

4.1 Introduction

Biological and biochemical processes are dominant in the formation of carbonate sediments, although inorganic precipitation of CaCO_3 from seawater also takes place. Once deposited, the chemical and physical processes of diagenesis can considerably modify the carbonate sediment. Limestones occur throughout the world in every geological period from the Cambrian onwards and reflect the changing fortunes, through evolution and extinction, of invertebrates with carbonate skeletons. In the Precambrian, carbonates are also abundant, but they are commonly dolomite and many contain stromatolites, produced largely by microbes, especially the cyanobacteria ('blue-green algae').

The economic importance of limestones today lies chiefly in their reservoir properties, as about half of the world's major petroleum reserves are contained within carbonate rocks. Limestones also are hosts to epigenetic lead and zinc sulphide deposits of the Mississippi Valley type and they have a wide variety of chemical and industrial uses, including the manufacture of cement.

As a result of recent geological events, notably the Pleistocene glaciation and a global sea-level lowstand, shallow-marine carbonate sediments are not widely developed at the present time. In the past, shallow epeiric seas periodically covered vast continental areas so that limestones were deposited over many thousands of square kilometres. On a broad scale, extensive carbonate deposition correlates with global sea-level highstands. Organisms with carbonate skeletons occur throughout the world's seas and oceans so that carbonate sediments can develop anywhere. However, there are several factors, of which the most important are temperature, salinity, water depth and siliciclastic input, that control carbonate deposition. Many carbonate skeletal organisms, such as the reef-building corals and many calcareous green algae, require warm waters in which to flourish. The majority

of carbonate sediments therefore occur in the tropical-subtropical belt, some 30° north and south of the Equator, and most limestones of the Phanerozoic formed in low latitudes. Biogenic carbonate productivity is highest in sea water of normal salinity in the shallow (less than 10 m), agitated part of the photic zone (the depth down to which light penetrates, of the order of 100–200 m). Skeletal carbonate sands do occur in higher latitudes, such as along the western coast of Ireland and Norway where calcareous red algae (especially *Lithothamnion*) and Mollusca dominate the sediments, and also off southern Australia, where bryozoans are especially important. However, there are few ancient examples of these so-called *cool-water carbonates* (see papers in James & Clarke (1997) and Lukasik *et al.* (2000) for a Miocene example from South Australia).

Non-skeletal grains, such as ooids and lime mud, are precipitated only in the warm shallow waters of the tropics. In the deeper-water pelagic environment, calcareous oozes are developed extensively and they are composed principally of the skeletons of pelagic organisms, Foraminifera and coccoliths, which live in the photic zone. High rates of carbonate dissolution at depths of several thousand metres result in little carbonate deposition below this, the *carbonate compensation depth*. Limestones also form in lakes and soils. One overriding control on carbonate deposition is the input of siliciclastic material. Many carbonate-producing organisms cannot tolerate the influx of large quantities of terrigenous mud.

For the petrographic study of limestones thin-sections are examined routinely, but acetate peels are useful too and quick to make. The surfaces of a limestone are polished and then etched with dilute acid (5% HCl or acetic acid for 30–60 s). The surface is then covered in acetone and a piece of acetate sheet is rolled on, with care taken not to trap air bubbles. After at least 10 min, the acetate is peeled off, and placed between glass. It is now ready for the microscope. All textural details are faithfully replicated, but of course

polarizers cannot be crossed. See Miller (1988a) for further details.

Much petrographic information is obtained from thin-sections viewed in transmitted light but further detail can be revealed through the technique of cathodoluminescence (CL): bombarding a polished thin-section (no cover slip) with electrons in a special small vacuum chamber mounted on a microscope stage (see Marshall (1988) and Miller (1988b) for details). An example of CL is given in Plate 13c. Examining a thin-section under UV light and observing the fluorescence can also reveal 'hidden' textures (see Dravis & Yurewicz, 1985). It is common practice to stain a thin-section of a limestone, or the polished surface of a hand specimen before taking a peel, to show the mineralogy (calcite or dolomite), and the iron content (ferroan or non-ferroan). Alizarin Red S and potassium ferricyanide are used (for the recipe see Miller, 1988a), and calcite (non-ferroan) stains pink and ferroan calcite blue to mauve; dolomite (non-ferroan) does not take up the stain whereas ferroan dolomite is turquoise-blue. Table 4.1 summarizes the features to look for in a carbonate thin-section and Table 4.2 gives a simple table for notes when describing a slide.

4.2 Mineralogy of carbonate sediments

In Recent sediments, two calcium carbonate minerals dominate: aragonite (orthorhombic) and calcite (trigonal). Two types of calcite are recognized depending on the magnesium content: low-magnesium calcite with less than 4 mol.% MgCO_3 and high-magnesium calcite with greater than 4 mol.%, but typically ranging between 11 and 19 mol.% MgCO_3 . By comparison, aragonite normally has a very low Mg content (less than 5000 p.p.m.) but it may contain up to 10 000 p.p.m. (1%) strontium, substituting for calcium. The mineralogy of a modern carbonate sediment depends largely on the skeletal and non-skeletal grains present. Carbonate skeletons of organisms have a specific mineralogy or mixture of mineralogies (Table 4.3), although the actual magnesium content of the calcites may vary, being partly dependent on ambient water temperature.

Aragonite is unstable at surface temperatures and pressures and in time high-Mg calcite loses its Mg. Thus all carbonate sediments with their original mixed mineralogy are converted to low-Mg calcite during

diagenesis. Grains and cement composed originally of low-Mg calcite generally are perfectly preserved in limestones; those originally of high-Mg calcite are mostly well preserved, like low-Mg calcite, but they may show some microstructural alteration and minor dissolution. Grains of aragonite are either replaced by calcite with some retention of original structure (the process of *calcitization*), or dissolved out completely to leave a mould, which later may be filled with calcite (a cement). A limestone also may be *dolomitized*, whereby dolomite, $\text{CaMg}(\text{CO}_3)_2$, replaces the CaCO_3 minerals and is precipitated as a cement (dolomitization is discussed in Section 4.8). Non-carbonate minerals in limestones include terrigenous quartz and clay, and pyrite, hematite, chert and phosphate of diagenetic origin. Evaporite minerals, in particular gypsum-anhydrite, may be closely associated with limestones (see Chapter 5).

4.3 Components of limestones

Limestones are very varied in composition but broadly the components can be divided into four groups: (i) non-skeletal grains, (ii) skeletal grains, (iii) micrite and (iv) cement. The common cement, sparite, and others are discussed in the section on diagenesis (Section 4.7).

4.3.1 Non-skeletal grains

Ooids and pisoids

Modern ooids are spherical-sub-spherical grains, consisting of one or more regular concentric lamellae around a nucleus, usually a carbonate particle or quartz grain (Figs 4.1 & 4.2). Sediment composed of ooids is referred to as an *oolite*. The term ooid (formerly oolith) has been restricted to grains less than 2 mm in diameter and the term pisoid (formerly pisolith) is used for similar grains of a larger diameter. If only one lamella is developed around a nucleus, then the term superficial ooid is applied (Fig. 4.1). Composite ooids consist of several small ooids enveloped by concentric lamellae. *Coated grain* is a general term frequently used for ooids and pisoids, and includes *oncooids*, grains with a microbial coating (see Section 4.3.3).

The majority of modern ooids range from 0.2 to 0.5 mm in diameter. They typically form in agitated

Table 4.1 Scheme for petrographic description for carbonate rocks**Hand specimen**

Note the colour; type(s) of grain(s); grain size; grain shape if significant; evidence of fossil breakage, disarticulation, abrasion, packing, etc.; presence of micrite and/or sparry calcite if observable; physical and biogenic sedimentary structures; stylolites, etc. Any evidence of diagenetic alteration, e.g. dolomitization, silicification, dissolution, compaction

Thin-section

Check macroscopic features of thin-section by holding up to light and noting any lamination or large fossils or grains

Grains: type(s) — bioclasts, ooids, peloids, intraclasts/aggregates, etc.; size, sorting, shape, packing of grains; mineralogy/composition of grains (in stained sections — ferroan/non-ferroan calcite/dolomite). Identify bioclasts from shape, internal structure and preservation.

Ooids — determine original composition (aragonite or calcite). Peloids — micritized grains or faecal pellets?

Micrite (lime mud): grains mostly less than 4 μm and normally almost opaque and brownish in colour, but may be peloidal. Any aggrading neomorphism?

Cements: Usually coarser than 10 μm ; identify types of cement: fibrous calcite, bladed calcite, syntaxial overgrowths, drusy calcite spar, poikilotopic calcite spar; note geometry: meniscus, isopachous, pore-filling; spar may be non-ferroan, ferroan, or zoned (use staining). Determine original mineralogy (aragonite, calcite, high-Mg calcite). Timing of cement precipitation, early versus late, pre- versus post-compaction. Where cemented — marine, meteoric, burial?

Replacement, recrystallization and neomorphism: minerals such as dolomite, silica (chert) or phosphate may replace calcite grains, micrite or cement. Neomorphism of grains, e.g. aragonitic shells, ooids or cements replaced by calcite (calcitization) and micrite replaced by microspar — aggrading neomorphism. Timing of replacement relative to compaction?

Dolomitization: scattered rhombs or pervasively dolomitized? Any fabric control: grains or micrite preferentially dolomitized? Determine dolomite texture and crystal shape/size/zonation. Is dolomite early or late, pre- or post-compaction; timing relative to calcite diagenesis. Any dedolomitization?

Compaction: look for evidence of mechanical compaction (broken bioclasts, broken micrite envelopes, spalled oolitic coatings or early cements). Check for chemical compaction: sutured contacts between grains, pre-burial spar cementation, and stylolites: through-going pressure dissolution seams, post-spar cementation

Porosity: identify type: primary (intergranular, intragranular, cavity, growth, etc.) and secondary (fracture, dissolutional, intercrystalline — as through dolomitization)

Classify using the Dunham scheme (or Folk) after assessing the proportion of grains, cement and micrite, and dominant grain types. Where diagenetic alteration is significant, then qualify name, e.g. partly dolomitized bioclastic grainstone, silicified oosparite, recrystallized lime mudstone

Interpretation

Depositional environment: use grain types and texture, e.g. most grainstones represent moderate to high energy shallow subtidal, many lime mudstones—wackestones represent lower-energy, lagoonal or outer-shelf/ramp environments. Check bioclasts; these can indicate open-marine, restricted, deep-water, shallow, non-marine, etc.

Diagenesis: identify early (near-surface) and late (burial) processes, determine timing of cementation. Assess degree of compaction; any dolomitization? Try to interpret in terms of pore-water chemistry (e.g. marine or freshwater from cement fabric, and Eh/pH from ferroan/non-ferroan nature of calcite/dolomite). Deduce succession of diagenetic events and porosity evolution

waters where they are frequently moved as sandwaves, dunes and ripples by tidal and storm currents, and wave action. On the Bahama platform, the ooids form shoals close to the edge of the platform. In the Arabian Gulf the ooids form in tidal deltas at the mouth of tidal inlets between barrier islands along the Trucial Coast. Along the Yucatan shoreline (northeast Mexico), ooids are being precipitated in the shoreface and fore-shore zones. Depths of water where ooids precipitate usually are less than 5 m, but they may reach 10–15 m.

Practically all marine ooids forming today, such as in the Bahamas and Arabian Gulf, are composed of aragonite and they have a high surface polish (Fig.

4.2). Bimineralic high-Mg calcite–aragonite ooids have been recorded from Baffin Bay, Texas. Relict early Holocene ooids of high-Mg calcite occur off the Great Barrier Reef of Queensland and on the Amazon Shelf.

The characteristic microstructure of modern aragonitic marine ooids is a tangential orientation of acicular crystals or needles, 2 μm in length. Lamellae of microcrystalline aragonite and of randomly oriented aragonite needles also occur. The sub-Recent high-Mg calcite ooids have a radial fabric. Ooids contain organic matter, located chiefly between lamellae and in the microcrystalline layers.

Table 4.2 Scheme for describing carbonate rocks in thin-section

Features	Thin-section 1	Thin-section 2
<i>Grains present and percentage</i>		
Bioclasts		
Ooids		
Pellids		
Intraclasts		
Micrite: uniform, peloidal, microsparitic		
<i>Texture</i>		
Roundness, sorting, fabric, binding, framework		
<i>Cavity structures</i>		
Umbrella, geopetal, fenestrae—birdseye, laminar, burrow, rootlet		
<i>Cements</i>		
Fibrous, bladed, drusy calcite, sparite, geometry		
<i>Replacements</i>		
Calcitization, neomorphism, microspar, micritization, dissolution		
<i>Dolomite</i>		
Crystal shape, size, distribution, texture, fabric preservation, timing		
<i>Evidence of compaction</i>		
Broken grains and micrite envelopes, sutured contacts, stylolites		
<i>Porosity</i>		
Intergranular, intragranular, framework, mouldic, intercrystalline, fracture, etc.		
<i>Name/microfacies</i>		
<i>Depositional environment</i>		
<i>Diagenetic events</i>	1:	
	2:	
	3:	

Ooids and pisoids can also form in quieter-water marine locations, such as in lagoons (e.g. Baffin Bay, Texas) and on tidal flats (e.g. the Trucial Coast). Ooids–pisoids also are formed in association with tepee structures on intertidal–supratidal flats (see Section 4.6.1), being precipitated in local pools and beneath cemented crusts. These low-energy coated grains commonly have a strong radial fabric, so that they do break relatively easily (e.g. Fig. 4.3). Those forming in a vadose situation, such as in association with tepees, are commonly asymmetric; they may also show downward thickening laminae and a fitted fabric from *in situ* growth. Such *vadose pisoids* are a feature

of the back-reef facies in the Capitan Reef Complex of Texas–New Mexico.

Ooids can form in high-energy locations in lakes, as in the Great Salt Lake, Utah and Pyramid Lake, Nevada. These lacustrine ooids are commonly dull, and they may have a cerebriform (bumpy) surface. The Great Salt Lake ooids are composed of aragonite and many have a strong radial fabric. The Pyramid Lake ooids are biminerally (low-Mg calcite–aragonite).

Structures resembling ooids–pisoids do form in calcareous soils. They are usually composed of fine-grained calcite and have a poorly developed concentric lamination, which may be asymmetric. They form

Organism	Mineralogy			
	Aragonite	Low-Mg calcite	High-Mg calcite	Aragonite + calcite
Mollusca				
Bivalves	X	X		X
Gastropods	X			X
Pteropods	X			
Cephalopods	X		(X)	
Brachiopods		X	(X)	
Corals				
Scleractinian	X			
Rugose + Tabulate		X	X	
Sponges	X	X	X	
Bryozoans	X		X	X
Echinoderms			X	
Ostracods		X	X	
Foraminifera				
Benthic	(X)		X	
Pelagic		X		
Algae				
Coccolithophoridae		X		
Rhodophyta	X		X	
Chlorophyta	X			
Charophyta		X		

Table 4.3 The mineralogy of carbonate skeletons (X, dominant mineralogy; (X), less common). During diagenesis, these mineralogies may be altered or replaced; in particular, aragonite is metastable and is invariably replaced by calcite, and high-Mg calcite loses its Mg

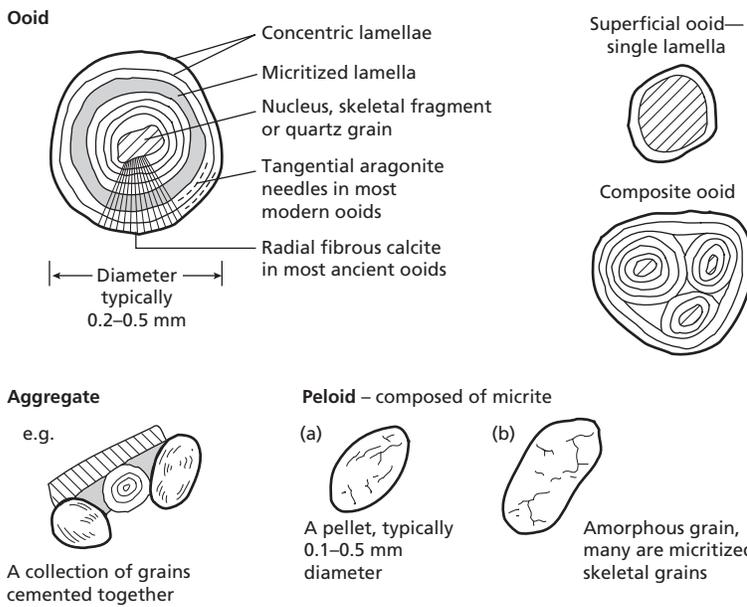


Fig. 4.1 The principal non-skeletal grains in limestones: ooids, peloids and aggregates.

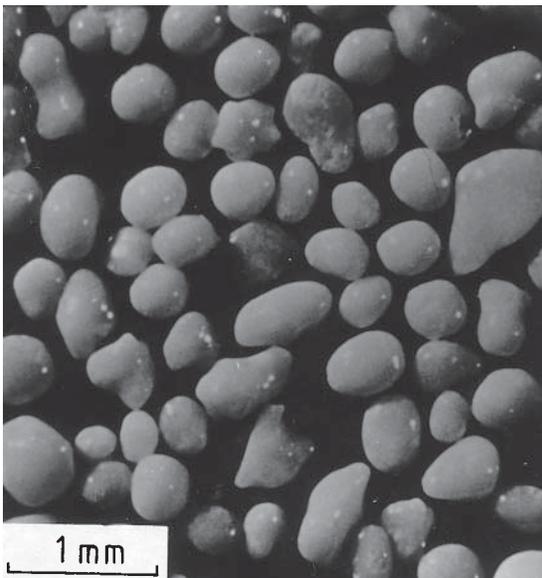


Fig. 4.2 Modern aragonitic ooids from the Bahamas with polished surface.

largely by microbial processes and calcification of fungal and bacterial filaments. These *calcrete pisoids* are commonly associated with laminated crusts, formed by the calcification of root mats (see Section 4.6.1).

Ancient marine ooids

Ooids in the rock record are composed of calcite (low Mg), unless dolomitized or silicified. However, although there has been much discussion over the matter, it is clear that some were originally calcite, whereas others were originally aragonite. Former bimineralic ooids also have been reported. Primary calcite ooids, whether in high-energy or low-energy facies, typically have a radial texture of wedge-shaped, fibrous crystals (see Plate 6c). Under crossed polars, an extinction cross is seen. The cortex of larger, originally calcitic ooids may have an inner radial part and an outer radial-concentric part. It is not easy to determine whether calcitic ooids originally had a low or high Mg content. The presence of small dolomite crystals (microdolomites), evidence of minor dissolution of the cortex or a moderate to high iron content may indicate an original high-Mg calcite composition (see Section 4.7.1).

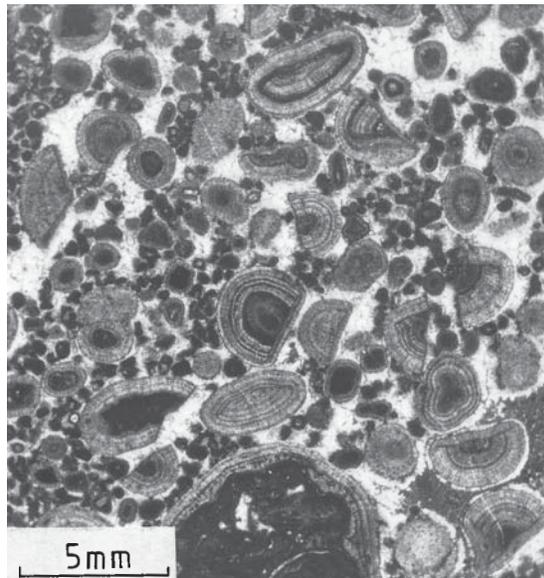


Fig. 4.3 Pisoids, probably of vadose origin (vadoids) showing strong radial structure and broken and re-coated grains. Smaller grains are peloids. Rock is a grainstone, but it is 100% dolomite. The sediment was dolomitized with perfect preservation of original texture. The clear spar is a dolomitic cement. Dürrenstein Formation, Triassic. The Dolomites, Italy.

Ancient ooids originally of aragonite will have been altered during diagenesis to a greater or lesser extent (see Plate 6d). They may be replaced by calcite with some retention of the original tangential structure (if high energy) or radial structure (if quiet water) through the presence of minute inclusions of organic matter and/or relics of the aragonite. This replacement process is termed *calcitization* (also see Section 4.7.4). Alternatively, the aragonite of the ooids may be dissolved out completely, to leave oomoulds. These holes may be left empty to give the limestone an oomouldic porosity, or filled with calcite cement (see Plate 6d). Some ancient ooids have a fine-grained micritic texture. As with modern ooids, this may be the result of micritization by endolithic microbial organisms (see Section 4.3.3) or it may result from diagenetic alteration (*neomorphism*, see Section 4.7.3).

Origin of ooids

There has been much discussion on the origin of ooids; current ideas invoke biochemical or inorganic pro-

cesses, a direct microbial origin being largely discarded now. Although a precise mechanism of inorganic precipitation has not been demonstrated, seawater in shallow tropical areas is supersaturated with respect to CaCO_3 , so that this, together with water agitation, CO_2 degassing and elevated temperature, might be sufficient to bring about carbonate precipitation on nuclei. A biochemical origin hinges on the organic mucilage that coats and permeates the ooids. One view is that bacterial activity within the organic matter creates a microenvironment conducive to carbonate precipitation. Some ooids have a proteinaceous matrix, also suggesting a biochemical process, because in organisms it is amino acids that induce calcification. A biological origin is supported by SEM examination of modern marine ooids, which shows that the aragonite rods and nanograins that form the cortex are identical to those associated with endolithic and epilithic bacteria and mucilaginous films occurring on and within the ooids. Laboratory synthesis of ooids has suggested that organic compounds in the water are instrumental in the formation of quiet-water ooids with their radial fabric, but that ooids formed in agitated conditions are precipitated inorganically.

The factors determining the primary mineralogy of ooids are water chemistry, especially P_{CO_2} , Mg/Ca ratio and carbonate saturation, and possibly the degree of water agitation. It is believed that aragonite and high-Mg calcite ooids are precipitated when P_{CO_2} is low and Mg/Ca ratio high, and that low-Mg calcite ooids form when P_{CO_2} is high and Mg/Ca ratio low. It is unclear what controls the precipitation of aragonite ooids as opposed to high-Mg calcite ooids but carbonate supply rate has been implicated. High carbonate supply, as would occur in high-energy locations, is thought to favour aragonite precipitation.

