored entirely by finer carbonate sediment; lagoons behind barrier reefs can have mostly siliciclastic sediment. Lagoons often have numerous coral knolls, or isolated coral colonies, each a miniature reef itself. Depths in lagoons are seldom more than 50 m. There is usually a great variety of

sediment types in lagoons: typical sediments are coral debris from the living reef, foraminiferal and algal sediment, skeletal fragments of larger invertebrates, and aragonite mud in the deepest parts.

Reef Flat: Behind the growing reef front is *a broad expanse of dead reef rock with a flat surface*, partly or wholly emergent at low tide. There are patches of sand and coral rubble and scattered small coral colonies, as well as shallow pools, irregular gullies, and potholes. Islands are common on the reef flat; these could be built by storms or be left from a higher stand of sea level. Most of the major islands on reef flats today are a few meters above sea level and are undergoing erosion; they probably were formed by the higher stand of sea level during times of milder climate within the last few thousand years.

Reef Front: The heart of the reef is the growing reef front, *a narrow zone of flourishing coral colonies*. Waves and storms beating against the coral colony keep breaking material away, throwing it onto or over the reef flat or carrying it down the outer reef slope. *It's here that most of the reef growth takes place*.

Outer Slope: Often the reef descends almost vertically for a hundred meters, then at slopes as much as 30° down to deep ocean depths. The reason for the steepness of the slopes is that the coral colonies grow fastest in their uppermost part, tending to produce an overhang. The lower slopes are a jumble or talus of jagged broken reef material.

8.7 Bioherms and biostromes

8.7.1 Whenever you see large carbonate skeletal deposits in the sedimentary record, you have to think about whether they were part of a reef, or, more specifically, whether they formed a mass that stood up from the sea floor or whether they had no relief but built up in thickness along with the surrounding sea floor. See Figure 5-20.

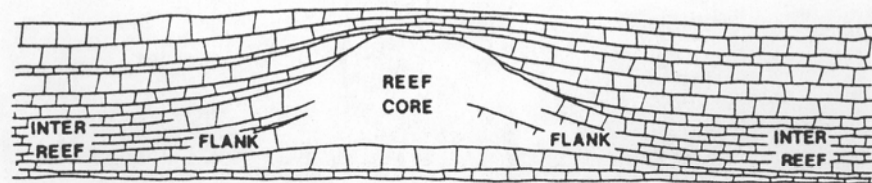


Figure 5-20 Schematic cross section through a carbonate buildup that presented positive relief on the sea floor

8.7.2 A *bioherm* is the general term for carbonate buildups, large or small, produced either by *in-place production of large numbers of individual organisms or by colonial framework building or encrustation*, as in reefs. These structures had substantial positive relief on the sea floor. They have low width-to-height ratios ($W:H < 30$)

8.7.3 A *biostrome* is the general term for *blanketlike deposits of carbonate sediment produced by in-place organisms and surrounded by other sediment types*. There are shell biostromes (beds of unsorted and nontransported skeletal remains that grew and died in place) and algal biostromes (soft, sticky mats of sediment-binding algae, which catch fine carbonate sediment and then grow upward through that sediment to reestablish the mat). They have larger width-to-height ratios ($W:H > 30$).

8.7.4 Finally, the general term in use for any carbonate body that develops in such a way as to stand above the surrounding seafloor is called a *buildup*. The essential feature of a buildup is that it has topographic relief above the seafloor.

9. LIMESTONES

9.1 Definition

9.1.1 The official definition of a *limestone* is a *sedimentary rock that contains at least 50% carbonate minerals, of which at least 50% is a calcium carbonate mineral*. Most limestones you will deal with have far more carbonate material than siliciclastic material: approximately equal mixtures are not as common as dominantly carbonate or dominantly siliciclastic—although approximately equal mixtures are not vanishingly uncommon.

9.2 General Stuff

9.2.1 *Limestones are a highly varied group of rocks*. All they have in common is their composition. Some are grown in place, as reefs, and others are deposited as particles, analogous to shale, sandstone, and conglomerate. Many geologists have thought of limestone as a "wastebasket" term, and have tried to get rid of it by coining new and more specific words for the various kinds--but they have not been successful.

The keys to understanding most limestones are these:

- *Most carbonate sediments started out as accumulations of particles, large and small, analogous to siliciclastic gravel, sands, and muds*. These particles are produced locally in the carbonate environment, and they are deposited with or without subsequent transportation. So you can think about most carbonate rocks in terms of *framework, matrix, and cement*, just as with sandstones, but keep in mind that most if not all the material is just carbonate in composition.

- *Cementation of carbonate sediments is usually very early; little burial is needed*. So reduction of pore space is much more a matter of filling by carbonate

cement than by deformation of framework grains. Cementation commonly occurs before compaction of sediments.

- *Carbonate rocks are much more susceptible to diagenesis, early and late, than siliciclastic rocks.* Diagenesis obliterates primary structures and textures in carbonate rocks long before siliciclastic rocks. This is due to their greater solubility in diagenetic pore fluids.