Surveys of the original mineralogy of ooids through the Phanerozoic have shown that there is a secular variation, with aragonitic ooids, which may be associated with calcitic ooids (presumably high-Mg calcite originally) in the late Precambrian/early Cambrian, mid-Carboniferous through Triassic and Tertiary to Recent, and calcitic ooids (presumed to have had a low-to-moderate Mg content) dominant in the mid-Palaeozoic and Jurassic–Cretaceous (Fig. 4.4). This pattern suggests that there have been subtle fluctuations in seawater chemistry through time, in P_{CO_2} and/or the Mg/Ca ratio. See Sandberg (1983) and the review in Stanley & Hardie (1998). Some anomalies

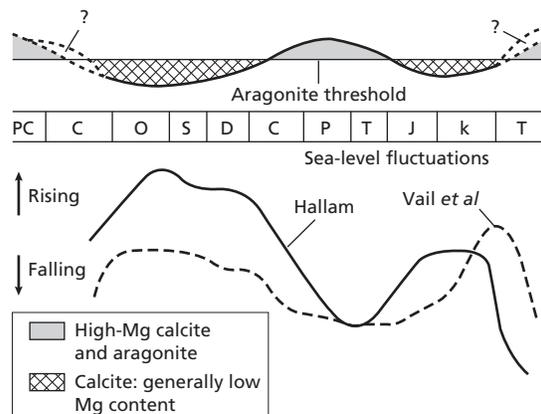


Fig. 4.4 Mineralogy of marine, abiogenic, carbonate precipitates through the Phanerozoic, compared with the first-order global sea-level curve.

do occur in the broad trend, notably in Upper Jurassic strata where aragonitic ooids are recorded, as in the Smackover Formation of the US Gulf Coast subsurface (see Plate 6d; Heydari & Moore, 1994).

The general trend shown in Fig. 4.4 appears to tie in with the first-order, global sea-level curve, suggesting that a geotectonic mechanism(s) is causing the subtle variations in seawater chemistry, giving rise to the secular variation in ooid mineralogy. High sea-level stands, correlating with calcite seas, are times of high rates of sea-floor spreading, when P_{CO_2} can be expected to be relatively high from increased metamorphism at subduction zones and the Mg/Ca ratio relatively low from increased extraction of Mg^{2+} at mid-ocean ridges as sea water is pumped through (see Stanley & Hardie, 1998). Temporal exceptions to the broad trend, as in the Late Jurassic, could be the result of local conditions, the most likely being a raised Mg/Ca ratio from evaporite deposition, leading to aragonite precipitation.

Peloids

Peloids are spherical, ellipsoidal or angular grains, composed of microcrystalline carbonate, but with no internal structure (Fig. 4.1 and Plates 6a, 7a,c & 8c). The size of peloids may reach several millimetres but the majority are in the range of 0.1–0.5 mm in diameter. Most peloids are of faecal origin and so can be referred to as pellets. Organisms such as gastropods,

crustaceans and polychaetes produce pellets in vast quantities. Faecal pellets have a regular shape and they are rich in organic matter. They are most common in the sediments of protected environments such as lagoons and tidal flats. Pellets are very common in limestones and many micritic limestones, seemingly without sand-sized grains, may actually be pelleted. The definition of pellets is commonly lost as a result of diagenetic processes, and the limestones may show a flocculent or clotted texture.

The term peloid includes micritized bioclastic grains formed by alteration of skeletal fragments by microboring microbes and recrystallization (Section 4.3.3). They are more irregular in shape and are an important component of modern Bahamian carbonate sediments.

Aggregates and intraclasts

Aggregates consist of several carbonate particles cemented together by a microcrystalline cement or bound by organic matter. Such grains in the Bahamas are known as grapestones and form in relatively protected shallow subtidal areas, usually beneath a thin, surficial microbial mat.

Intraclasts are fragments of lithified or partly lithified sediment. A common type of intraclast in carbonate sediments is a micritic flake or chip, derived from desiccation of tidal-flat muds or disruption by storms of partially lithified or cemented subtidal lime muds (Fig. 4.5). The latter are particularly common in the Precambrian and Cambrian. An abundance of these

flakes produces flat-pebble or edgewise conglomerates, also called flakestones. They may show an imbrication of clasts (see Section 2.2.4).

One distinctive intraclast is a *black pebble*. These are carbon-impregnated pebbles, commonly with soil fabrics, which are associated with palaeosoils, laminated crusts and palaeokarsts, and may be reworked into intraformational conglomerates. They form in soil horizons and may be the result of forest fires, or more likely organic-matter impregnation in water-logged, reducing conditions (Shinn & Lidz, 1988).

4.3.2 Skeletal components (excluding algae)

The skeletal components of a limestone are a reflection of the distribution of carbonate-secreting invertebrates through time and space (Fig. 4.6). Environmental factors, such as depth, temperature, salinity, substrate and turbulence, control the distribution and development of the organisms in the various carbonate environments. Throughout the Phanerozoic, various groups expanded and evolved to occupy the niches left by others that were declining or becoming extinct. The mineralogy of carbonate skeletons also varies through the Phanerozoic, like the inorganic precipitates (Section 4.3.1), and this probably is also a reflection of tectonically forced shifts in seawater chemistry (see Stanley & Hardie, 1998).

The main skeletal contributors to limestones are discussed in the following sections, with comments on their recognition. Detailed accounts of skeletal struc-

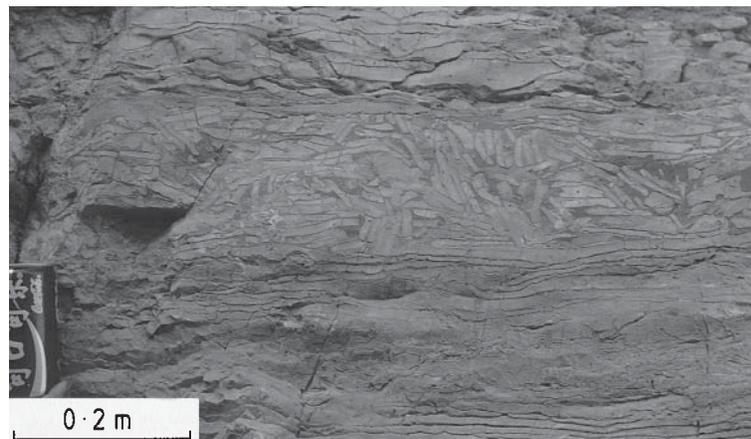


Fig. 4.5 Intraclasts. Storm bed composed of rip-up clasts of micritic limestone. Cambrian. Qinhuandao, northeast China.

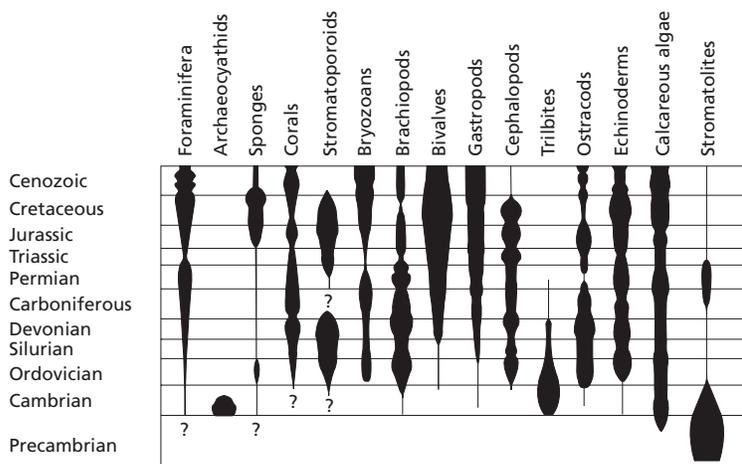


Fig. 4.6 Age range and generalized taxonomic diversity of principal carbonate-secreting organisms.

ture and thin-section appearance are given in Bathurst (1975), Scholle (1978), Flügel (1982), Adams *et al.* (1984) and Adams & Mackenzie (1998). For the identification of skeletal particles in thin-section the points to note are:

- 1 shape (and size), bearing in mind that under the microscope a two-dimensional view only is given; look for other sections of the same fossil to determine the three-dimensional shape;
- 2 internal microstructure, which may be modified or obliterated by diagenesis;
- 3 mineralogy—although in a limestone everything will be calcite, unless dolomitized or silicified, fabric evidence can be used to decide if a skeletal particle was originally aragonitic (staining a thin-section for ferroan calcite and dolomite using Alizarin red S and potassium ferricyanide (see Section 4.1) may provide additional information—for example, skeletal components originally of high-Mg calcite may be replaced preferentially by ferroan calcite);
- 4 other features likely to be diagnostic, such as presence of spines or pores.

Mollusca

Bivalves, gastropods and cephalopods occur in limestones from the Early Palaeozoic onwards. The bivalves are a very large group with species occupying most marine, brackish and freshwater environments. Bivalves have been important contributors to marine carbonate sediments, particularly since the Tertiary following the decline of the brachiopods. The modes

of life are very varied, too, including infaunal (living within the sediment), epifaunal (attached to a hard substrate), vagile (crawlers), nektonic (free-swimming) and planktonic (free-floating). Certain bivalves, such as oysters, may form reef-like structures. During the Cretaceous, masses of aberrant, coral-like bivalves called rudists formed reefs in Mexico, southern USA, the Mediterranean region and the Middle East, for example. Fresh- and brackish-water limestones may be composed largely of bivalves; examples occur in the Upper Carboniferous, Upper Triassic (Rhaetic) and Upper Jurassic (Purbeck) of western Europe.

The majority of bivalve shells are composed of aragonite; some are of mixed mineralogy (the rudists, for example, Plate 7e); others, such as the oysters and scallops, are calcitic. Bivalve shells consist of several layers of specific internal microstructure, composed of micron-sized crystallites (see Plate 7b,c). One common shell structure is of an inner nacreous layer consisting of sheets of aragonite tablets, and an outer prismatic layer of aragonite (or calcite). If composed originally of aragonite, the internal structure of a fossil bivalve shell is likely to be poorly preserved or not preserved at all (see Fig. 4.7). The aragonite may be dissolved out completely to leave a mould, which subsequently may be filled by calcite (a cement). This is the most common mode of preservation so that most bivalve fragments in limestones are composed of clear, coarse drusy sparite (see Plate 7d; also Plates 6c & 9e). Alternatively, the aragonite of the shell may be replaced by calcite (calcitized) so that faint relics of the

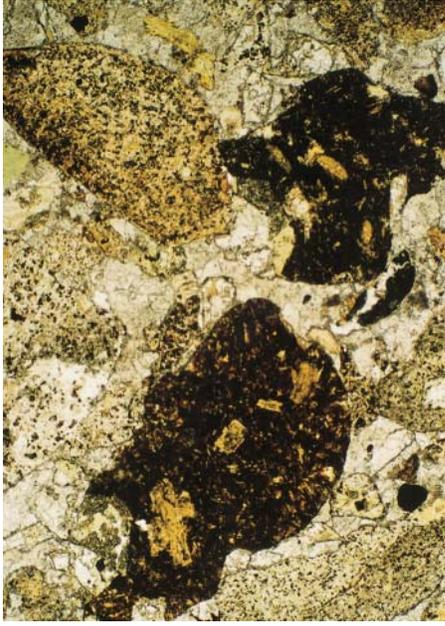
Plate 1

(a) Lithic grains of sedimentary origin—laminated and silty shale, also angular quartz grains. Plane-polarized light. Carboniferous fluvial sandstone. Cantabrians, Spain. Field of view 3 × 2 mm.

(b) Lithic grains of volcanic origin. The two dark grains consist of mafic phenocrysts in a very altered, originally glassy, dark groundmass. The other grains consist of minute feldspar laths in a glassy groundmass, and have numerous small dark iron-rich spots. The cement is calcite. Plane-polarized light. Triassic shallow-marine sandstone. The Dolomites, Italy. Field of view 3 × 2 mm.

(c) Quartz grains: several monocrystalline quartz grains showing both uniform and unit extinction; polycrystalline quartz grains with several and many subcrystals, some of the latter with sutured contacts. The cement is poikilotopic calcite. Crossed polars. Permian aeolian sandstone. Durham, England. Field of view 3 × 2 mm.

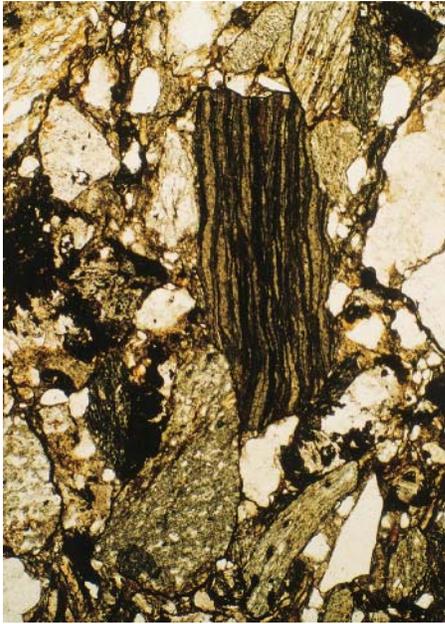
(d) Feldspar grains: microcline on left with grid-iron twinning and orthoclase on right with bright specks of alteration material (sericite). Quartz grains also present (mostly monocrystalline with unit extinction), the one in lower centre with overgrowth. Muscovite flake extreme right showing blue colour. Grains are coated with hematite, giving red/brown rim. Crossed polars. Precambrian fluvial sandstone. Torridon, Scotland. Field of view 1.2 × 0.8 mm.



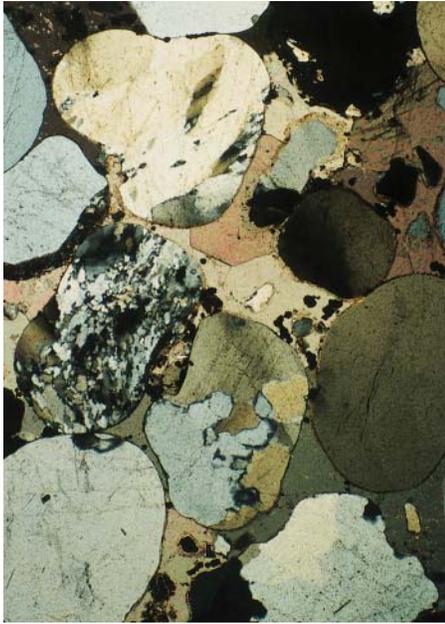
(b)



(d)



(a)



(c)

Plate 2

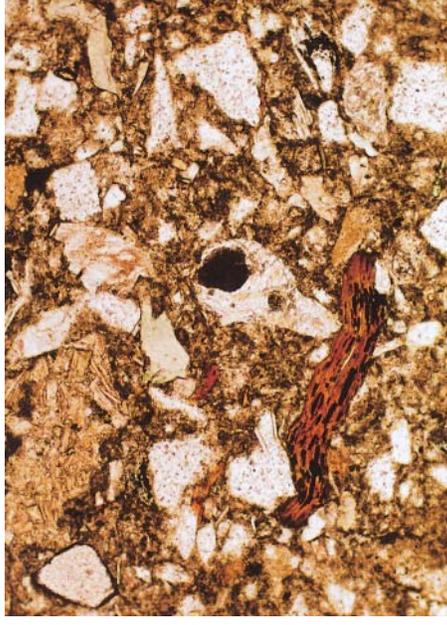
(a) Muscovite and kaolinite in quartz arenite. Muscovite shows orange colour. Kaolinite forms the small dark crystals between the quartz grains, which are mostly monocrystalline with unit extinction; the one on the left has minute fluid and mineral inclusions. Crossed polars. Carboniferous fluvial sandstone, Northumberland, England. Field of view 1.2×0.8 mm.

(b) Biotite mica (brown) showing effects of compaction, and light brown matrix consisting of minute, unresolvable clay minerals, iron minerals (dark brown/black), and silt-grade quartz. Also present are many angular quartz grains (white), and feldspar and lithic grains. Plane-polarized light. Silurian turbidite greywacke. Scotland. Field of view 1.2×0.8 mm.

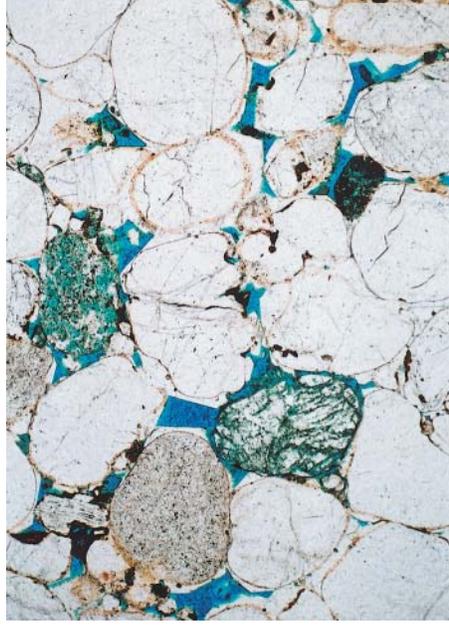
(c, d) Quartz arenite with well-developed overgrowths on quartz grains, which are mostly monocrystalline with unit extinction. Dusty looking quartz grain on left is of hydrothermal origin and is full of fluid inclusions. Note clear overgrowth. Red hematite coating around grains. Two feldspar grains show effects of dissolution. Rock impregnated with blue resin to show porosity (reduced intergranular and dissolutional intragranular). **(c)** Plane-polarized light; **(d)** crossed polars. Permian aeolian sandstone. Cumbria, England. Field of view 3×2 mm.



(a)



(b)



(c)



(d)

Plate 3

(a) Quartz arenite with angular looking quartz grains from overgrowth cement. Dust-line around grains visible in some cases. Supermature sandstone consisting of monocrystalline, unit-extinguishing quartz grains. Zircon, a heavy mineral, upper left (red). Crossed polars.

Carboniferous marine sandstone. Durham, England. Field of view 1.2×0.8 mm.

(b) Litharenite with sedimentary rock fragments of fine sandstone and mudrock (some showing lamination). Quartz grains also present (clear). Note the effects of compaction: a tight-fitting arrangement of grains, and some interpenetration and squashing of grains too. Plane-polarized light. Carboniferous fluvial sandstone. Cantabrians, Spain. Field of view

6×4 mm.

(c) Arkose with many feldspar grains (dusty/dirty looking compared with clearer quartz grains), and hematite coatings around sand grains; note that hematite is absent where grains are in contact. Plane-polarized light. Precambrian fluvial sandstone. Torridon, Scotland. Field of view 1.2×0.8 mm.

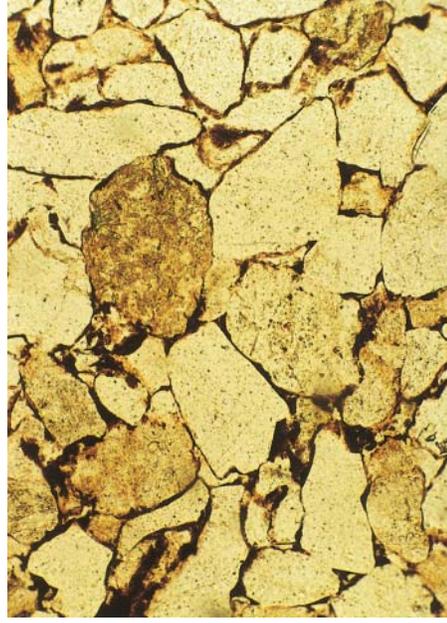
(d) Arkose with feldspar grains (orthoclase, microcline) showing incipient replacement by sericite (minute bright crystals). Crossed polars. Precambrian fluvial sandstone. Torridon, Scotland. Field of view 1.2×0.8 mm.



(a)



(b)



(c)

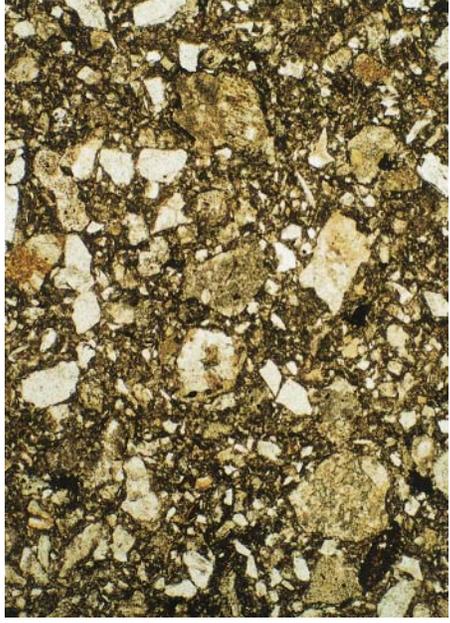


(d)

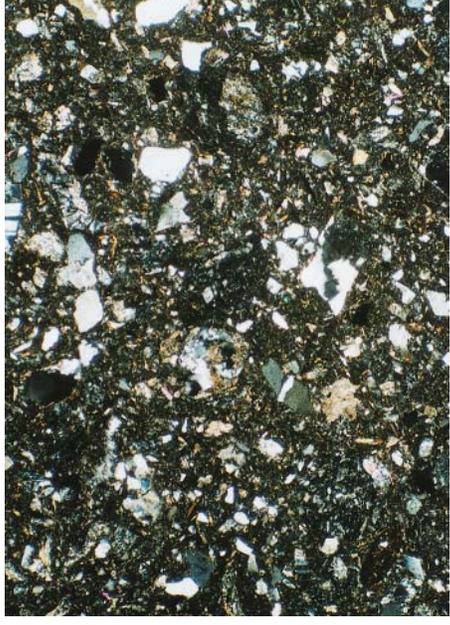
Plate 4

(a, b) Greywacke. Quartz, feldspar and lithic grains are contained in a fine-grained matrix of chlorite and silt-grade quartz: **(a)** plane-polarized light; **(b)** crossed polars. Silurian turbidite greywacke. Southern Uplands, Scotland. Field of view 3 x 2 mm.

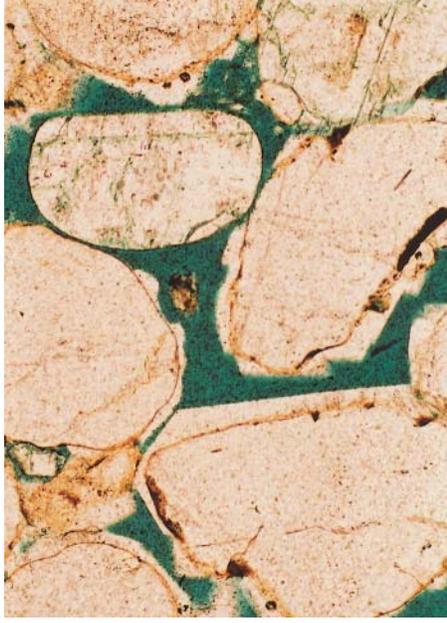
(c, d) Quartz overgrowths on quartz grains. The grain surface is shown by the red hematite coating. Better overgrowths on monocrystalline quartz than polycrystalline quartz grains. Feldspar grain upper right has no overgrowth and shows some effects of dissolution. Rock impregnated with blue resin to show porosity (reduced intergranular): **(c)** plane-polarized light; **(d)** crossed polars. Permian aeolian sandstone. Cumbria, England. Field of view 1.2 x 0.8 mm.



(a)



(b)



(c)



(d)

Plate 5

(a, b) Calcite cement in quartz arenite. Large poikilotopic calcite crystals enclosing several grains. Quartz grains are well-rounded monocrystalline, with unit and undulose extinction, and polycrystalline. Feldspar grain showing splitting by calcite crystal; detail in (b). Crossed polars. Permian aeolian sandstone. Durham, England. **(a)** field of view 3×2 mm; **(b)** field of view 1.2×0.8 mm.

(c) Kaolinite (small black and white crystallites) between quartz grains (monocrystalline and polycrystalline), probably replacing a feldspar grain. Distorted muscovite flake (showing blue colour) between quartz grains. Crossed polars. Carboniferous fluvial sandstone. Northumberland, England. Field of view 1.2×0.8 mm.

(d) Mudrock showing effects of compaction with fracture of shells (thin brachiopods), folding of laminae around shells and flattening of burrows (upper left). Minute silt-grade quartz and shell debris disseminated throughout mud; round white grain (centre left) is crinoid ossicle. Carboniferous marine mudrock. Northumberland, England. Field of view 6×4 mm.



(a)



(b)



(c)



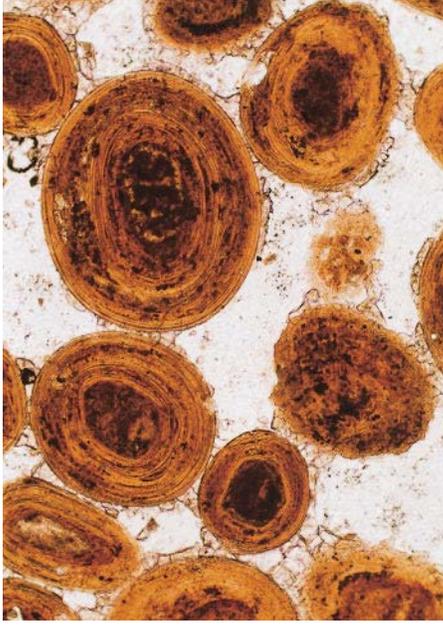
(d)

Plate 6

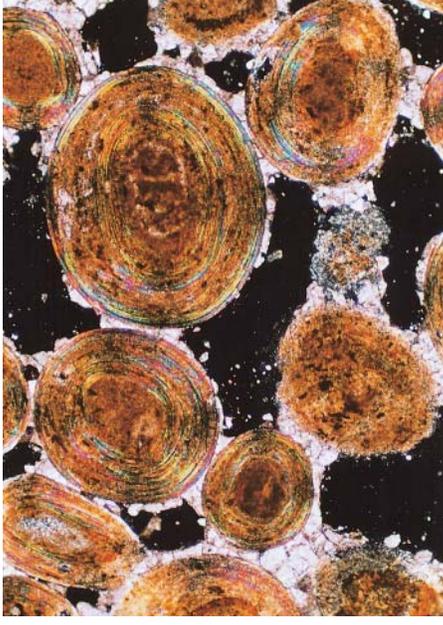
(a, b) Holocene ooids composed of aragonite showing concentric structure and a nucleus, several of which are peloids, and one in the centre of the biggest ooid is a bioclast. Structureless oval grain at lower left is a peloid. Also present is a meniscus cement of calcite precipitated in the meteoric vadose zone. The white areas between grains in (a) and the black areas in (b) are pore space: **(a)** plane-polarized light; **(b)** crossed polars. Joulters Cay, Bahamas. Field of view 1.2 × 0.8 mm.

(c) Primary, calcitic ooids with strong radial-concentric structure and nuclei mostly of peloids. Bivalve fragment, now composed of clear calcite spar crystals, is also coated. Notice contact between grains; a little interpenetration indicating some burial compaction before cementation. The ooids are contained in a very large poikilotopic calcite cement (here appearing white). Oolitic grainstone, Jurassic. Lincolnshire, England. Field of view 3 × 2 mm.

(d) Formerly aragonitic ooids, now composed of calcite with poor preservation of original concentric structure and oomoulds (filled with blue resin). Some compaction of oomoulds indicating that the drusy calcite spar cement is a burial precipitate. Oolitic grainstone, Smackover Formation, Jurassic. Subsurface Arkansas, USA. Field of view 3 × 2 mm.



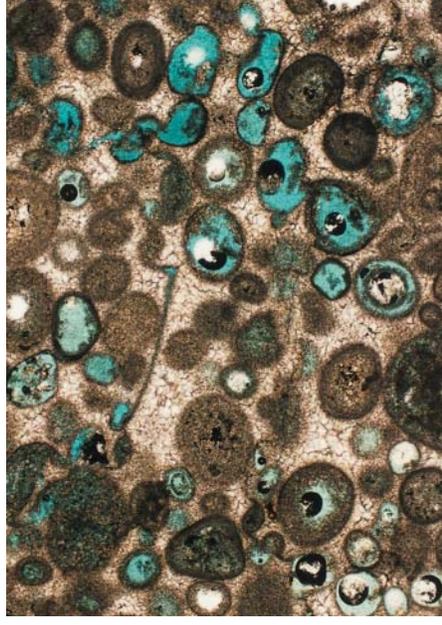
(a)



(b)



(c)



(d)

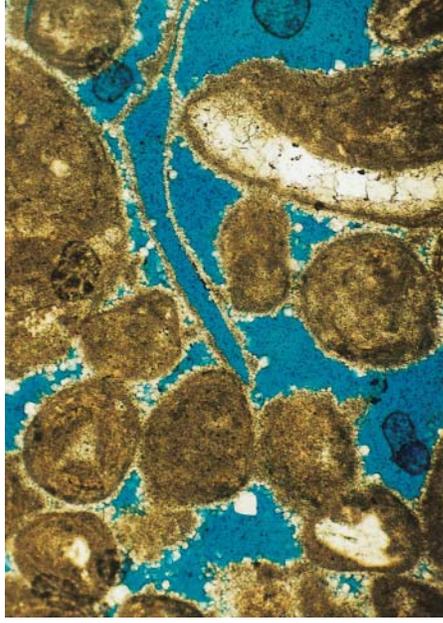
Plate 7

(a) Peloids; many are micritized bioclasts and ooids, some are faecal pellets. Micrite envelope defines a bivalve shell that has dissolved away; bivalve shell within coated grain is replaced by coarse calcite crystals. The sparse cement consists of small, stubby calcite crystals of probable meteoric phreatic origin. Small dolomite rhombs also present. Blue resin shows porosity. Jurassic. Dorset, England. Field of view 0.8×0.8 mm.

(b,c) Modern bivalve fragment with micrite envelope. Shell, composed of aragonite, consists of minute crystallites giving a sweeping extinction under crossed polars **(c)**. Abu Dhabi, UAE. Field of view 0.8×1.0 mm.

(d) Bivalve fragment (the elongate grain) in centre with a prominent micrite envelope and shell now composed of drusy calcite spar, a cement. The micrite envelope has fractured (in the centre) as a result of compaction. Also present are numerous peloids, most of which are micritized bioclasts, and some crinoid fragments, in a sparite cement. Urgonian, Cretaceous. Vercors, France. Field of view 3×2 mm.

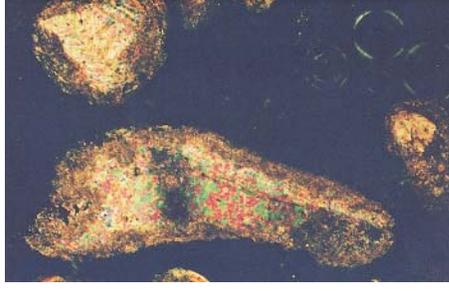
(e) Hippuritid rudist bivalve (the conical, attached valve) showing brown, well-preserved fibrous calcite outer wall, and thinner inner wall with tabulae, which originally were aragonite, but are now composed of calcite spar. Round areas of sediment (micrite) are the fills of sponge borings. Cretaceous. Provence, France. Field of view 6×4 mm.



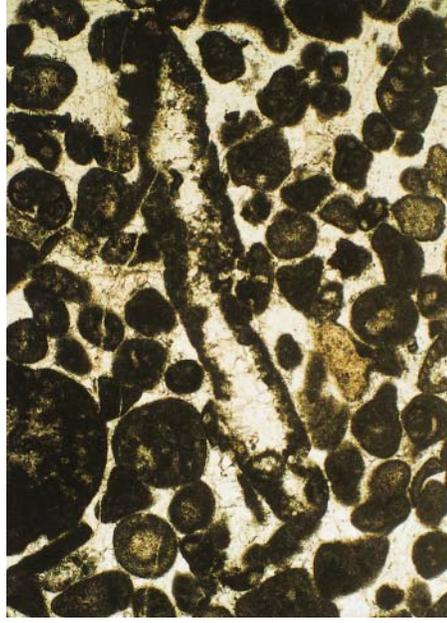
(a)



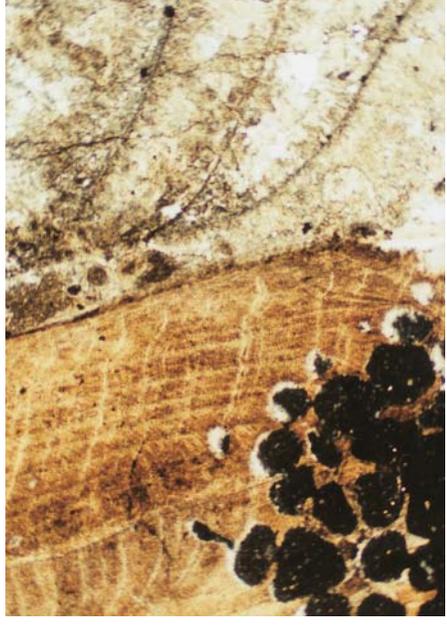
(b)



(c)



(d)



(e)

Plate 8

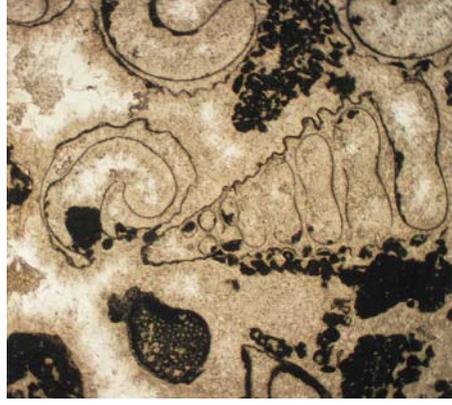
(a, b) Calcitized bivalve shells. Shells originally composed of aragonite but replaced by calcite with some retention of original shell structure. Calcite crystals cross-cut the shell structure and are pseudopleochroic (different shades of brown on slide rotation): **(a)** plane-polarized light and **(b)** crossed polars. Jurassic. Dorset, England. Field of view 0.8 × 0.8 mm.



(a)

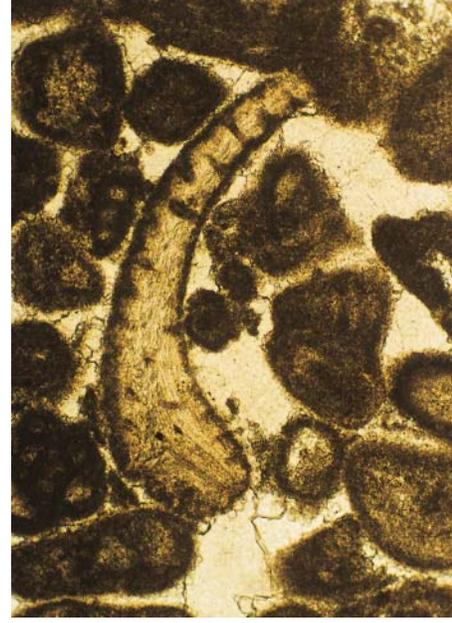


(b)

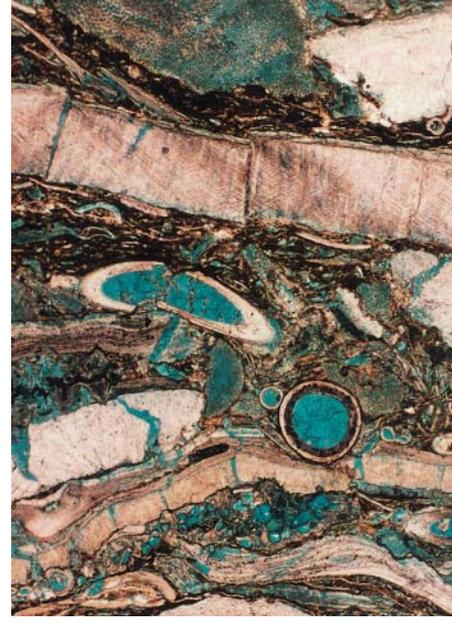


(c)

(c) Gastropods, in long- and cross-section, defined by thin micrite envelopes (black). The shells, originally aragonite, dissolved out and voids were filled by marine fibrous calcite cement (pale brown). A later clear calcite cement filled remaining pores (white). Jurassic. Sicily, Italy. Field of view 4 × 4 mm.



(d)



(e)

(d) Brachiopod shell with puncti, some filled with lime mud sediment, showing preservation of internal structure consisting of obliquely arranged fibres. Other grains are micritized bioclasts (peloids). Bioclastic grainstone. Cretaceous. Vercors, France. Field of view 3 × 2 mm.

(e) Brachiopod shells and spines with well-preserved shell structure, but limestone has suffered compaction and many shells are broken. Thin-section is stained with Alizarin Red S and potassium ferricyanide; bioclasts are pink (calcite) and cement is blue (ferroan calcite). Cement occurs within cracks showing that it is a burial precipitate. Crinoids and clay (brown) also present. Bioclastic packstone.

Carboniferous. Northumberland, England. Field of view 6 × 4 mm.

Plate 9

(a) Rugose coral (*Lithostrotion* sp.) showing internal plates (septa, tabulae and dissepiments). The pores within the coral are partly filled with an initial isopachous fibrous marine cement and then by internal sediment. Carboniferous. Durham, England. Field of view 6 × 4 mm.

(b) Scleractinian coral from a Jurassic patch reef showing variable preservation of structure as a result of replacement of the original aragonite by calcite. Also present, on the right, is a boring made by a lithophagid bivalve (shells present); the boring shows a geopetal structure, with sediment below (dark) and drusy calcite spar above (white). Yorkshire, England. Field of view 6 × 4 mm.

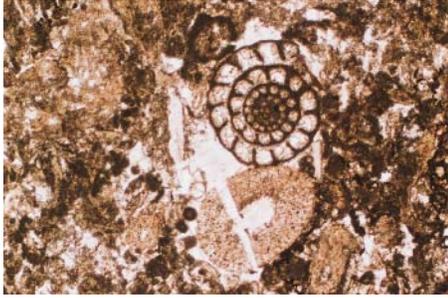
(c, d, e) Foraminifera. **(c)** Endothyraeid foram, crinoid to left, cut by a calcite vein. Carboniferous. Clwyd, Wales. Field of view 4 × 2 mm. **(d)** *Nummulites*. Bioclastic grainstone, Eocene. Tunisia. Field of view 6 × 4 mm. **(e)** Miliolids, also bivalve fragments. Cretaceous. Vercors, France. Field of view 4 × 2 mm.



(a)



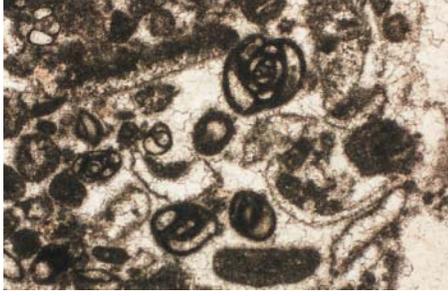
(b)



(c)



(d)



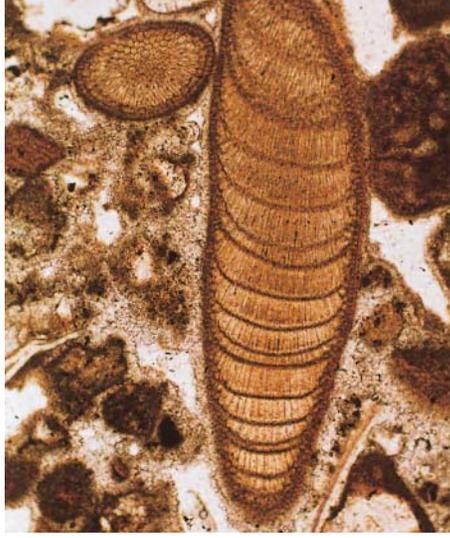
(e)

Plate 10

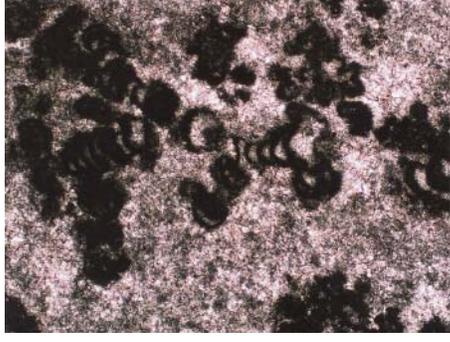
- (a)** Dasyclad algae. Capitan, Permian. Texas, USA. Field of view 4 x 2 mm.
- (b)** Calcareous red algae, *Lithothamnion* in longitudinal section (showing seasonal growth zones) and cross-section. The grains here are cemented by isopachous high-Mg calcite marine cement. Recent. Belize. Field of view 1.0 x 0.8 mm.
- (c)** Calcified microbes, *Renalcis*. Devonian. Guilin, China. Field of view 4 x 2 mm.
- (d)** Modern microbial mat composed of dolomite. The filaments of the cyanobacteria are clearly visible, but the dolomite crystals are submicroscopic. Recent. Bahamas. Field of view 4 x 2 mm.
- (e)** Stromatolite (microbial mat) composed of micrite laminae and laminoid fenestrae, with some intraclasts from desiccation. Carboniferous. Glamorgan, Wales. Field of view 5 x 4 mm.
- (f)** Stromatolite composed of micritic and grainy laminae, and small spar-filled fenestrae. Precambrian. Flinders, Australia. Field of view 4 x 2 mm.



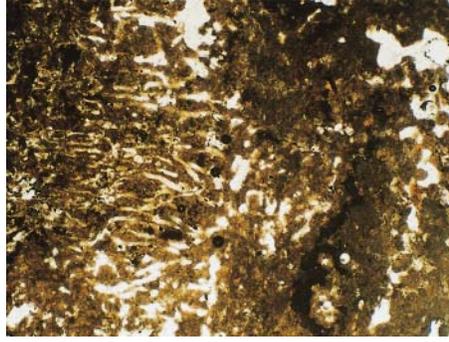
(a)



(b)



(c)



(d)



(e)



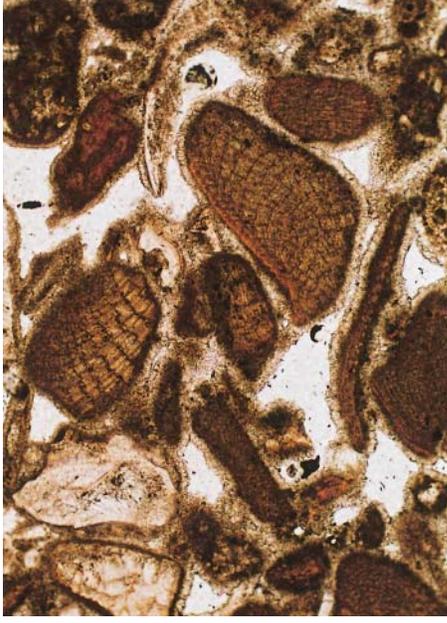
(f)

Plate 11

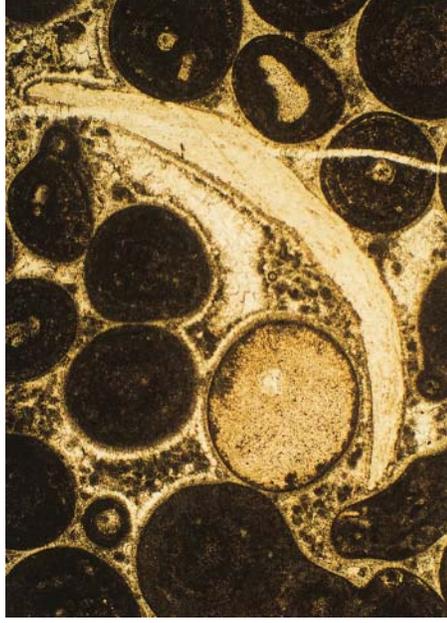
- (a)** Coral with aragonite cement — needles and a botryoid within the corallites, and lime mud internal sediment (black) and peloids in other pores. Crossed polars. Recent. Belize. Field of view 6 x 4 mm.
- (b)** Isopachous high-Mg calcite cement around skeletal grains (including calcareous red algae). Recent fore-reef debris. Belize. Field of view 1.2 x 0.8 mm.
- (c)** Oolitic grainstone with brachiopod and crinoid fragments (with thin micrite envelope — black) cemented by early isopachous fibrous marine cement (calcite), then some internal sediment of peloids, followed by drusy calcite spar cement. Carboniferous. Glamorgan, Wales. Field of view 6 x 4 mm.
- (d)** Hardground with ooids surrounded by thin isopachous marine cement fringe and then pore space filled by lime mud, now micrite. Cemented rock then cut by annelid borings, which later filled with quartz grains. Jurassic. Gloucestershire, England. Field of view 6 x 4 mm.