9.3 Classification

9.3.1 In the late 1950s, Folk revolutionized the classification of limestones by proposing that *most limestones can be treated as particulate rocks involving framework, matrix, and cement just like siliciclastic rocks.* Although the terms he coined for the various classificatory categories are not widely used these days, his ideas have been of great significance.

9.3.2 Folk called the coarse clastic constituents ***allochems***. These are the kinds described in an earlier section: *intraclasts, skeletal grains, ooids, and pellets.* He used the term ***micrite*** for the carbonate mud (short for *microcrystalline calcite mud*; something of a misnomer because it's now known that most of the mud is aragonite, not calcite) and ***spar*** for the *sparry calcite cement*. Figure 5-21 shows where most limestones fall in a ternary composition diagram. Figure 5-22 shows one version of Folk's classification, and Figure 5-23: Folk, R. L., 1959, Practical petrographic classification of limestone: American Association of Petroleum Geologists, Bulletin, v. 43, p. 1-38, Figure 4, and Figure 5-24: Folk, R.L., 1962, Spectral subdivision of limestone types, in Ham, W.E., ed., Classification of Carbonate Rocks: American Association of Petroleum Geologists, Memoir 1, p. 62-84. are two of Folk's original figures, showing about the same thing as above but in more detail.

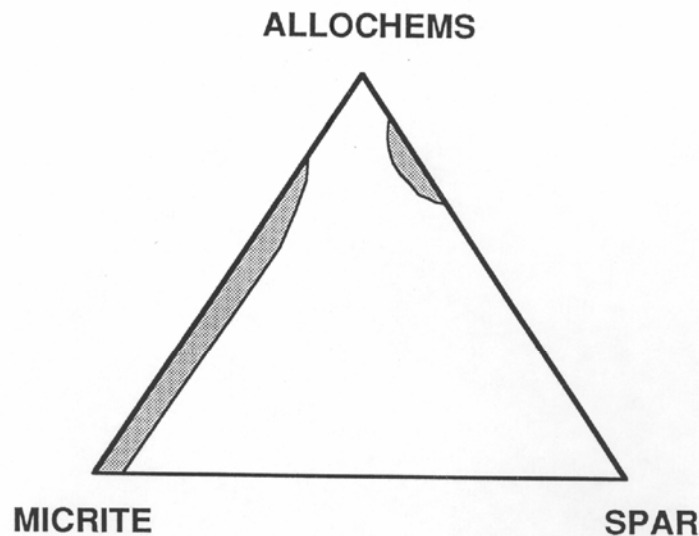


Figure 5-21 Ternary composition diagram showing regions of most common limestones

	Sparry rx	Micritic rx		
Intraclasts	Intrasparite	Intra micrite	Intraclasts > 25%	
Oolites	Oosparite	Oomicrite	Oolites > 25%	Intraclasts < 25%
Fossils	Biosparite	Biomicrite	F>P	
Milits	Pelsparite	Pelmicrite	P>F	
No allochems		Micrite		No allochems

Figure by MIT OCW.

Figure 5-22: Folk's classification of limestones

9.3.3 In the early 1960s Dunham developed a simpler and broader classification using Folk's concepts; the names are in very wide use today. Figure 5-25 shows Dunham's original classification table, and a later modification by Embry and Klovan.

Depositional Texture recognizable				Depositional texture not recognizable
Original components not bound together during depositions			Lacks mud and is grain-supported	
Contains mud (particles of clay and fine silt size)		Original components were bound together during deposition... as shown by intergrown skeletal matter, lamination contrary to gravity, or sediment-floored cavities that are roofed over by organic or questionably organic matter and are too large to be interstices.		
Mud-supported			Grain-supported	
Less than 10% grains	More than 10% grains			
Mudstone	Wackstone	Packstone	Grainstone	Boundstone
				Crystalline carbonate
				(Subdivide according to classifications designed to bear on physical texture or diagenesis.)

Figure by MIT OCW.

Figure 5-25: Dunham's classification of limestones

The key terms here are **grainstone**, **packstone**, **wackestone**, and **mudstone**. Get used to trying to apply these names when you look at a limestone. They are by far the most common terms used by practicing sedimentary geologists to describe limestones. Dunham's term **boundstone**, also important to remember, applies to a carbonate rock, typically a reef rock, whose original constituents were bound together during deposition and subsequently remained mostly in position of growth. A few others, however, not covered by Dunham's classification, are also common: a **rudstone** is a carbonate rock composed of bioclasts or other carbonate fragments, over 2 mm in diameter, that are close-packed and in physical contact. Rudstones are the coarse-grained equivalents of grainstones and packstones—terms that are customarily restricted to sand-size carbonate clasts. A **floatstone** is a carbonate rock containing a small percentage of bioclasts or other carbonate fragments over 2 mm in size, that are widely spaced, embedded in (“floating in”) finer carbonate sediment. Floatstones are the coarse-grained equivalents of wackestones. A **bindstone** is a kind of boundstone that is composed of sheetlike colonies that encrust large fossil fragments or finer carbonate sediment, forming a layered mass that is partly in-place skeletal, like boundstone, and partly bioclastic, like packstone or wackestone.