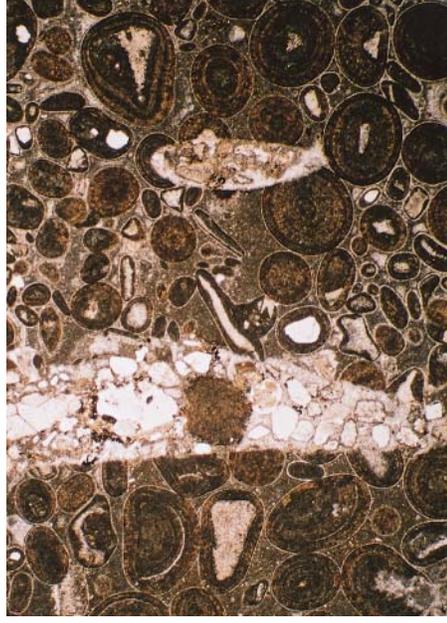
(a)



(b)



(c)



(d)

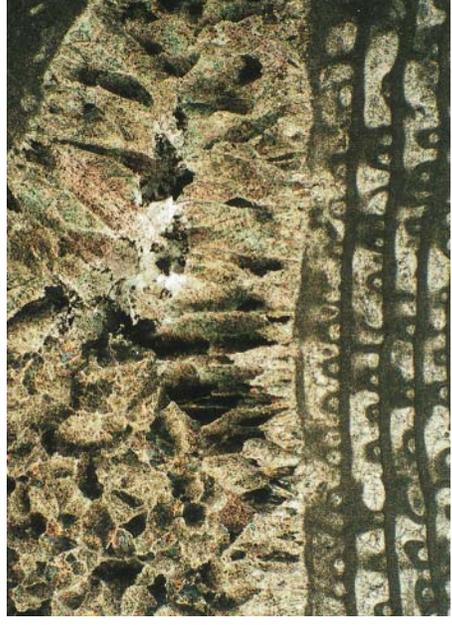
Plate 12

(a,b) Fusulinid foraminifer cemented by radial fibrous calcite. The crystals are columnar and cloudy with inclusions, and have undulose extinction under crossed polars. Small area of clear calcite spar: **(a)** plane-polarized light, field of view 6 x 4 mm; **(b)** crossed polars, field of view 3 x 2 mm. Capitan, Permian. Texas, USA.

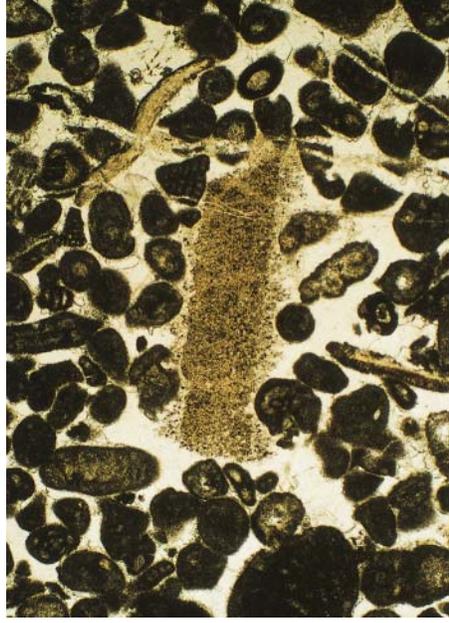
(c,d) Syntaxial calcite overgrowth cement on crinoid grain. Early part of overgrowth marine precipitate; clear later overgrowth a burial precipitate. Grains mainly peloids, micritized bioclasts and faecal pellets, showing concavo-convex/interpenetrative contacts, indicating some compaction. Very thin isopachous calcite cement fringes around grains seen in **(d)** are probably marine precipitates. Rock also cut by thin calcite veins. **(c)** Plane-polarized light; **(d)** Crossed polars. Cretaceous. Alps, France. Field of view 3 x 2 mm.



(a)



(b)



(c)



(d)

Plate 13

(a) Calcite cement at grain contacts and irregularly around grains, indicating near-surface, meteoric vadose environment. Later precipitation of large poikilotopic calcite (black, in extinction) took place during burial after some compaction. Crossed polars. Carboniferous. Glamorgan, Wales. Field of view 0.8×0.8 mm.

(b, c) Calcite spar.

(b) Calcite spar under plane-polarized light showing drusy fabric (crystal-size increase away from the substrate) and prominent twin planes. **(c)** Same field of view under cathodoluminescence showing delicate growth zones resulting from subtle variation in manganese and iron contents. Triassic. Glamorgan, Wales. Field of view 2×2 mm.

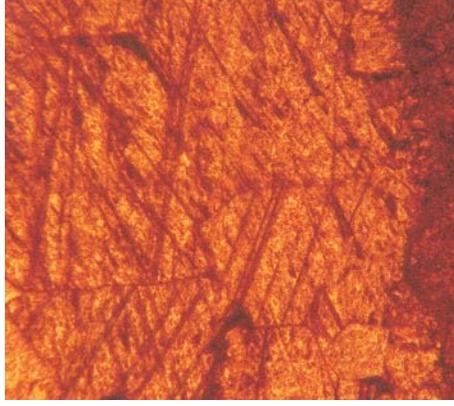
(d) Sutured (microstylolitic) contacts and concavo-convex contacts between grains (micritized ooids and bioclasts) and mechanical fracture of lower grain. Calcite spar between grains is a post-compaction burial cement. Jurassic. Burgundy, France. Field of view 2×2 mm.

(e) Microspar-pseudospar formed through aggrading neomorphism with fossil relics (brachiopod spine on right). Carboniferous. Yorkshire, England. Field of view 2×2 mm.

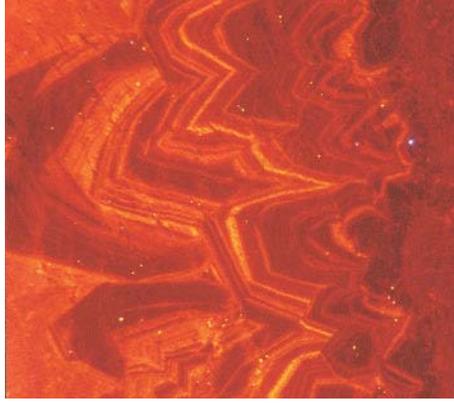
(f) Oolitic grainstone with scattered dolomite rhombs precipitated after early compaction (see grain contacts), before calcite spar cement. Carboniferous. Glamorgan, Wales. Field of view 2×2 mm.



(a)



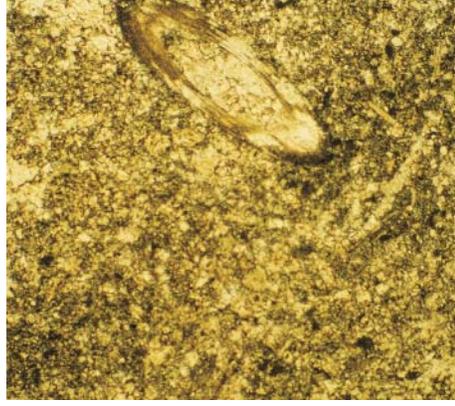
(b)



(c)



(d)



(e)



(f)

Plate 14

(a) Dolomitized oolite (no relics of original grains) with stylolite, highlighted by iron-rich clay. Xenotopic dolomite (anhedral crystals) below stylolite and idiotopic dolomite (euhedral crystals) above.

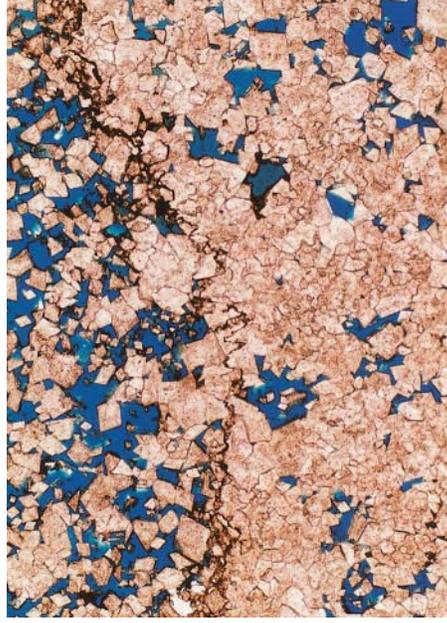
Intercrystalline porosity shown through impregnation with blue resin. Arab Formation, Offshore UAE. Field of view 3×2 mm.

(b) Dolomitized grainstone with moderate preservation of original ooids. Intercrystalline porosity is present, shown up by impregnation with blue resin. Cretaceous, Offshore Angola. Field of view 6×4 mm.

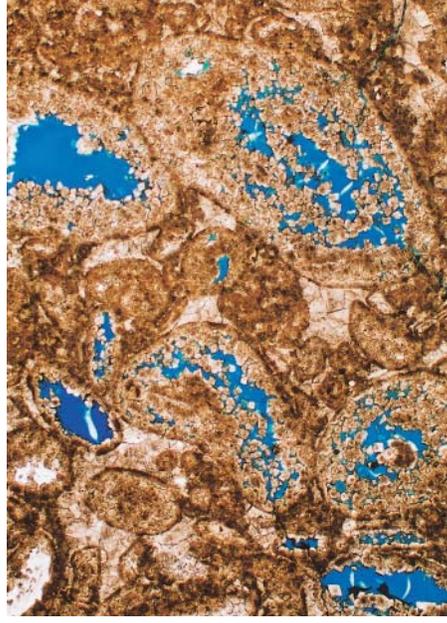
(c) Baroque dolomite: coarse crystals with undulose extinction. Rock is an oolite but there are no relics. Crossed polars. Carboniferous. Glamorgan, Wales. Field of view 2×2 mm.

(d) Dedolomite: crinoidal grainstone with overgrowths containing scattered dolomite rhombs that have been replaced by calcite. The dark material is iron oxide/hydroxide, suggesting the dolomite was originally ferroan. Carboniferous. Northumberland, England. Field of view 2×2 mm.

(e) Dolomite moulds: grainstone with scattered dolomite rhombs that have been dissolved out to give a good porosity, as shown by blue resin. Stylolitic contacts between grains. Jurassic. Burgundy, France. Field of view 0.8×0.8 mm.



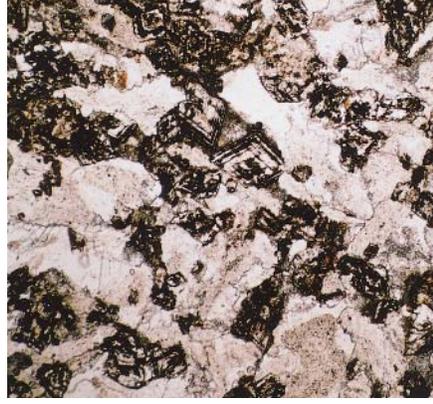
(a)



(b)



(c)



(d)



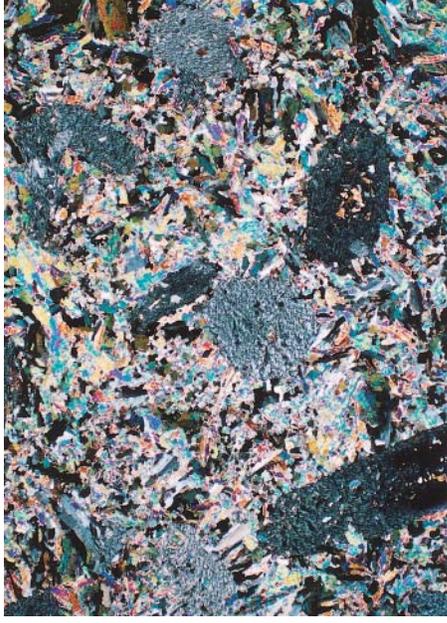
(e)

Plate 15

(a) Anhydrite: minute crystals of anhydrite forming nodules with some clay sediment between (brown) and large replacement anhydrite crystals. Crossed polars. Permian. Cumbria, England. Field of view 6 × 4 mm.



(b) Anhydrite and secondary gypsum: small crystals and laths of anhydrite (bright colours) being replaced by large, low birefringence, porphyrotropic gypsum crystals. Crossed polars. Permian (Zechstein). Teesside, England. Field of view 6 × 4 mm.

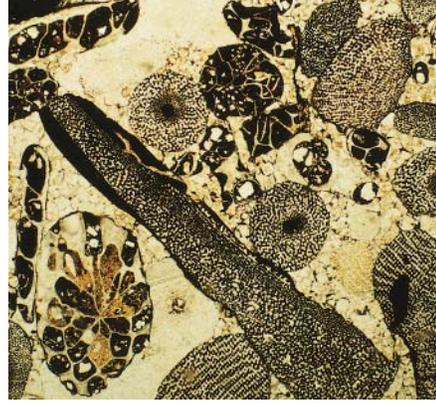


(c, d) Hematitic ironstone with hematite-impregnated crinoid and bryozoan fragments in a calcite cement. **(c)** plane-polarized light; **(d)** crossed polars.

(b)

(a)

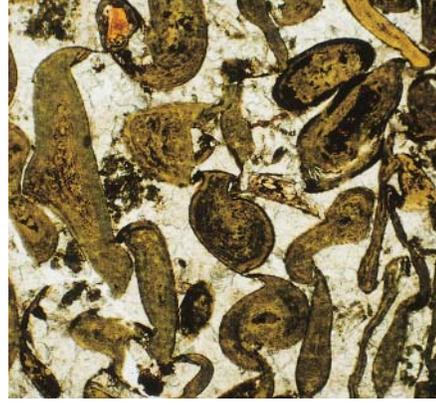
Bioclastic grainstone, Carboniferous Rhiwbina Ironstone, Glamorgan, Wales. Field of view 2 × 2 mm.



(c)



(d)



(e)

(e) Berthierine-chamosite ooids showing distorted shapes and elephantine features in calcite cement. Plane-polarized light. Jurassic. Raasay, Scotland. Field of view 2 × 2 mm.

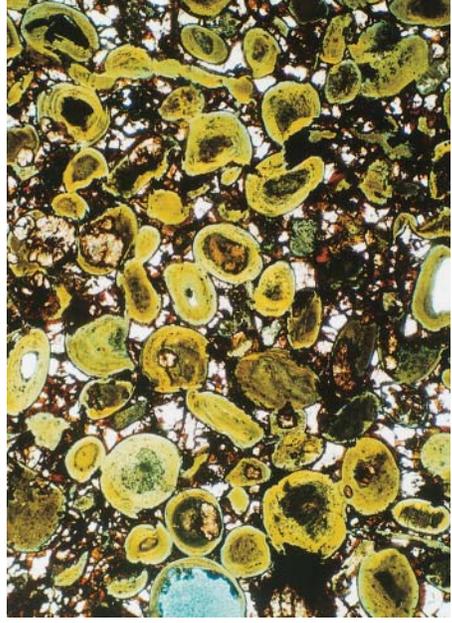
Plate 16

(a) Berthierine-chamosite ooids, with some shape distortion, in a siderite cement partly altered to goethite (brown). Plane-polarized light. Jurassic. Yorkshire, England. Field of view 3 x 2 mm.

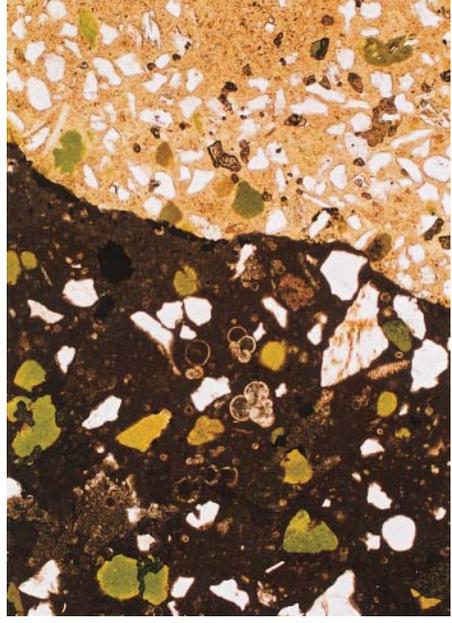
(b) Glauconite grains (green) in dark chalk with planktonic foraminifers and in a pebble (on right) of phosphate (pale brown), with angular quartz grains in both. Plane-polarized light. Cretaceous. Bornholm, Denmark. Field of view 3 x 2 mm.

(c, d, e) Phosphorites: all plane-polarized light; all fields of view 2 x 2 mm. **(c)**

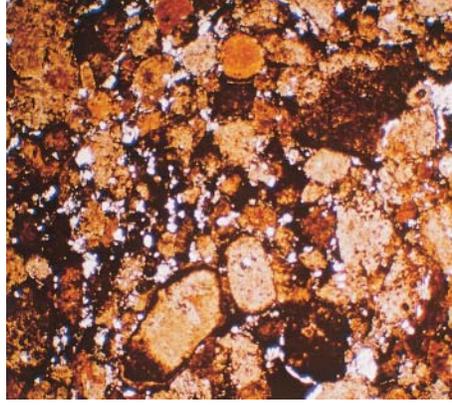
Phosphorite pellets. Permian. Idaho, USA. **(d)** Fish scales and bones in phosphorite mudstone. Permian. Idaho, USA. **(e)** Bone fragments. Triassic. Gloucestershire, England.



(a)



(b)



(c)



(d)



(e)

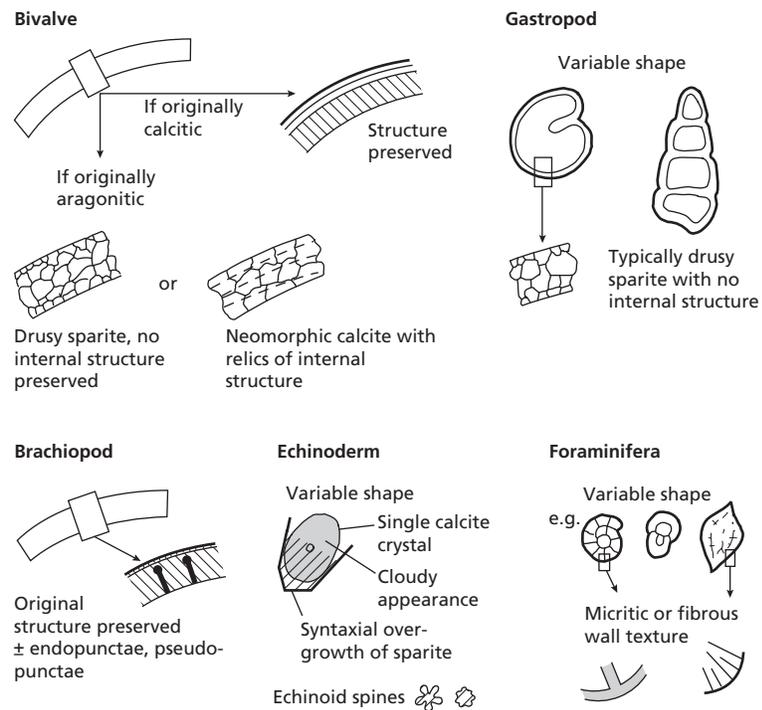


Fig. 4.7 Typical thin-section appearance of bivalve, gastropod, brachiopod, echinoderm and foraminiferal skeletal grains in limestone.

internal structure (growth lines) are preserved (see Plate 8a), and there are minute inclusions of aragonite left in the calcite (see Section 4.7). Calcitic bivalves normally will retain their original structure and the most common types are foliaceous (thin parallel sheets) and prismatic. Bivalve fragments in thin-section will be seen as elongate, rectangular to curved grains, typically disarticulated.

Gastropods are ubiquitous throughout shallow-marine environments. They also occur in vast numbers, but low species diversity, in hypersaline and brackish waters, such as on tidal flats and in estuaries, because certain species are able to tolerate fluctuations and extremes of salinity. Most gastropods are benthic, vagile creatures. The encrusting vermetiform gastropods, often confused with serpulids, form reef-like structures in the tropics and Carboniferous. The small, conical pteropods are important in Cenozoic pelagic sediments.

The majority of gastropods have shells of aragonite with similar internal microstructures to bivalves. The internal microstructure of fossil gastropods also is rarely seen because the original aragonite is mostly dissolved out and the void filled by calcite cement. Gas-

tropod fragments can be recognized easily under the microscope by their shape, although this is very much dependent on the plane of section (Fig. 4.7 and Plate 8b). Gastropods may resemble certain foraminifers but the latter are usually much smaller and composed of dark micritic calcite. (See Plate 9c,d,e.)

Of the *cephalopods*, nautiloids and ammonoids are relatively common in limestones of the Palaeozoic and Mesozoic and belemnites occur in Mesozoic limestones. They were wholly marine animals with a dominantly nektonic or nekto-planktonic mode of life, as with the modern *Nautilus*, octopus and cuttlefish. The cephalopods are more common in pelagic, relatively deep-water deposits. Examples include the Ordovician–Silurian *Orthoceras* limestones of Sweden, the Devonian Cephalopodenkalk and Griotte of western Europe and the Jurassic Ammonitico Rosso of the European Alpine region. Nautiloid and ammonoid shells were originally aragonitic and so in limestones they are typically composed of calcite spar with little internal structure. The shape, normally large size and presence of septa are the features to note. Belemnite guards were made of calcite and have a strong radial-fibrous fabric in cross-section.

Brachiopods

Brachiopods are particularly common in Palaeozoic and Mesozoic limestones of shallow-marine origin. These were largely benthic, sessile organisms; a few species were infaunal. Only in rare cases, such as in the Permian of west Texas, did the brachiopods contribute to reef development. At the present time brachiopods are an insignificant group of marine invertebrates.

Although in section brachiopod shells are similar to those of bivalves in shape and size, most articulate brachiopods were composed of low-Mg calcite, so that the internal structure is invariably well preserved. The common structure is a very thin, outer layer of calcite fibres orientated normal to the shell surface, and a much thicker, inner layer of oblique fibres (Fig. 4.7). It can be difficult to distinguish brachiopod shells from bivalves with foliaceous calcite. However, certain brachiopods have modifications to the shell, with punctae and pseudopunctae. In the punctate brachiopods, such as the terebratulids, fine tubes (endopunctae) perpendicular to the shell surface perforate the inner layer and are filled with sparite or micrite. Pseudopunctae, as occur in the strophomenid group, are prominent rod-like prisms within the shell. Inarticulate brachiopods, composed mostly of chitin or chitinophosphate, are rare in limestones. (See Plates 8c,d & 11c.)

Cnidaria (especially corals)

The Cnidaria include the Anthozoa (corals), of which two ecological groups exist today: hermatypic corals that contain symbiotic dinoflagellate algae (zooxanthellae) in their polyps and ahermatypic corals without such algae. Because of the algae, hermatypic corals require shallow, warm and clear seawater. They are the reef-forming corals at the present time, being mainly responsible for the reef framework, which is reinforced by red algae. Ahermatypic corals can occur at much greater depths and tolerate colder waters. They locally form build-ups. The rugose and tabulate corals were important in Silurian and Devonian reefs, and many Triassic reefs contain scleractinian corals. Some of the latter may well have been ahermatypic. Corals, both solitary and colonial, and coral debris occur in many non-reefal limestones.

The Palaeozoic rugose and tabulate corals were

composed of calcite, most probably high-Mg calcite, so that preservation generally is very good (see Plate 9a). Scleractinian corals (Triassic to Recent), on the other hand, have aragonitic skeletons and so normally are poorly preserved in limestones (see Plate 9b). Identification of coral is based on such internal features as septa and, where present, other internal plates in the Rugosa and Scleractinia, and tabulae in the Tabulata. Corallite form and colonial organization are also important. The microstructure of Palaeozoic calcitic and later aragonitic corals is very similar, chiefly consisting of fibres in spherulitic or parallel arrangements, which form linear structures called trabeculae, or sheets.

Echinodermata

Echinoderms are wholly marine organisms that include the echinoids (sea-urchins) and crinoids (sea-lilies). In modern seas, echinoids inhabit reef and associated environments, locally in great numbers, but crinoids are restricted to deeper waters and are insignificant as producers of carbonate sediment. In the Palaeozoic and Mesozoic, fragments of echinoderms, especially the crinoids, are a major constituent of bioclastic limestones. Many deep-water limestone turbidites are composed of crinoidal debris, derived from shallow platforms.

Echinoid and crinoid skeletons are calcitic; modern forms generally have a high Mg content. Echinoderm fragments are easily identified because they are composed of large, single calcite crystals, individual grains thus showing unit extinction. In many cases, a sparite cement crystal has grown syntaxially around the echinoderm fragment (Fig. 4.7). Echinoderm grains have a dusty appearance, especially relative to a sparite cement overgrowth, and they may show a porous structure filled with micrite or sparite. (See Plates 9c, 12a,b & 15c,d.)

Bryozoa

Although these small, colonial marine organisms are significant suppliers of carbonate sediment only locally at the present time (notably to cool-water carbonates, as off southern Australia), they have in the past contributed to the formation of reef and other limestones, particularly in the Palaeozoic. Examples include the Mississippian mud-mounds of

southwestern USA and Europe, the Permian reefs of Texas and western Europe, and the Danian Chalk of Denmark.

Modern bryozoan skeletons are composed of either aragonite or calcite (commonly high-Mg calcite) or a mixture of both. There are many types of bryozoans but the fenestrate variety, including the fenestrellids, are seen most frequently in sections of Palaeozoic limestones. The skeleton consists of foliaceous calcite with round holes (the zooecia, where the individuals of the colony used to live) filled with sparite or sediment (see Plate 15c).

Foraminifera

Foraminifera are dominantly marine Protozoa, mostly of microscopic size. Planktonic foraminifers dominate some pelagic deposits, such as the *Globigerina* oozes of ocean floors and some Cretaceous and Tertiary chalks and marls. Benthic foraminifers are common in warm, shallow seas, living within and on the sediment, and encrusting hard substrates.

Foraminifera are composed of low- or high-Mg calcite, rarely aragonite. Foraminifera are very diverse in shape but in section many common forms are circular to subcircular with chambers. The test wall is dark and microgranular in many thin-walled foraminifers such as the endothyracids and miliolids, but light-coloured and fibrous in larger, thicker species, such as the rotalids, nummulitids and orbitolinids. (See Plate 9c,d,e.)

Other carbonate-forming organisms

There are many other organisms that have calcareous skeletons but contributed in only a minor way to limestone formation, or were important for only short periods of geological time.

Sponges (Porifera). Spicules of sponges, which may be composed of silica or calcite, occur sporadically in sediments from the Cambrian onwards. The importance of spicules is as a source of silica for the formation of chert nodules and silicification of limestones (Section 4.8.4). At times, sponges provided the framework for reefs and mounds; examples include lithistid sponges in the Ordovician, calcisponges in the Permian of Texas and Triassic of the Alps, and silicisponges (now calcitized) in the Jurassic of southern Germany. Sclerosponges are the dominant reef-forming organ-

ism in some modern Caribbean reefs. *Stromatoporoids*, once considered hydrozoans, are now classified as a subphylum of the Porifera. Stromatoporoids were marine colonial organisms that had a wide variety of growth forms, ranging from spherical to laminar, depending on species and environmental factors. Stromatoporoids were a major reef organism in the Silurian and Devonian, commonly in association with rugose and tabulate corals, and grew up to a metre or more in size. *Archaeocyathids*, also probably sponges, formed reefs in the lower Cambrian of North America, Morocco, Siberia and South Australia.

Arthropods. Of this group, the *ostracods* (Cambrian to Recent) are locally significant in Tertiary limestones and some others. They live at shallow depths in marine, brackish or freshwater environments. Ostracods have small (around 1 mm in length), thin bivalved shells, smooth or ornamented, composed of calcite with a radial-fibrous structure. *Trilobites* (Cambrian to Permian), with a similar skeletal structure, occur locally in Palaeozoic shelf limestones, but never in rock-forming quantities.

Calcispheres. These are simple spherical objects, up to 0.5 mm in diameter, composed of calcite (usually sparite), in some cases with a micritic wall. They are probably some form of alga, although an affinity with Foraminifera has been suggested. They occur in many Palaeozoic limestones, particularly fine-grained micrites of back-reef or lagoonal origin.

4.3.3 The contribution of algae and microbes to limestones

Algae and microbes make a major contribution to limestones by providing skeletal carbonate particles, trapping grains to form laminated sediments and attacking particles and substrates through their boring activities. Many of the Precambrian limestones were at least in part produced by microbes and algal-microbial limestones are widely distributed throughout the Phanerozoic. Four groups of algae are important: red algae (Rhodophyta), green algae (Chlorophyta), yellow-green algae (Chrysophyta) and cyanobacteria (formerly blue-green algae). Relevant texts are Walter (1976), Flügel (1977), Wray (1977), Monty (1981), Toomey & Nitechi (1985), Riding (1990) and Riding & Awramik (2000).

Rhodophyta (red algae)

Calcareous algae of the Rhodophyta, such as the Corallinaceae (Carboniferous to Recent) and Solenoporaceae (Cambrian to Miocene) have skeletons composed of cryptocrystalline calcite, which is precipitated within and between cell walls. In section, a regular cellular structure is present (see Plates 10b & 11b). Modern coralline algae have a high Mg content in the calcite, which is related to water temperature (higher values for a given species in warmer waters). Many of the coralline algae encrust substrates and if this is a pebble or shell then nodules, referred to as *rhodoliths*, develop. Encrustations may be massive and rounded, or delicately branched, depending on ecological factors. One of the most important roles of these red algae is in coating, binding and cementing the substrate, particularly in modern reefs. In temperate and arctic carbonate sands, the red alga

Lithothamnion is a major contributor. In the Palaeozoic and Mesozoic, red algae of the Solenoporaceae are locally abundant and in some cases participate in reef formation.

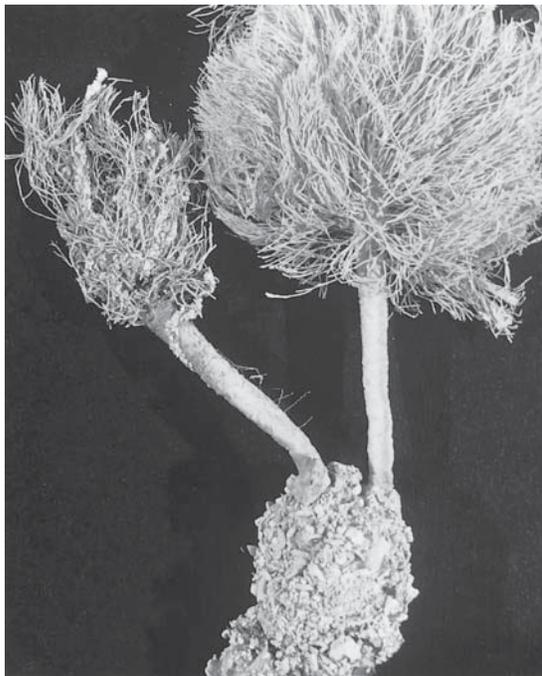
Chlorophyta (green algae)

Three algal groups are important: the Codiaceae, Dasycladaceae (both Cambrian to Recent) and Characeae (Silurian to Recent). The Characeae are incompletely calcified (low-Mg calcite), so that only stalks and reproductive capsules are found in limestones. Modern (and fossil) forms are restricted to fresh or brackish water.

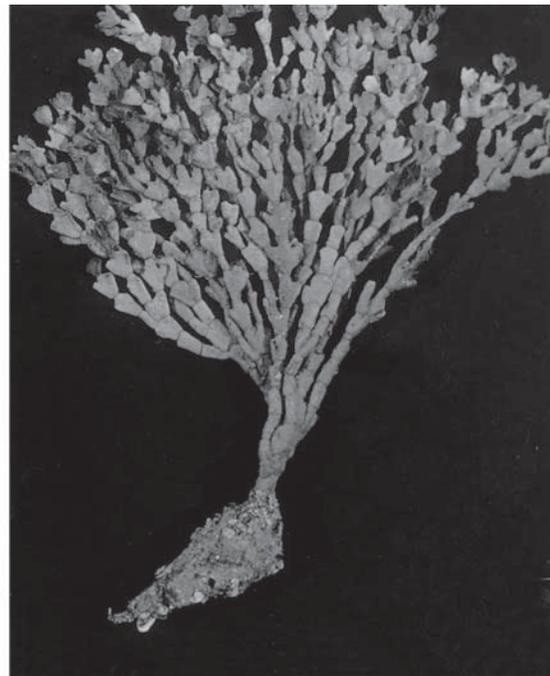
With the dasyclad algae, calcification is also incomplete and involves the precipitation of an aragonitic crust around the stem and branches of the plant. Dasyclads are marine algae that tend to occur in shallow, protected lagoonal areas of the tropics. Under the microscope, circular, ovoid or elongate shapes are seen, representing sections through the stems or branches (see Plate 10a).

The codiacean algae include *Halimeda* and *Penicillus* (Fig. 4.8), two common genera of Caribbean and

Fig. 4.8 Calcareous green algae. (a) *Penicillus*, which on death and disintegration produce micron-sized aragonite needles. (b) *Halimeda*, which gives rise to sand-sized grains. Recent. Florida Keys, USA.



(a)



(b)

Pacific reefs and lagoons. *Halimeda* is a segmented plant that on death and disintegration generates coarse, sand-sized particles. In thin-section, they look like Swiss cheese. *Penicillus*, the shaving-brush alga, is less rigid, consisting of a bundle of filaments coated in needles of aragonite. Death of this and other algae provides fine-grained carbonate sediment (lime mud or micrite, Section 4.3.4).

Phylloid algae are a group of late Palaeozoic algae that have a leaf or potato-crisp shape. Some belong to the codiaceans, whereas others have more affinities with red algae. Phylloid algae are one of the main components of shelf-margin mud-mounds in the Upper Carboniferous to Lower Permian of southwest USA.

Chrysophyta (coccoliths)

Coccolithophorids (Jurassic to Recent) are planktonic algae that have a low-Mg calcite skeleton consisting of a spherical coccosphere (10–100 μm diameter) composed of numerous calcareous plates, called coccoliths. In view of their size, these algae are studied with the scanning electron microscope. The coccoliths are chiefly disc-shaped, commonly with a radial arrangement of crystals (Fig. 4.9). Coccoliths are a significant

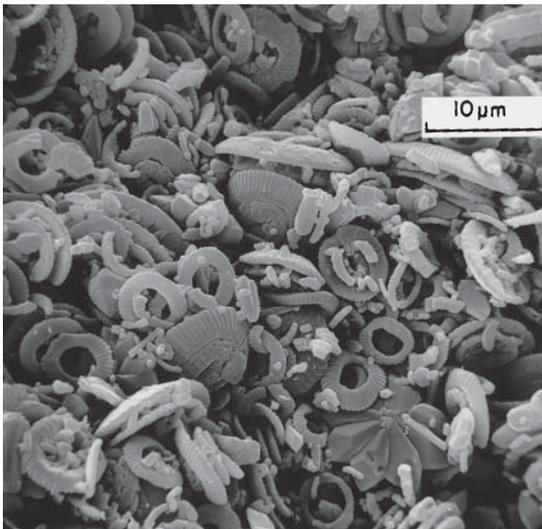


Fig. 4.9 Scanning electron micrograph of coccoliths from pelagic ooze, Shatsky Rise, northwest Pacific. Also present (lower right) is a larger discoaster, another type of nanoplanktonic alga.

component of modern deep-water carbonate oozes, particularly those of lower latitudes. They form the bulk of Tertiary and Cretaceous chalks and occur in red pelagic limestones of the Alpine Jurassic (see Fig. 4.17).

Micrite envelopes

A large proportion of skeletal fragments in modern and ancient carbonate sediments possess a dark *micrite envelope* around the grains (see Plates 7b,c,d, 8b & 11c). The envelope is produced mostly by endolithic bacteria that bore into the skeletal debris. Following vacation of the microbores (5–15 μm in diameter), they are filled with micrite. Repeated boring and filling result in a dense micrite envelope, which is the altered outer part of the skeletal grain. This process of grain degradation may eventually produce a totally micritized grain, i.e. a *peloid* (Section 4.3.1), devoid of the original skeletal structure (see Plates 7a, 8c, 12c & 13d). Some bioclasts, especially the corals, molluscs and foraminifers, are more susceptible to microboring than others, and this has implications for the preservation of grain assemblages (Perry, 1998). Inorganic recrystallization of grains on the sea floor also takes place, destroying the original skeletal structure (Reid & Macintyre, 1998).

Many other organisms bore into skeletal grains and carbonate substrates; examples include clionid sponges (in Plate 7e), bivalves (such as *Lithophaga*, see Plate 9b and Fig. 2.42d), and polychaetes, all of which produce larger borings and cavities, and fungi, which produce bores of 1–2 μm in diameter. The micrite filling the borings may be precipitated physico-chemically or biochemically through decomposition of the microbes. Micrite envelopes due to endolithic cyanobacteria can be used as a depth criterion, indicating deposition within the photic zone (less than 100–200 m), but the grains can be transported to greater depths.

Stromatolites and microbialites

An important role of cyanobacteria (with other organisms such as diatoms, fungi and nematodes) is the formation of *microbial mats*, formerly called algal mats. These organic mats occur on sediment surfaces in many low- to mid-latitude, marine and non-marine environments, from moderate depth subtidal through



Fig. 4.10 Tidal-flat microbial mats from Abu Dhabi, Arabian Gulf. The microbial mats (dark areas) are desiccated into small polygons and have a thin covering of recently deposited carbonate sediment. Large polygonal desiccation cracks occur in the slightly higher, partly lithified gypsiferous carbonate sediment (light area). Knife (circled) for scale.



Fig. 4.11 Columnar stromatolites from agitated intertidal zone, Shark Bay, Western Australia. Columns increase in size upwards and join with adjacent columns to form large domal structures.

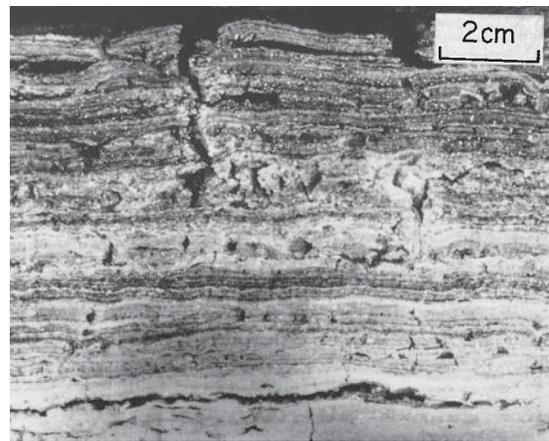


Fig. 4.12 Microbially laminated carbonate sediments from an intertidal flat. Shark Bay, Western Australia. Desiccation cracks and cavities (fenestrae) are present.

to supratidal marine areas, and fresh to hypersaline lakes and marshes. They form planar sheets, columns and domes (Figs 4.10–4.13). The cyanobacteria are mainly filamentous varieties, common mat-forming genera being *Lyngbya*, *Microcoleus*, *Schizothrix*, *Scytonema* and *Oscillatoria*, although unicellular coccoid forms such as *Endophysalis* also occur. A mat usually has a specific community that together with environmental factors produces a mat of particular morphology and structure. Areas where microbial mats of various types are developing today include the Bahamas, the Arabian Gulf and Shark Bay, Western Australia.

The cyanobacteria are mucilaginous and this, together with their filamentous nature, results in the trapping and binding of sedimentary particles to produce a laminated sediment, a *microbialite* or *stromatolite* (Figs 4.12 & 4.14). Stromatolites occur throughout the geological record but are particularly important in the Precambrian, where they have been used for stratigraphic correlation. The lamination in many modern intertidal mats consists of couplets of dark organic-rich layers alternating with light, sediment-rich laminae. Microbial filaments may be

Fig. 4.13 Stromatolite domes and columns in a few metres of water in a high-energy tidal channel. Exuma Cay, Bahamas.



Fig. 4.14 Stromatolite consisting of a large dome with internal crinkly laminae. Late Precambrian. Anti-Atlas, Morocco.

preserved (see Plate 10d). Laminae are usually less than several millimetres thick, but some sediment laminae may reach a centimetre or more. The alternating laminae reflect growth of the microbial mat (organic layer) followed by sedimentation, and then trapping and binding of the sediment particles into the mat, as the microbial filaments grow through to form a new organic layer at the surface once more. In ancient stromatolites, the laminae are usually alterations of dense micrite, perhaps dolomitized, and grains, such as peloids and fine skeletal debris (see Plate 10f). Microbially laminated sediments commonly show small corrugations and irregularities in thickness, which serve

to distinguish them from laminae deposited purely by physical processes. There may be evidence of desiccation—broken laminae, intraclasts and laminoid fenestrae (see Plate 10e). The laminae may also constitute larger-scale domes and columns. A diurnal growth pattern has been demonstrated for the laminae in some subtidal mats, but in other cases, such as on tidal flats and in ephemeral lakes, mat growth is probably seasonal or related to periodic wettings, and sedimentation is erratic, being controlled largely by storm floodings.

Microbial mats give rise to a range of laminated structures (microbialites). The simplest are planar stromatolites (or microbial laminites). They typically develop on protected tidal flats and so may show desiccation polygons (Fig. 4.10) and contain *laminoid fenestrae* (elongate cavities: Fig. 4.12, Plate 10e and Section 4.6.3) and evaporite minerals or their pseudomorphs. Domal stromatolites, where the laminae are continuous from one dome to the next, occur on the scale of centimetres to metres. Columnar stromatolites are individual structures, which may be several metres high (Figs 4.13 & 4.14). Complex stromatolites, such as occur in Precambrian strata, may be combinations of domes and columns, such as a large columnar structure with linked domes internally, or they may show branching of columns. They can form build-ups many tens of metres high and laterally be very extensive.

One further type of microbial-sediment structure is a nodule (ball), or *oncoïd* (Fig. 4.15). Some have an internal concentric lamination, which may be asym-

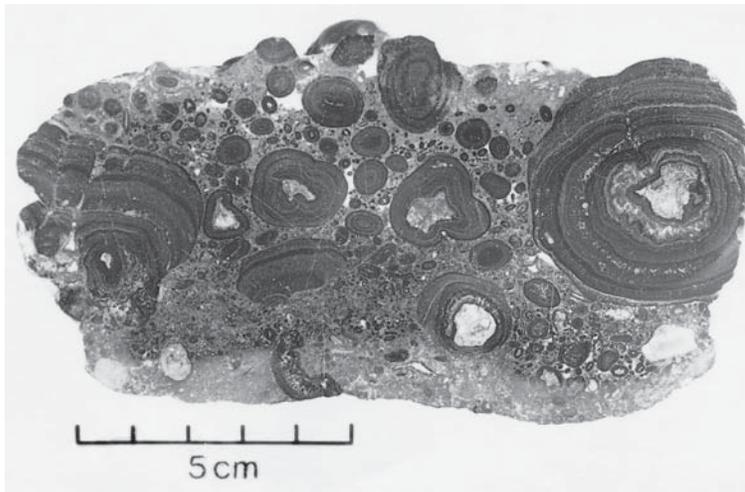


Fig. 4.15 Oncoids: spherical to irregular microbial balls. Some showing marked asymmetry, through periods of stationary growth. Carboniferous. Fife, Scotland.

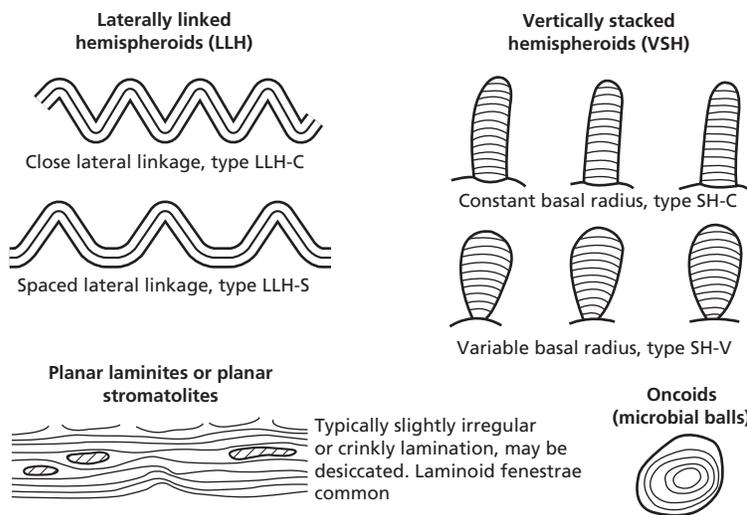


Fig. 4.16 Common types of stromatolite with terminology.

metric. Many oncoids are composed of dense micrite, whereas others are more clotted. The latter fabric also occurs in stromatolites where ones with a poor lamination and a clotted texture are called *thrombolites*. They probably have been formed by coccoid cyanobacteria.