9.4 *Microbial Carbonates*

9.4.1 The foregoing material neglects a very important kind of carbonate depositional environment and the limestones and dolostones deposited therein: **microbial mats**.

9.4.2 Since far back in the Precambrian, *certain kinds of microbes (mainly filamentous and coccoid cyanobacteria) have developed extensive mats in very shallow marine environments*, ranging from very shallow subtidal up through intertidal. These mats present a dense network of growing filaments exposed to the surface. Fine carbonate sediment (aragonite needles, or dolomite crystals to which the aragonite has been changed within the sedimentary environment) is now and then washed onto the mat and adheres to the sticky surface. The filaments then grow upward among the sediment particles to reestablish the mat. The result is a laminated fine carbonate deposit. In some cases, the microbes cause precipitation of carbonate directly on the surface, which accretes upward over time.

9.4.3 Depending upon environmental conditions that are still not well understood, the surface geometry of the microbial mat can either be planar or nearly so (these deposits are called **microbial laminites**) or take on a variety of complicated geometries involving heads or domes of various shapes. These structures are called **stromatolites**. Lamination in these stromatolites is characteristically convex upward, although there are some striking exceptions to this. At the margins of the stromatolites the laminae dip at angles greater than the angle of repose, which is a tip-off for the cohesive or precipitated nature of the deposition. In some cases the laminae end abruptly at the margins of the

stromatolites. It was originally thought that stromatolites could be treated as actual organisms, and were given names like *Cryptozoon*, *Anabaria*, and *Tungussia*, based on variations in their form and lamination fabrics. Although such names are still used, it is recognized that the range of morphologies and lamination fabrics is due more to environmental factors than to biologic factors. Some stromatolites have poor lamination and a clotted texture; these are called **thrombolites**. Collectively, microbial laminates and stromatolites are termed **microbialites** (pronounced mike-ROW-bee-al-ites, not mike-row-BUY-al-ites!).

9.4.4 In slightly different environments, especially with strong and regular currents, *microbes bind or precipitate fine carbonate sediment around small spheroids that roll around while accreting concentrically*. These spheroids are called **oncolites** (also called **oncoids** or **microbial accretionary grains**). They are analogous to ooids, but they are different in origin, and usually they are larger. They are usually from a few millimeters to more than a centimeter in size. When larger, up to ten centimeters, they are called **microbial biscuits**.

9.4.5 Telling microbial laminites from non-microbial laminites in thinly laminated carbonates is a tricky business. Microbial laminites tend to show less regular lamination, and they sometimes show a characteristic structure known as **fenestrae**: *little irregular cavities elongated parallel to bedding*, filled with carbonate cement or sometimes even geopetal fine clastic sediment. These cavities seem to have developed during gas production in response to photosynthesis or respiration by the microbial community, or during occasional drying and shrinkage of the mat.

10. STRUCTURES IN LIMESTONES

10.1 Many limestones have bed configurations (ripples and dunes) and cross stratification much as in siliciclastic rocks. That should not surprise you, inasmuch as carbonate grainstones, especially, consist of particles that were transported by water currents. Certain other structures, however, are specific to limestones. Below are descriptions of two of these. (There are others that could be added here as well.)

hardgrounds: Hardgrounds are surfaces or horizons with syndimentary cementation that develops at, or slightly below, the sediment surface. It's a feature that develops underwater, on the seafloor. The surface is typically encrusted by sessile benthic organisms (that is, organisms that live attached to the seafloor). Species of various phyla of marine invertebrates have that mode of life. The sediment lying immediately below such surfaces also tend to be bored by organisms that make tunnels into the sediment. The burrows are manifested by slight differences in the nature of the sediment that fills the burrows. Hardgrounds are significant because they record times when the rate of sedimentation was low, or zero. They often mark unconformities.

paleokarst: You probably have heard of karst. Karst, a geomorphic feature, is a kind of topography that develops on limestone terranes by solution when the land surface in a region with abundant rainfall lies well above the groundwater table. The surface manifestation of karst is a distinctive pattern of sharp hills and bowl-shaped depressions, that latter commonly floored by sinkholes. Below the surface, karst is associated with networks of caves and caverns. In a limestone succession, karst (paleokarst, that is) can be recognized by the presence of a sharp, irregular surface that caps a zone of dissolution marked by solution breccias and irregular solutional passageways extending often far down into the underlying succession. The great significance of paleokarst is that it marks a time of emergence: a transition from a submarine depositional setting to a subaerial erosional setting.

11. STEPS IN STUDYING CARBONATE ROCKS

- Is it a limestone or dolostone?
- Can you recognize depositional textures/structures?
- Is the rock clastic, or is it a bound carbonate?
- If bound, describe it on its merits.
- If clastic, identify clasts, matrix, cement.
- Think about giving it a Dunham name.
- What about diagenesis?
- What about sedimentary structures?