In a notation for describing stromatolites (Fig. 4.16), domal stromatolites are referred to as laterally linked hemispheroids (LLH), columnar stromatolites are vertically stacked hemispheroids (VSH) and oncoids are spherical structures (SS). There is also a binomial classification with form genera and species, particu-

larly used for Precambrian examples. Common genera include *Conophyton*, *Collenia* and *Cryptozoon*.

The morphological variation of microbialites depends largely on environmental factors such as water depth, tidal and wave energy, frequency of exposure and sedimentation rate. For example, large columnar structures in Shark Bay (Fig. 4.11) are restricted to intertidal and subtidal areas in the vicinity of headlands; small columns and domes occur in less agitated bay waters, and low domes and planar mats dominate protected tidal flats (Fig. 4.10). In the Bahamas, large stromatolite columns (almost 'reefs') are growing in

shallow, high-energy tidal channels (Fig. 4.13; Macintyre *et al.*, 1996). The microstructure of microbial mats and stromatolites is also variable; it is thought largely to be a reflection of the microbial community, as noted above.

At the present time, microbial mats are quite restricted in their occurrence, being developed extensively only in hypersaline tidal-flat and freshwater environments. Planar stromatolites are a feature of peritidal facies throughout the geological record. The dearth of domal and columnar stromatolites in the Phanerozoic, notably in shallow, normal-marine facies, is attributed to the grazing activities of organisms, especially the gastropods. The absence of such metazoans in the Precambrian and earliest Phanerozoic was one of the main factors in the widespread and diverse stromatolite development at that time, including many in subtidal and deeper-water settings.

Modern marine microbial mats are largely unlithified, whereas those of fresh and hypersaline waters may be cemented through biochemical or physicochemical precipitation of carbonate. There are particular microbes, common in the Palaeozoic, that do have calcified filaments. The group name Porostromata has been applied to these, which include *Girvanella*, *Ortonella*, *Garwoodia* and *Cayeuxia*. The structures consist of tubes or filaments with micritic walls, considered to be calcified sheaths, arranged in an irregular spaghetti-type or more ordered, radial fashion. They typically form nodules (skeletal oncoids) but also stromatolites. In some cases, the algae are associated intimately with encrusting foraminifers, as in *Osagia* and *Sphaerocodium* nodules. More bush-like, branching structures (dendrolites) of calcified microbes are common in the early Palaeozoic, commonly forming reefal structures, and include *Renalcis* and *Epiphyton* (see Plate 10c).

For a discussion of the geological history of microbial carbonates see Riding (2000), and papers in Riding & Awramik (2000).

4.3.4 Lime mud and micrite

Many grainy limestones have a fine, usually dark matrix and many others are composed entirely of fine-grained carbonate. This material is *micrite* (microcrystalline calcite), with a grain size generally less than 4 μm . Electron microscope studies have shown that the micrite is not homogeneous but has areas of finer or

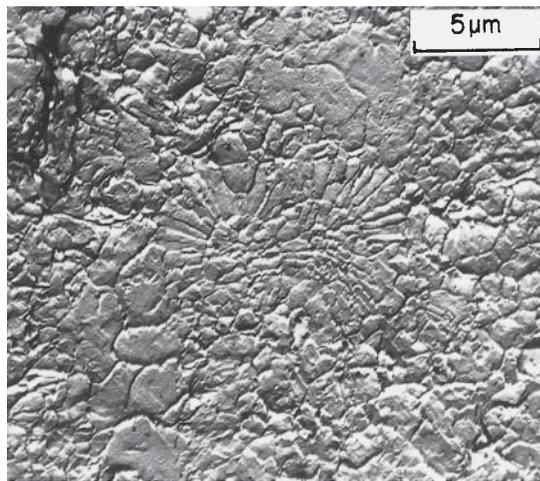


Fig. 4.17 Transmission electron micrograph of a Jurassic micritic pelagic limestone showing variation in size and shape of micrite crystals, and two coccoliths consisting of elongate, radially arranged crystals. Ammonitico Rosso, Jurassic. Austrian Alps.

coarser crystals, and intercrystalline boundaries that may be planar, curved, irregular or sutured (Fig. 4.17). Micrites are susceptible to diagenetic alteration and may be replaced by coarser mosaics of *microspar* (5–15 μm) through aggrading neomorphism (Section 4.7.4).

Carbonate muds are accumulating in many modern environments, from tidal flats and shallow lagoons to the deep-sea floor. There are many possible sources of lime mud. Carbonate muds of south Florida and the Bahama Platform have been the subject of much research. They occur in the shallow subtidal, less-agitated central parts of the platform to the west of Andros Island, and in lagoons, such as the Bight of Abaco. Fine carbonate sediments also occur on tidal flats and on the slopes and in deep basins around the platform ('periplatform ooze'). The mud in many subtidal areas consists predominantly of aragonite needles and laths a few microns in length; only some 20% of the sediment is recognizably biogenic. Inorganic precipitation as a result of evaporation has been postulated for the Great Bahama Bank. The occasional 'whiting', a sudden milkiness of the sea resulting from suspended aragonite needles, may be the actual inorganic precipitation taking place, although stirring of bottom muds by shoals of fish can produce the same effect (see Shinn *et al.*, 1989). The disintegration of calcareous

green algae has been widely regarded as a major process of lime-mud production, from studies in South Florida and the Bight of Abaco. When algae such as *Penicillus* (Fig. 4.8a) break down, a vast quantity of aragonite is released. Measurements of growth rates and calculations of the mass of aragonite produced have shown that sufficient mud is produced to account for all the fine-grained sediments. Indeed, it appears that there is an overproduction so that algal disintegration in lagoons could be the source of mud for neighbouring tidal flats and peri-platform areas (Fig. 4.18). Hence the shallow subtidal zone is referred to as the 'carbonate factory'. However, detailed studies with the SEM and trace elements (e.g. Milliman *et al.*, 1993) have concluded that much of the lime mud on the Great Bahama Bank is a direct precipitate, whereas that of Florida Bay is largely green algal in origin. The Sr/Mg ratio of algae is less than two, whereas it is more than four in inorganic aragonite.

Three other processes that produce lime mud, but in variable or limited quantities are:

- 1 bioerosion, where organisms such as boring sponges and microbes attack carbonate grains and substrates;
- 2 mechanical breakdown of skeletal grains through waves and currents;
- 3 biochemical precipitation through microbial photosynthesis and decomposition.

Carbonate mud, largely of skeletal origin, forms

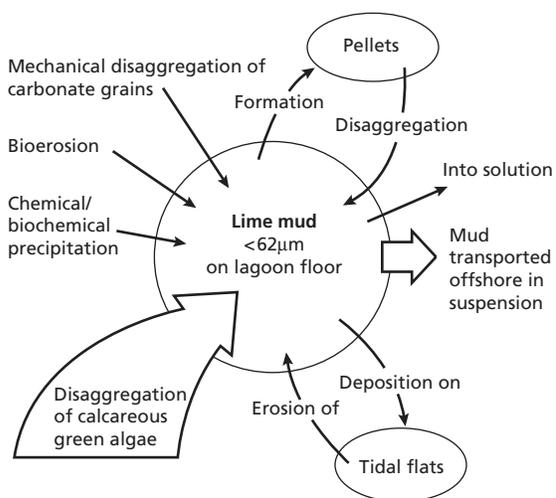


Fig. 4.18 Lime-mud budget for a lagoon in the Bahamas.

subtidal banks in Florida and Shark Bay, where sea-grasses and algae trap and bind the sediment (e.g. Bosence, 1995). Carbonate muds of the deep-ocean floors are oozes composed chiefly of coccoliths, with larger foraminiferal and pteropod grains.

In the lagoons along the Trucial Coast of the Arabian Gulf, inorganic precipitation is probably taking place. The aragonite needle muds contain high strontium values (9400 p.p.m.), close to the theoretical for direct precipitation from that lagoonal water. There is also a paucity of calcareous algae in the region and the possibility of other aragonitic skeletons contributing is precluded by their low strontium values.

There usually is little evidence in a limestone for the origin of the micrite. Nanofossils, in particular coccoliths, can be recognized with the electron microscope in some pelagic limestones (Fig. 4.17), but on the whole there is little to indicate a biogenic origin for the majority of ancient micrites. From the studies of modern lime muds, it is tempting to suggest that many ancient shallow-marine micrites were the product of calcareous green-algal disintegration. However, the possibility of inorganic precipitation in the past cannot be ruled out.

The original mineralogy of lime muds will have been important in their diagenesis; those with a high content of aragonite will have been more susceptible to neomorphism and microspar formation. Original aragonite-dominated precursor muds can be recognized through SEM study of lightly etched micrites, revealing minute crystal relics of aragonite (Lasemi & Sandberg, 1984). In grainstones and other coarse limestones, micrite could well be a cement, rather than a matrix (Section 4.7.1). In addition, fine carbonate sediment up to silt grade may filter into a porous limestone soon after deposition or during early diagenesis. Such geopetal or internal sediment can be recognized by its cavity-filling nature (see Plates 9a,b & 11a,c,d).

4.4 Classification of limestones

Three classification systems are currently used, each with a different emphasis, but the third, that of Dunham, based on texture, is now used more widely.

- 1 A very simple but often useful scheme divides limestones on the basis of grain size into calcirudite (most grains > 2 mm), calcarenite (most grains between 2 mm and $62\mu\text{m}$) and calcilutite (most grains less than $62\mu\text{m}$).

2 The classification scheme of R.L. Folk (Fig. 4.19), based mainly on composition, distinguishes three components: (a) the grains (allochems), (b) matrix, chiefly micrite and (c) cement, usually drusy sparite. An abbreviation for the grains (*bio*—skeletal grains, *oo*—ooids, *pel*—peloids, *intra*—intraclasts) is used as a prefix to micrite or sparite, whichever is dominant. Terms can be combined if two types of grain dominate, as in biopelsparite or bio-oosparite. Terms can be modified to give an indication of coarse grain size, as in biosparrudite or intramicrudite. Other categories of Folk are *biolithite*, referring to a limestone formed *in situ*, such as a stromatolite or reef-rock; and *dismicrite*, referring to a micrite with cavities (usually spar-filled), such as a birdseye limestone (Section 4.6.3).

3 The classification of R.J. Dunham (Fig. 4.20) divides limestones on the basis of texture into: *grainstone*, grains without matrix (such as a bio- or

oosparite); *packstone*, grains in contact, with matrix (this could be a biomicrite); *wackestone*, coarse grains floating in a matrix (could also be a biomicrite); and a *mudstone*, micrite with few grains. Additional terms of A.F. Embry & J.E. Klovan give an indication of coarse grain size (*floatstone* and *rudstone*), and of the type of organic binding in *boundstone* during deposition (*bafflestone*, *bindstone* and *framestone*). The terms can be qualified to give information on composition, e.g. oolitic grainstone, peloidal mudstone or crinoidal rudstone.

As a result of diagenetic modifications to limestones, care must be exercised in naming the rock. For example, a homogeneous-looking micrite may be a peloidal mudstone, and micrite in a bioclastic, grain-supported rock could be cement, compacted pellets (i.e. grains), primary sediment (i.e. matrix) or internal sediment (infiltrated geopetal sediment). The second

Principal grains in limestone	Limestone types			
	Cemented by sparite		With a micrite matrix	
Skeletal grains (bioclasts)	Biosparite		Biomicrite	
Ooids	Oosparite		Oomicrite	
Peloids	Pelsparite		Pelmicrite	
Intraclasts	Intrasparite		Intramicrite	
Limestone formed in situ	Biolithite		Fenestral limestone-dismicrite	

Fig. 4.19 Classification of limestones based on composition.

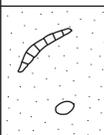
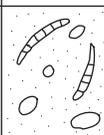
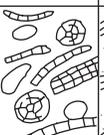
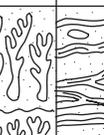
Original components not bound together during deposition				Original components bound together	Depositional texture not recognizable	Original components not organically bound during deposition		Original components organically bound during deposition			
Contains lime mud		Grain-supported	Lacks mud and is grain supported			>10% grains >2mm	Matrix supported	Supported by > 2mm components	Organisms act as baffles	Organisms encrust and bind	Organisms build a rigid framework
Less than 10% grains	More than 10% grains										
Mudstone	Wackestone	Packstone	Grainstone			Boundstone	Crystalline	Floatstone	Rudstone	Baffle stone	Bindstone
											

Fig. 4.20 Classification of limestones based on depositional texture.

example is basically a question of separating a grainstone from a packstone/wackestone. Many packstones are actually compacted wackestones.

The depositional environments and facies of limestones are considered in Section 4.10, where the typical limestone rock types of each environment are discussed briefly. The composition of limestones can be taken further by point-counting the different grain types present and calculating their percentages. A visual estimation of the grain percentages can be made using Fig. 2.49 in Chapter 2. Triangular diagrams can illustrate the composition for three principal components (e.g. ooids, bioclasts and peloids, or crinoids, brachiopods and bivalves). In this way different *microfacies* can be recognized.

4.5 Limestone grain size and texture

For the most part carbonate sediments are formed *in situ*. Although some may be transported from shelf to basin by turbidity currents or slumps, and from an inner to outer shelf by storms, the majority of limestones accumulated where the component grains were formed or have been subjected to only limited transport by wave and tidal currents. The skeletal grains of carbonate sediments vary greatly in size and shape. Interpretations of limestone deposition are thus to a large extent based on the types of grain present because these will often provide concise information on the depth, salinity, degree of agitation, etc. This is not to say that the grain size and degree of sorting and rounding are unimportant. Although the grain size will be a reflection largely of the size of the carbonate skeletons of the organisms living in the area and of the many biological factors involved in their breakdown, the physical factors of waves and currents will also contribute, and in cases dominate. A measure of the grain size, then, will often give useful additional information, reflecting the energy level of the environment, or energy gradient of the area. Where one is dealing with grainstones, the grain-size parameters discussed in Section 2.2.1 can be applied. It must be borne in mind, however, that carbonate particles are hydrodynamically different from quartz grains (see Kench & McClean, 1996). Apart from complications arising out of shape, carbonate grains commonly have a lower density because of pores and contained organic matter. The degree of sorting and rounding of skeletal grains (use Figs 2.4 and 2.6) can be useful in certain bioclastic

rocks, such as those of shelves and ramps where changes in these features could indicate proximity to a shoreline or zone of higher wave and tidal current activity. Some limestones of course, such as oolitic and peloidal grainstones, are very well sorted and rounded anyway. In interpreting the energy level of a depositional environment from a rock's grain-size parameters and texture, one is assuming that the sediment surface was in equilibrium with the hydrodynamic regime. With carbonate sediments this may not have been the case. In the modern shallow-marine environment, the sediment surface is commonly covered in a surficial microbial mat that stabilizes the sediment, enabling it to withstand current velocities up to five times those eroding nearby sediments lacking a microbial cover. During diagenesis, evidence of the mat could be destroyed. The probable wackestone would be interpreted as a low-energy deposit, whereas in fact periodically it was subjected to high current velocities.

Measurements of carbonate grain size in ancient limestones are made by point-counting slides or acetate peels under the microscope. For loose carbonate sands, the use of a settling chamber/sedimentation balance is recommended because this gives a better indication of the hydraulic behaviour than the grain-size distribution obtained by sieving.

In a general way the amount of micrite or lime mud in a limestone reflects the degree of agitation; lime muds tend to be deposited in quiet lagoons or on outer ramps, as well as on tidal flats and in the deep sea, in basins and periplatform areas. Increasing agitation leads to a decrease in the micrite content and increase in grain-support fabric and sparite content; sorting and rounding of grains then improves in the grainstone/biosparite. Interpretations must be made with care though because lime muds can accumulate in higher-energy environments, trapped and stabilized by sea-grass or a surficial microbial mat, which leaves no record in the sediment, and micrite can be precipitated as a cement during early diagenesis (Section 4.7.1).

4.6 Sedimentary structures of limestones

Limestones contain many of the sedimentary structures occurring in sandstones described in Section 2.3, but some structures are found only in carbonate sediments. Demicco & Hardie (1995) presented a description of sedimentary structures in limestones.

4.6.1 Bedding planes, hardgrounds, tepees and palaeokarstic surfaces

As in siliciclastic sediments, *bedding planes* generally represent a change in the conditions of sedimentation. The changes may have been subtle or short-lived. The bedding planes are mostly the result of changes in sediment grain size or composition. Thin clay seams also commonly define the bedding in limestone successions. However, with limestones it is not uncommon to find that bedding planes have been affected by dissolution as a result of overburden pressure (Section 4.7.5). Through this, originally gradational bed boundaries, such as a limestone passing up into a mudrock, or a grainstone into a lime mudstone, may become sharp. In many platform limestones, the ‘bedding’ planes are not primary depositional surfaces, but they have been produced by pressure dissolution during burial. This is most obvious where, for example, a ‘bedding’ surface occurs *within* a graded bed, or where it cross-cuts, at a low angle, a clear primary bedding surface. These *pseudobedding planes* actually account for much of the stratification in shallow and deep-water limestones (Simpson, 1985).

One particular type of bedding plane is a *hardground surface*. Hardgrounds are horizons of symsedimentary cementation, taking place at or just below the sediment surface. Where a hardground surface formed the sea floor, it was commonly encrusted by sessile benthic organisms, such as corals, serpulids, oysters, foraminifers and crinoids, and bored by polychaete annelids, certain bivalves and sponges. Hardground surfaces may cut across fossils and sedimentary structures. Two types of hardground surface can be recognized: a smooth, planar surface, formed by abrasion (Fig. 4.21), and an irregular, angular surface formed by dissolution (a corrosional hardground surface). The first type is more common in shallow subtidal sediments where waves and currents are able to move oolitic and skeletal sands across lithified sediment to produce a planar erosional surface. Corrosional hardground surfaces are more common in pelagic limestones where periods of non-sedimentation allow sea-floor cementation and dissolution. The identification of a hardground is important because it demonstrates symsedimentary submarine cementation. Hardground surfaces may become mineralized and impregnated with iron hydroxides, Fe–Mn oxides, phosphate and glauconite. Hard-

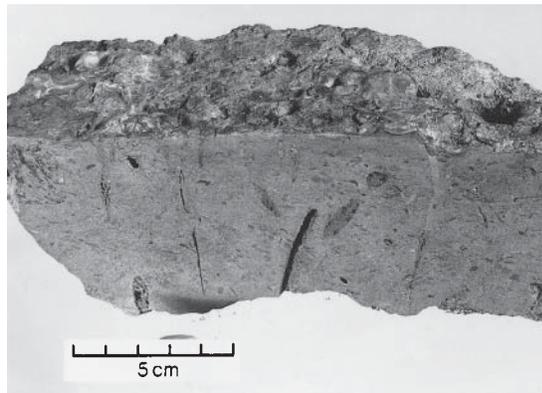


Fig. 4.21 Planar hardground surface encrusted with *Ostrea* (oysters) and penetrated by annelid and bivalve borings (seen on cut face). These features demonstrate symsedimentary sea-floor cementation of the sediment (see also Plate 11d). Jurassic. Gloucestershire, England.

grounds develop from loose sediments through firmgrounds to lithified layers, and associated with this there may be a change in the fauna, particularly of the burrowing organisms, as the sedimentation rate slowed down. Hardgrounds occur throughout the Phanerozoic and modern ones are forming at the present time off Qatar in the Arabian Gulf and on Eleuthera Bank, Bahamas (also Section 4.7.1).

One distinctive feature of peritidal limestones is the *tepee structure* (Fig. 4.22). Tepees are disruptions of the bedding into ‘pseudoanticlines’ and in plan view the tepee crests form a polygonal pattern. Tepees occur on the scale of tens of centimetres to several metres across. They mostly form on intertidal–supratidal flats as a result of the cementation and expansion of the surface-sediment layer. Upward movement (resurgence) of ground water, marine or meteoric, is a contributory factor in some cases. Elongate cavities (sheet cracks) commonly form beneath the uplifted slabs and in these, pisoids may form, as well as vadose cements such as dripstone and flowstone. These tepees usually are associated with planar stromatolites, desiccation cracks and intraclast conglomerates. Modern examples of these tepees, some with the spelean–pisoid association, occur on supratidal flats and around saline lakes in South and Western Australia. Tepees also form in the submarine environment, for example, where hardground surfaces have expanded through the cement precipitation. These are well developed off the

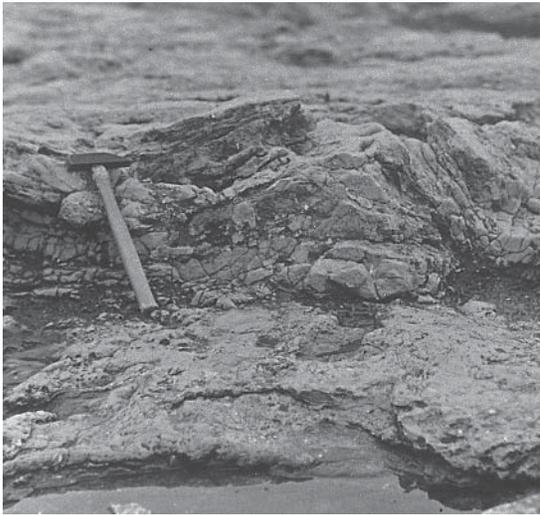


Fig. 4.22 Tepee structure in Triassic lacustrine littoral dolomite. Glamorgan, Wales.

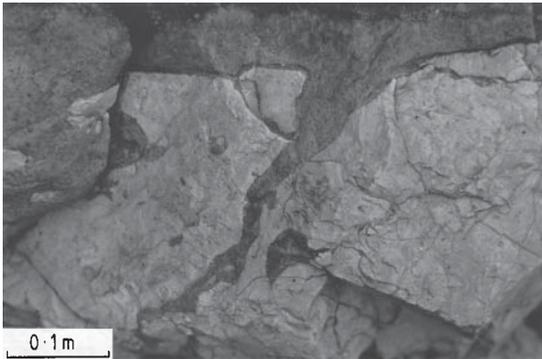


Fig. 4.23 Palaeokarst. Cretaceous limestone (white) is penetrated by Eocene sands (grey), which also contain blocks of the host rock. Sub-Alpine Chains, France.

Qatar Peninsula. Tepees are prominent in many ancient peritidal sequences; classic examples occur in the back-reef facies of the Permian Capitan reef complex of Texas–New Mexico and in the Triassic of the Dolomites of northern Italy. See the review of Kendall & Warren (1987) for more information and M. Mutti (1994).

Another particular type of bedding discontinuity peculiar to limestones is a *palaeokarstic surface* (Fig. 4.23). When carbonate sediments become emergent, then dissolution through contact with meteoric water

produces an irregular, pot-holed surface. This dissolution commonly takes place beneath a thin soil cover, and the soil itself may be preserved as a discontinuous clay seam or bed immediately above the dissolution surface. The term *karst* is applied to these dissolution features, which are typical of more humid climatic areas. In a well-developed karst system, pot holes and caverns may form many tens or even hundreds of metres below the surface. Breccias form by cave collapse and deposition from subterranean streams; speleothems and flowstones are precipitated too. Important karst developed in the Lower Ordovician of Texas and Oklahoma, and in central China (Ordos Basin) and are major hydrocarbon reservoirs. Examples are given in Esteban & Klappa (1983) and James & Choquette (1988); also see Vanstone (1998), Molina *et al.* (1999) and Purdy & Waltham (1999).

Laminated crusts form upon and within uplifted carbonate sediments, as a type of calcrete or caliche (Section 4.10.1). In most cases, they are calcified root mats (Wright *et al.*, 1988), and usually they are closely associated with vadose pisoids and black pebbles. In ancient limestones, laminated crusts can be mistaken for stromatolites, but their association with palaeokarstic surfaces and palaeosoils indicates a sub-aerial, pedogenic origin.

Bedding surfaces and their significance in carbonates are discussed in Hillgärtner (1998).

4.6.2 Current and wave structures

All the current structures of siliciclastic rocks occur in limestones: wave and current ripples, cross-lamination, cross-bedding on all scales, planar or flat bedding, small scours to large channels, HCS in storm beds, bundled cross-beds and reactivation surfaces in tidal sands and sole structures on the bases of storm beds and turbidites. Post-depositional structures resulting from dewatering and loading are also common. For details of these sedimentary structures see Section 2.3. The lack of clay in many limestones, together with the effects of surface weathering, may make internal structures difficult to discern in the field. Careful observations, perhaps aided by polishing, etching or staining cut blocks, will often bring them to light.

The same importance is attached to current structures in limestones as in sandstones. They are essential to environmental interpretation and facies analysis,

giving valuable information on depositional process, palaeocurrents, depth and water turbulence. Intra-clasts of lime mud (flakes) and large fragments of skeletal debris are generally more common in limestones. They may be concentrated through current winnowing to form lag deposits, or transported by storms to give rudstones and floatstones (flakestones, Fig. 4.5). These beds may show imbrication of clasts, reverse or normal grading. Elongate fossils are commonly aligned parallel to, or normal to, the current direction, so that they too can give a palaeocurrent indication.

4.6.3 Cavity structures

Depositional and early diagenetic cavity structures are common in limestones, and there are many types. Some are partly filled with sediment that has been washed in to occupy the lower part of the cavity, with the space above occupied by sparite cement (see Plates 9a,b & 11a,c,d). Such cavity fills are known as *geopetal structures*, and they are a most useful way-up indicator (sparite at the top of course). Geopetal structures also record the horizontal at the time of sedimentation (acting as a spirit level) and in some cases show that there was an original depositional dip (as in fore-reef limestones, for example). *Umbrella structures* are simple cavities beneath convex-up bivalve and brachiopod shells and other skeletal fragments. *Intra-skeletal cavities* occur in enclosed or chambered fossils, such as gastropods, foraminifers and ammonoids. *Growth cavities* are formed beneath the skeletons of frame-building organisms, corals and stromatoporoids, for example, where they build out above the sediment or enclose space within the skeletons.

Fenestral cavities or ‘birdseyes’ are small cavities that occur particularly in peloidal mudstones of intertidal–supratidal environments. The majority are spar-filled only, but some may be sediment-filled. Three main types can be distinguished:

- 1 irregular fenestrae, the typical ‘birdseyes’ (Fig. 4.24), several millimetres across, equidimensional to irregular in shape;
- 2 laminoid fenestrae, several millimetres high and several centimetres long, parallel to bedding (Figs 4.12 & 4.25);
- 3 tubular fenestrae, cylindrical, vertical to subvertical in arrangement, several millimetres in diameter.

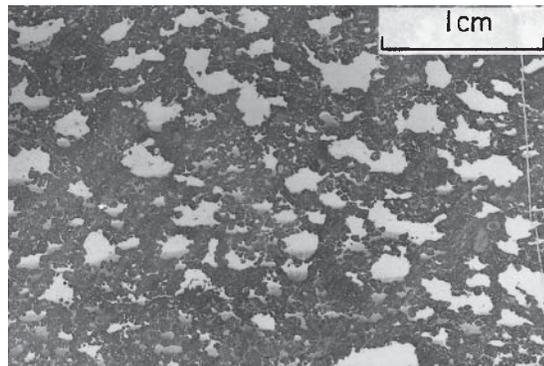


Fig. 4.24 Birdseyes (fenestrae) in lime mudstone. Fenestrae are filled with sparry calcite although some contain internal sediment. Carboniferous. Clwyd, Wales.

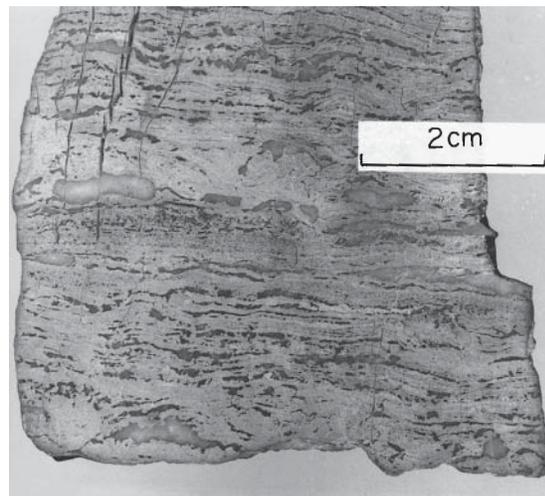


Fig. 4.25 Laminoid fenestrae and birdseyes filled with calcite in microbial and peloidal dolomite. The dolomite is probably penecontemporaneous in origin. Triassic. Austrian Alps.

Irregular fenestrae in abundance form the so-called birdseye limestones (my favourite). They are ascribed to gas entrapment in the sediment and desiccation and so are a characteristic intertidal-facies indicator. Similar structures also occur in subtidal grainstones associated with early cementation and hardgrounds. Beach sands also contain irregular fenestrae, called *keystone vugs*, from the movement of water and air through the sediment.

Laminoid fenestrae develop in laminated sedi-

ments, particularly planar stromatolites, from the decay of organic matter, and desiccation and parting of laminae. *Tubular fenestrae* are formed mainly by burrowing organisms, but plant rootlets produce similar tubes. Although common in tidal-flat sediments, tubular fenestrae will also occur in shallow, subtidal sediments. Fenestral limestones occur throughout the geological record; one notable example is the *loferite* of the Alpine Triassic (Fig. 4.25).

Small vugs to extensive cavern systems on the scale of tens of metres form as a result of surface and near-surface karstic dissolution of limestones (see Section 4.6.1). Dissolution also may take place during deep burial in the formation of hydrothermal karst.

A further particular type of cavity is *stromatactis* (Fig. 4.26). This is common in carbonate mud-mounds of the Palaeozoic such as the Waulsortian ‘reefs’ and other mounds of the European Carboniferous and Devonian, the Devonian Tully Limestone of New York, Silurian sponge ‘reefs’ of Quebec and Ordovician mud-mounds of Nevada. Stromatactis cavities have an irregular, unsupported roof and a flat floor, formed by internal sediment. The cement is invariably a first generation of fibrous calcite, followed by drusy sparite. Most stromatactis cavities are a few centimetres long, but they may reach tens of centimetres. The origin of the cavities has led to much discussion and speculation. Stromatactis in Upper Devonian mud-mounds has been interpreted as the product of recrystallization of algal, cyanobacterial and bacterial colonies. Fibrous calcite, which is a feature of most stromatactis cavities and normally is thought of as a



Fig. 4.26 Stromatactis cavities filled with fibrous calcite and drusy calcite. Mud-mound, Upper Devonian. Ardennes, Belgium.

marine cement (see Section 4.7.1), also has been interpreted as microbial in origin. Various inorganic origins are available: sediment collapse and dewatering; winnowing of uncemented sediment beneath lithified crusts or beneath gelatinous microbial mats; and dissolution during deep burial. A popular mechanism now is that stromatactis formed within a sponge framework, from the decay of uncemented sponge tissue (Bourque & Boulvain, 1993; and papers in Monty *et al.*, 1995). It is likely that stromatactis structures can form through several of these processes.

Two types of cavity that form in partly lithified or cemented limestone are sheet cracks and neptunian dykes. *Sheet cracks* are cavities generally running parallel to the bedding, which have planar walls, although some may have irregular roofs. *Neptunian dykes* cut across the bedding and may penetrate down many metres from a particular bedding plane. Both sheet cracks and neptunian dykes are filled with internal sediment, in some cases with fossils a little younger than the adjacent limestone, if the cavities opened on to the sea floor. Spectacular examples of both types, in pelagic limestones and penetrating down into underlying platform carbonates, occur in the Triassic and Jurassic of the Alps and Spain (e.g. Molina *et al.*, 1995). Sheet cracks and neptunian dykes form through small tectonic movements during sedimentation and/or some slight downslope movement of sediment, causing fracturing of the lithified or partly lithified limestone mass.

Fractures and veins filled with calcite, of various types, are common in limestone. Although they can be early structures, like neptunian dykes, the majority form through later fracturing, and this is usually in response to tectonic stresses rather than purely diagenetic processes.

4.7 Carbonate diagenesis

The diagenesis of carbonates involves many different processes and takes place in near-surface marine and meteoric environments, down into the deep-burial environment. It is most important in occluding and generating porosity in the sediment. Six major processes can be distinguished: cementation, microbial micritization, neomorphism, dissolution, compaction and dolomitization. Carbonate diagenesis mostly involves the carbonate minerals, aragonite, calcite and dolomite, but other minerals such as quartz, feldspar, clays,

phosphates, iron oxides and sulphides, and evaporites also may be involved. The diagenesis of carbonate sediments begins on the sea floor; in fact depositional and diagenetic processes may be going on at the same time. As a reef is growing or carbonate sand is being moved by the waves, cements may be precipitated within intraskeletal cavities and grains altered by *micritization*. The latter process has been described in Section 4.3.3, and results in the formation of micrite envelopes around bioclasts and completely micritized grains. Micrite envelopes play an important role during diagenesis by maintaining the shape of an aragonite bioclastic grain after its dissolution.

Cementation is the major diagenetic process producing a solid limestone from a loose sediment and taking place principally where there is a significant throughput of pore-fluid saturated with respect to the cement phase. The mineralogy of the cements depends on water chemistry, particularly P_{CO_2} and the Mg/Ca ratio, and carbonate supply rate (see Section 4.7.1). *Neomorphism* is used to describe replacement and recrystallization processes where there may have been a change of mineralogy. Examples include the coarsening of crystal sizes in a lime mud/micrite (aggrading neomorphism) and the replacement of aragonite shells and cements by calcite (*calcitization*). Many limestones have suffered *dissolution* as a result of the passage of pore-fluids undersaturated with respect to the carbonate phase present. This is a major process in near-surface, meteoric diagenetic environments, and may lead to the formation of karst (see Section 4.6.1), but it can also take place on the sea-floor and during deep burial. The secondary porosity created by carbonate dissolution is important in some hydrocarbon reservoirs. *Compaction* takes place during burial, resulting in a closer packing of grains, their fracture and eventual dissolution where in contact. Chemical compaction leads to stylolites and dissolution seams, when burial exceeds many hundreds of metres of overburden. *Dolomitization* is a major alteration process for many limestones and the dolomite, $CaMg(CO_3)_2$, may be precipitated in near-surface and burial environments. There are a number of models for dolomitization, but the matter is still one of great debate.

Three major diagenetic environments are distinguished: marine, near-surface meteoric and burial (see Fig. 4.27). In the marine environment, diagenesis takes place on and just below the sea-floor in both shallow and deep water, and in the intertidal–supratidal

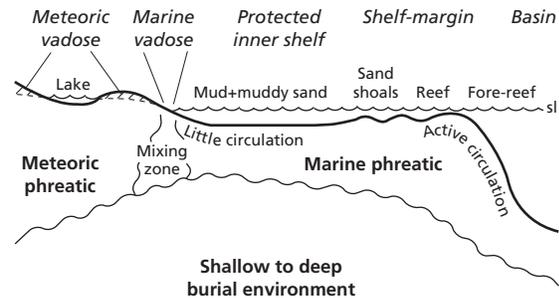


Fig. 4.27 Carbonate diagenetic environments, schematically drawn for a rimmed shelf with unconfined aquifers. Where there are confined aquifers as a result of impermeable strata, then it is possible, for example, for meteoric waters to penetrate deep beneath the marine shelf and even emerge on the sea floor as submarine springs.

zone. Meteoric diagenesis can affect a sediment soon after it is deposited if there is shoreline progradation or a slight sea-level fall, or it may operate much later when a limestone is uplifted after burial. The burial environment, the least well known, begins at a depth below the sediment surface of tens to hundreds of metres, that is, below the zone affected by surface processes, down to several kilometres where metamorphic dehydration reactions and wholesale recrystallization take over.

4.7.1 Marine diagenesis

Marine diagenesis in Recent carbonate sediments

Intertidal–supratidal diagenesis. Cementation in the intertidal zone produces cemented beach sands known as beachrock. Beachrocks are most common in the tropics and subtropics but they do occur along temperate shorelines. They are composed of the same sediment that forms the surrounding loose beach sand; this usually is calcareous but it may have a substantial or even dominant siliciclastic component. Beachrocks can form quickly, as is evidenced by the inclusion of anthropogenic objects such as beer cans. Beachrock formation probably takes place a few tens of centimetres below the surface of the beach, but in many places it is exposed through storm action, and then it can be eroded (to produce intraclasts), encrusted, and bored and grazed by intertidal organisms.

The cements in modern beachrocks are aragonite and/or high-Mg calcite. Aragonite typically occurs as fringes from 10 to 200 μm thick of acicular crystals, orientated normal to the grain surfaces (Figs 4.28 & 4.29). In many cases the cement fringes are isopachous, i.e. of equal thickness (Fig. 4.30), indicating marine phreatic (below the water table) precipitation

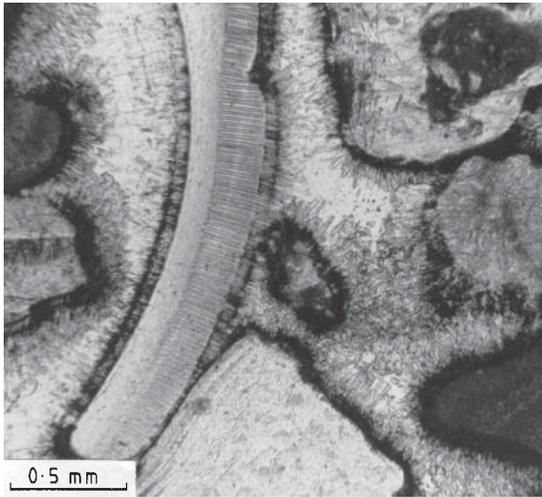


Fig. 4.28 Beachrock with cements of acicular aragonite and micritic high-Mg calcite: the latter is the dark coating around grains. The aragonite cement has grown syntaxially on the bivalve grain, which shows a two-layer shell structure. Plane polarized light. Great Barrier Reef, Australia.

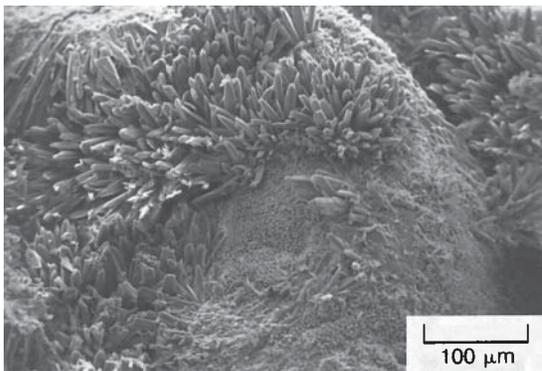


Fig. 4.29 Scanning electron micrograph of beachrock cement shown in Fig. 4.28. Where the coating of acicular aragonite crystals has come off the grain, the high-Mg calcite micrite cement is visible.

where pores were constantly water-filled. Asymmetric cement fringes, thicker on the underside of grains, and meniscus cements, concentrated at grain contacts (Fig. 4.30), are recorded from some beachrocks and indicate precipitation in the marine vadose zone. High-Mg calcite is usually a dark micritic cement coating grains or filling pores. Micritization of grains is common in beachrocks and calcified microbial filaments are also present. If the meteoric ground-water table is high in the backshore area, then low-Mg calcite cements may be precipitated in beachrocks there in the upper intertidal zone.

Two processes are important in beachrock formation:

- 1 purely physico-chemical precipitation through evaporation of seawater when the tide is out and CO_2 -degassing of seawater as it is pumped through the sand by waves and the rising and falling tide;
- 2 biochemical–microbial precipitation, involving microbial photosynthesis, bacterial calcification and the decomposition of organic matter.

Cemented surface crusts occur in the high intertidal to low supratidal zone of tidal flats in carbonate areas. Along the Trucial Coast of the Arabian Gulf, aragonite-cemented crusts are commonly brecciated or polygonally cracked, and expansion of the crusts as a result of the cementation leads to pseudoanticlines or *tepee structures* (Section 4.6.1; Kendall & Warren, 1987). The resurgence of continental ground water onto supratidal flats may contribute to crust and tepee formation. Beneath the surface crust, dripstone cements, vadose pisoids and aragonite botryoids

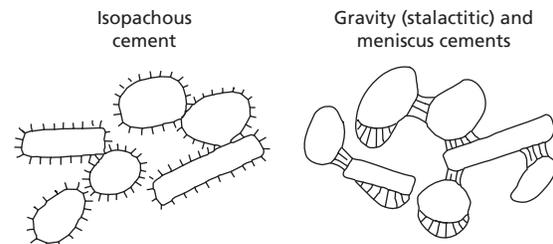


Fig. 4.30 The geometry of first-generation cements: isopachous cement, indicative of precipitation in phreatic zones where all pores are filled with water (typical feature of low-intertidal and subtidal cements); and gravity (stalactitic) and meniscus cements, indicative of vadose-zone precipitation, as occurs in high-intertidal, supratidal and shallow-subsurface meteoric situations.

may develop. Supratidal crusts in the Bahamas are cemented by dolomite (Section 4.8.1, see Plate 10d).

Shallow-subtidal cementation. Sea-floor diagenesis in low-latitude, shallow-marine areas mostly involves cementation and microbial micritization. The former is more common in high-energy areas, where seawater is pumped through the sediment, whereas the latter is more common in quieter-water areas, such as back-reef lagoons. Thus an active marine phreatic diagenetic environment can be distinguished from a stagnant one. In higher-latitude, shallow-water areas, seawater is undersaturated with respect to CaCO_3 and so inorganic precipitation of cement does not take place. Dissolution of grains may occur, however.

In the stagnant, marine phreatic diagenetic environment, cementation is restricted to intraskeletal cavities, such as occur within gastropod and foraminiferal bioclasts. The formation of grapestones and aggregates is common in this environment, with microbial binding and filament calcification usually involved. In these areas too, micritization of grains by recrystallization is taking place on the sea floor (Reid & Macintyre, 1998).

Shallow-subtidal cementation of loose carbonate sand to produce surface crusts and lithified layers is rare but it is taking place in a few metres of water off the Qatar Peninsula, Arabian Gulf and on Eleuthera Bank, Bahamas. Off Qatar, these modern hardgrounds are being bored and encrusted by organisms, and polygonal cracks and tepee structures have formed through expansion of the cemented layer. The cements are mainly acicular aragonite, with some micritic high-Mg calcite. The acicular cements form isopachous fringes (see Plate 11b), but if well developed and completely filling the pore space, then a polygonal pattern is observed in thin-section, from the meeting of the cement fringes on adjacent grains. The precipitation of cements is probably the result of turbulent bottom conditions and the pumping of CaCO_3 -supersaturated seawater through the sediments, particularly in areas of slow sedimentation. Microbial filaments, commonly calcified, are usually present, and they may have provided some initial stabilization of the sediment, permitting the cementation to begin.

Marine cements can be developed extensively in reefs (see papers in Schroeder & Purser, 1986). There is a wide variety of cement morphologies but the miner-

alogy is either aragonite or high-Mg calcite. Aragonite occurs mostly as acicular fringes (like those in Fig. 4.28), and needle meshworks, but one prominent type is the botryoid: isolated or coalesced mamelons, which reach 100 mm in diameter (see Plate 11a). They consist of fanning fibrous crystals, commonly twinned to give a pseudo-hexagonal cross-section. High-Mg calcite occurs as bladed cements, 20–100 μm long and less than 10 μm wide, forming isopachous fringes (see Plate 11b). They are more common than aragonite in some reefs, but rare in others. High-Mg calcite also occurs as a micritic cement. This coats grains and lines cavities, but more commonly forms peloidal structures, abundant in skeletal cavities and forming surficial crusts on corals in some instances. The peloids are 20–60 μm in diameter and are arranged in grainstone to packstone textures. The origin of the peloids has given rise to much discussion, principally over whether they are inorganic precipitates, microbial precipitates, faecal pellets, or detrital sediment. The consensus now is for a microbial origin with precipitation taking place within and around clumps of bacteria (see Chafetz, 1986).

The precipitation of cements in reefs contributes towards the generation of a solid framework, but reefs are also subject to extensive *bioerosion*. Microbial organisms, clionid sponges and lithophagid bivalves all bore into carbonate skeletons in reefs; sponges also generate much fine debris. Fish and other organisms graze on corals and generate sediment too. Internal sedimentation of this material into primary and secondary cavities is widespread, and these sediments may then become cemented.

The distribution of cements in many reefs appears to relate to the circulation of water; cementation is generally more intense along the windward margins, where seawater is constantly pumped through the reef. However, on a smaller scale, the occurrence of cements may be very patchy and varied, both in terms of extent, mineralogy and morphology. One cavity may contain acicular aragonite, an adjacent cavity high-Mg calcite peloids, whereas another close by may be empty (as in Plate 11a). One important control on the amount of cementation is local permeability, which determines the fluid-flow rates through the reef-rock (Goldsmith & King, 1987). This also may control the mineralogy itself.

Aragonite cements of intertidal to shallow-subtidal sediments usually have a high strontium content, up to

10 000 p.p.m., and Mg content of about 1000 p.p.m. or less. The high-Mg calcite cements are typically between 14 and 19 mol.% MgCO_3 , but Sr is low at around 1000 p.p.m.

Deeper-water cementation. Cemented carbonate sediments have been recovered from the ocean floors at depths down to 3500 m, mostly from areas of negligible sedimentation, such as seamounts, banks and plateaux. The limestones consist chiefly of planktonic foraminifera, molluscs and coccoliths cemented by a micritic calcite. The limestones are commonly bored and may be impregnated with phosphate and ferromanganese oxides. On the deeper-water slopes around the Bahama Platform (700–2000 m deep), periplatform ooze is being cemented by micritic low-Mg calcite. The sediment itself consists of planktonic foraminifera, coccoliths and shallow-water material derived from the platform. Cemented crusts and hardgrounds pass downslope into patchily cemented sediment and nodules, and then there is little cementation on the deeper-water slopes. Originally aragonitic bioclasts have been leached and high-Mg calcite grains have lost their Mg (Dix & Mullins, 1988). In the Mediterranean and Red Sea, where there are and/or have been warmer bottom waters, thin surface crusts and nodules of lithified pelagic sediment are cemented by micritic high-Mg calcite. Acicular and micritic aragonite cements pteropod layers on the floor of the Red Sea.

Carbonate cement precipitation in these deeper-water environments is mainly a reflection of the very slow sedimentation rates, which allow interaction between sediment and seawater. The CaCO_3 is derived from seawater and the dissolution of less stable grains in the sediment; the type of cement precipitated and its Mg content are determined by the precipitation rate and the water temperature; as the latter goes down, so does the mol.% MgCO_3 .

Marine dissolution. Dissolution of carbonate grains does take place on the shallow sea floor in higher latitudes where seawater is undersaturated with respect to the carbonate phase. The saturation state of seawater with regard to CaCO_3 decreases with lower temperature (and so broadly with increasing depth), and seawater becomes undersaturated in respect of aragonite and high-Mg calcite before low-Mg calcite. Dissolution of bioclasts is taking place in slope sedi-

ments off the Bahama Platform at depths of a few hundred metres, and in the geological record relatively shallow sea-floor dissolution of aragonite is recorded from hardgrounds in the Jurassic and Ordovician (Palmer *et al.*, 1988). These were times when shallow, low-latitude seawater was supersaturated with regard to calcite, but not aragonite, too, as is the case today. Sea-floor dissolution of aragonite is common in Mesozoic deeper-water pelagic limestones; this is clearly shown where, through sea-floor loss of their aragonitic shell, ammonites are preserved as casts encrusted by sessile organisms and Fe–Mn oxides. Carbonate dissolution increases with increasing depth until the CCD (carbonate compensation depth), below which limestones are not deposited (see Section 4.10.7).

Marine diagenesis in ancient limestones

In many limestones, there is abundant evidence for diagenesis taking place on the sea-floor or just below. Marine cements are common in ancient reef-rocks, and hardgrounds are well known too in the geological record (see Section 4.6.1). However, the cements themselves are variable; some were similar to those of the Recent and were composed of aragonite and high-Mg calcite, although they are now calcite (low Mg). On the other hand, other marine cements in ancient limestones do not appear to have exact modern equivalents and were precipitated as calcite with fabrics somewhat different from modern cements.

General features for the recognition of marine cements in limestones are:

- 1 they are the first-generation cement;
- 2 they usually form isopachous fringes around grains or cavity walls;
- 3 they usually are/were fibrous in nature;
- 4 they may be cut by borings or include skeletal debris;
- 5 they may be associated closely with internal sediments;
- 6 the crystals are usually non-ferroan and non-luminescent;
- 7 they are succeeded by clear calcite spar.

Ancient marine cements are shown in Plates 11c,d & 12a,b.

Ancient marine aragonite cements. Aragonite cements are very rarely preserved as aragonite in lime-

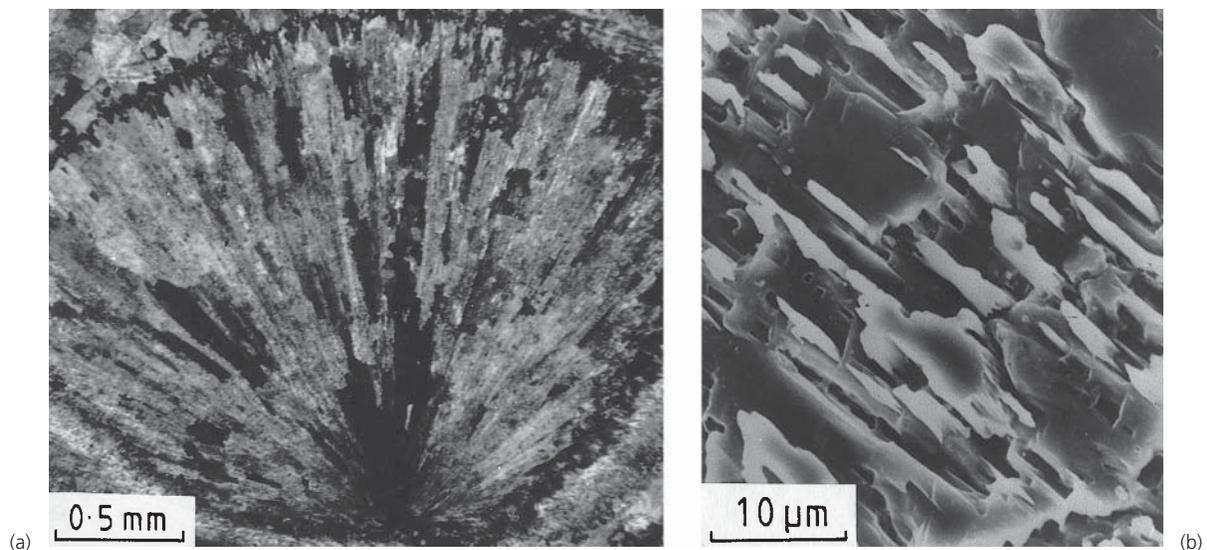
stones because the mineral is metastable. They are altered to calcite in a similar way to aragonite bioclasts, either through wholesale dissolution and then later filling of the void by calcite spar, or through calcitization, whereby calcite replaces the aragonite across a thin-fluid film, with dissolution of aragonite on one side and precipitation of calcite on the other. In this process, some traces of the original cement texture may be preserved within the replacement calcite crystals by the pattern of minute relics of the aragonite or inclusions of organic matter (Fig. 4.31). The replacement calcite crystals, a type of neomorphic spar (see Section 4.7.4), are usually irregular to equant in shape, cross-cutting the original acicular pattern of the aragonite cement. This calcite may have a relatively high (several thousand p.p.m.) strontium content inherited from the aragonite. Some ancient calcitized aragonite cements have distinctive 'square-ended terminations' along the cement fringes. The gross morphology of ancient aragonite cements is similar to that of Recent cases, isopachous fringes and botryoids. Although some marine isopachous cement fringes in limestones

were calcite originally, botryoids appear to have been aragonitic only; somewhat similar structures, in calcite, do form in meteoric, spelean-vadose diagenetic environments. Descriptions of ancient aragonite are included in Tucker & Hollingworth (1986), from the Permian of northeast England, and Roylance (1990), from the Carboniferous of the Paradox Basin, USA.

Ancient marine calcite cements. The most common marine calcite cement in ancient limestones is *fibrous calcite*: elongate crystals normal to substrate, and usually cloudy or dusty in appearance relative to later calcite spar. A columnar growth form, with length-to-width ratio of more than 6:1 and width greater than 10 μm , can be distinguished from a more acicular variety, with a much higher length/width ratio and width of around 10 μm . Columnar fibrous calcite is abundant in reef cavities and in stromatactis structures of Palaeozoic mud-mounds (see Fig. 4.26). The more acicular type is common in grainstones but occurs in reefs too (see Plate 11c,d).

Fibrous calcite has a range of fabrics from unit-extinguishing crystals to undulose-extinguishing radial fibrous calcite (RFC), with divergent optic axes, and fascicular-optic fibrous calcite (FOFC), with convergent optic axes (Fig. 4.32 and Plate 12a,b). In RFC (the most common type) the extinction swings across the crystal in the same direction as rotation of the

Fig. 4.31 Calcitized aragonite botryoid. (a) Photomicrograph of irregularly shaped neomorphic calcite cross-cutting the pattern of the original aragonite crystals. Crossed polars. (b) Scanning electron micrograph of minute aragonite crystal relics in calcite replacing an aragonite botryoid. Permian reef (Zechstein). Durham, England.



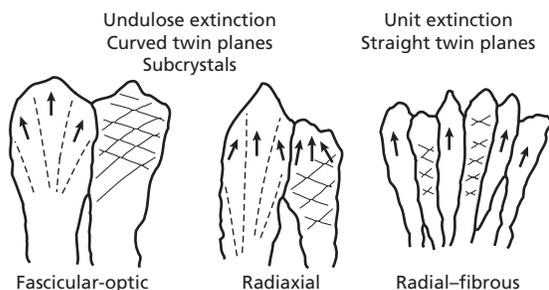


Fig. 4.32 Common types of fibrous calcite.

microscope stage. Twin planes are common in fibrous calcite and in RFC they are concave away from the substrate. There has been much discussion over the origin of fibrous calcite, with a replacement origin long popular, but it is now regarded generally as a primary precipitate, with the peculiar fabrics attributed to split-crystal growth (Kendall, 1985).

One particular point of interest with fibrous calcite is whether it was originally low- or high-Mg calcite. It is not easy to decide, as high-Mg calcite loses its Mg during diagenesis, but several lines of evidence can be used. Fibrous calcite usually is cloudy with inclusions, and some of these may be minute crystals of dolomite. Others will be fluid-filled inclusions or empty. *Microdolomites* are taken to indicate an original high-Mg calcite mineralogy; this is supported by their presence in formerly high-Mg calcite bioclasts such as echinoderms. In addition, fibrous calcite after high-Mg calcite may still contain several mol.% MgCO_3 after the diagenetic loss of most of the Mg. This *magnesium memory* can be detected only by geochemical means, such as with use of the electron probe. Fibrous calcites originally of high-Mg calcite may pick up some Fe^{2+} , if this is in the pore waters when the high- to low-Mg calcite transformation is taking place. Although the fabrics of fibrous calcite are mostly unchanged during diagenesis, some neomorphism may occur, especially if the crystals were originally high-Mg calcite. Dissolution may occur along twin boundaries, for example.

In addition to fibrous calcite, there is evidence in some limestones for equant sparry calcite being a marine precipitate. Although mostly of meteoric or burial origin, it does occur as the first-generation cement in some Jurassic and Ordovician hardgrounds. Syntaxial overgrowths on echinoderm debris, also generally re-

garded as near-surface meteoric or burial precipitates, formed early in some hardgrounds too. Early, inclusion-rich zones of crinoidal overgrowths are probably marine (see Plate 12c).

Similar textures to the micritic and peloidal high-Mg calcite precipitates in modern reefs do occur in ancient reefs, and there are the same arguments over their origin. The peloidal structures are particularly common in Triassic reefs and some Jurassic ones too (e.g. Sun & Wright, 1989).

Discussion of marine cements

The mineralogy of modern tropical shallow-marine cements is mostly either aragonite or high-Mg calcite, and in the geological record, marine cements originally of aragonite, high-Mg calcite and low-Mg calcite can be identified. In a similar manner to ooids (see Section 4.3.1), there appears to be a secular variation in marine cement mineralogy through the Phanerozoic (see Fig. 4.4). Aragonite botryoids, for example, one of the distinctive marine cement morphologies, are common in the Cenozoic and Permo-Triassic, but apparently absent from the mid-Palaeozoic and Jurassic-Cretaceous, where fibrous calcite is the dominant cement type. As with ooids, the controls on cement mineralogy are likely to be the seawater Mg/Ca ratio, P_{CO_2} and carbonate-supply rate (Fig. 4.33). Aragonite has a similar stability to magnesium calcite, with around 12 mol.% MgCO_3 . It is unclear what controls whether aragonite or high-Mg calcite is precipitated. The presence of Mg^{2+} and SO_4^{2-} in seawater does result in a kinetic hindrance to calcite precipitation, so favouring aragonite. High carbonate supply appears to favour aragonite precipitation too, so that where fluid-flow rates are higher, as in very permeable reef-rocks and lime sands, aragonite will be precipitated in preference to high-Mg calcite, which will tend to occur in less permeable sediments. In some situations, substrate control is another factor, with cement crystals being the same mineralogy as the substrate, and in optical continuity (syntaxial) too. See the review of Given & Wilkinson (1985) for more information.

4.7.2 Meteoric diagenesis

Near-surface meteoric diagenesis mostly involves fresh water and the major processes are carbonate dissolution, cementation and the formation of soils. The

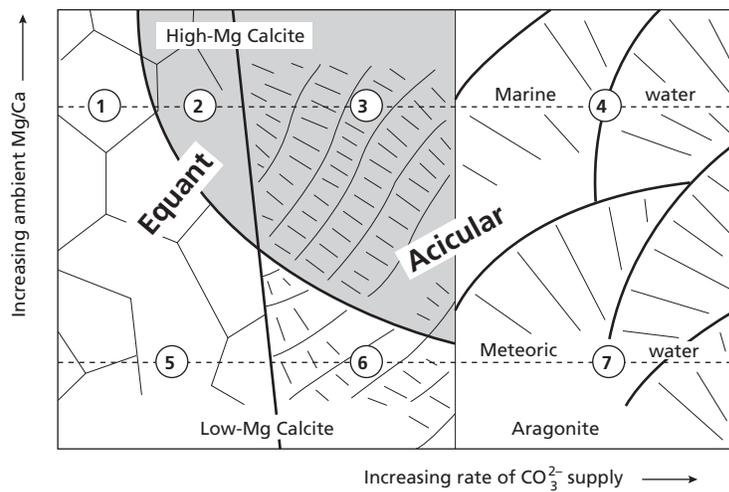


Fig. 4.33 Schematic illustration of relationship between fluid Mg/Ca ratio, rate of carbonate-ion supply, crystal morphology (equant or acicular) and mineralogy of inorganic precipitates (low-Mg calcite, high-Mg calcite or aragonite). The Mg/Ca ratio of the meteoric line is 0.3 and of the marine line 5.2. Low-Mg calcite is up to 9 mol. % MgCO₃ and high-Mg calcite is above. The position of modern, naturally occurring precipitates.

- 1 Equant calcite cement in cold, deeper-water, low-latitude sediments and shallow-water temperate sediments
- 2 Equant high-Mg calcite cements in reefs (rare)
- 3 Acicular high-Mg calcite cements in reefs and lime sands
- 4 Acicular aragonite cements in reefs and lime sands
- 5 Equant low-Mg calcite spar cements in meteoric environments
- 6 Acicular low-Mg calcite in speleothems and travertines
- 7 Acicular aragonite in speleothems (rare)

position of the ground-water table is important and the vadose zone above is distinguished from the phreatic zone below. In the vadose zone, pores periodically contain water, air or both, and an upper zone of infiltration is distinguished from a lower zone of percolation. Rainwater is undersaturated with respect to CaCO₃ so that dissolution is one of the main processes operating in the infiltration zone. As the water moves down through the vadose zone, it may become supersaturated with respect to CaCO₃ so that precipitation occurs. This is normally of calcite (low Mg), as meteoric water has a very low Mg/Ca ratio. In the phreatic zone, pores are fluid-filled all the time, and the normally fresh water gives way downwards to more saline water at depths of several hundred metres or more. In coastal regions, the phreatic meteoric ground-water realm passes into a mixing zone with seawater. It has been suggested frequently that the mixing zone is an important location for dolomitization (see Section 4.8.2), but there is now much doubt about this. Climate is a major influence on meteoric diagenesis because it controls the availability of meteoric water and

also affects the temperature and degree and nature of plant cover and soil development. Climate also determines the type and extent of karstic dissolution. Palaeokarstic surfaces (see Section 4.6.1) are now well known in the geological record and can be recognized from their morphology (an irregular pot-holed surface) and association with soil crusts and other features of subaerial exposure. Subsurface karst is the system of caves and fissures that may extend great distances down from the surface.

There have been several studies of modern to Pleistocene limestones to document the progressive alteration of marine sediments with increasing exposure to meteoric water. In the early stages, low-Mg calcite is precipitated on the surfaces of grains as an (a) isopachous (uniform thickness) fringe if precipitated in the phreatic zone, below the water table where all pores are completely filled with water, or (b) asymmetric fringe, thicker on the underside of grains (a dripstone effect) or located at grain contacts (a meniscus effect), if precipitated in the vadose zone (Fig. 4.30, and Plates 6a,b & 13a). Syntaxial overgrowths on echinoderm

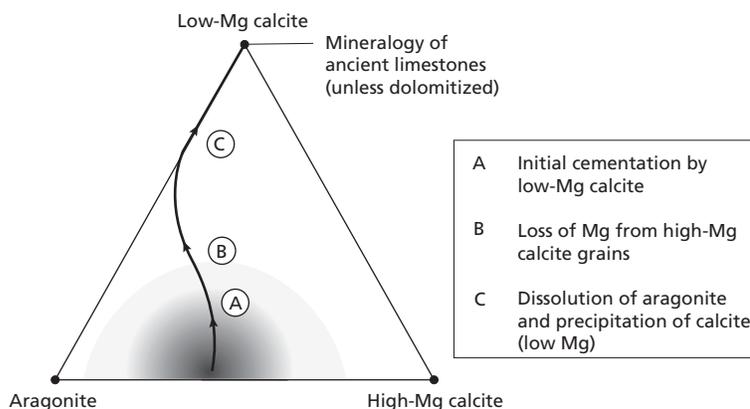


Fig. 4.34 Triangular diagram indicating the mineralogical range of modern marine carbonate sediments (stippled) and the path of meteoric diagenesis.

grains also begin to develop in this early stage. Next generally follows the loss of Mg from the high-Mg calcite, leaving a sediment of low-Mg calcite and aragonite (Fig. 4.34). The next stage is aragonite dissolution and reprecipitation of the CaCO_3 as drusy calcite spar. Aragonite skeletal grains are dissolved and the voids left are then filled with calcite, the shape of the grains being maintained by the earlier cement fringe or a micrite envelope (Section 4.3.3). The main feature of this dissolution–reprecipitation process is a loss of internal structure in the aragonitic skeletal grains. The reprecipitation of calcite may follow soon after aragonite dissolution, or, there may be a much longer time gap, allowing some compaction of the skeletal voids before calcite precipitation, producing broken and fractured cement fringes and micrite envelopes (seen in Plates 7c & 13d). Some aragonite grains are replaced *in situ*, i.e. calcitized (Section 4.7.4), so that the internal structure is preserved to some extent (see Plate 8a). The last stage of meteoric diagenesis would involve further precipitation of low-Mg calcite to fill all remaining voids. In most Pleistocene limestones, this final stage has rarely produced a fully cemented limestone; porosity may still reach 20%. A further phase of cementation is thus required to reduce the porosity to less than 5%, the typical value of ancient limestones.

Near-surface, meteoric vadose drusy calcite spar is generally non-ferroan, as pore waters are usually oxidizing. However, if there is decomposing organic matter, the water-flow rate is low and Fe^{2+} is available, then phreatic calcite spar may be ferroan.

Near-surface, meteoric calcite cementation takes place in many continental sediments, such as wadi

gravels, scree sediments and aeolian sands, where pore waters enriched in CaCO_3 are evaporated. Also in the meteoric diagenetic environment, soils are formed, especially calcretes (caliches). Near-surface sediments are cemented and altered through pedogenesis, and some distinctive fabrics are produced, such as needle-fibre calcite, alveolar texture, vadose pisoids and laminated crusts from calcification of root mats (see Section 4.10.1).

Ancient meteoric calcite cements

In the geological record, the most obvious products of meteoric diagenesis are the calcareous soils (calcretes) and palaeokarsts, but the vadose cement morphologies are not that uncommon (see Plate 13a) and it is possible that some drusy calcite spar, the ubiquitous cement of ancient limestones, was precipitated in a near-surface, meteoric phreatic environment. The original mineralogy of a carbonate sediment is also a factor in the extent of meteoric diagenesis. Where there is a high percentage of metastable aragonite and high-Mg calcite, then the sediment has a much higher potential for dissolution and cementation. Sediments dominated by the more stable low-Mg calcite will have a much lower diagenetic potential. This aspect of meteoric diagenesis is explored further in Hird & Tucker (1988) and James & Bone (1989).

4.7.3 Calcite spar

The cement that occupies the majority of the original pore space in many limestones is a clear, equant calcite,

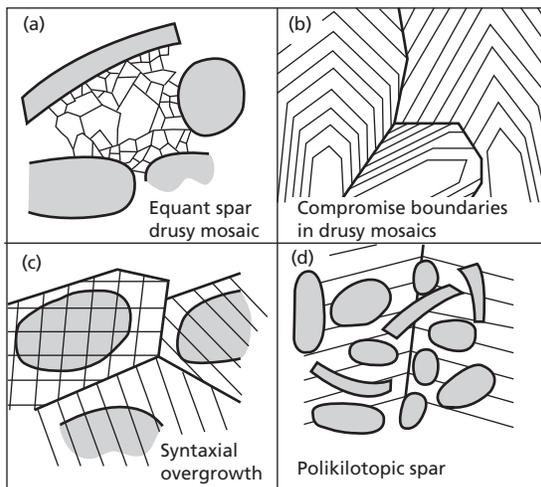


Fig. 4.35 Calcite spar. (a) Drusy calcite spar, the most common type, characterized by an increasing crystal size away from the substrate towards the cavity centre. (b) Growth zones in calcite spar showing how straight crystal boundaries (compromise boundaries) develop between adjacent crystals. (c) Syntaxial overgrowths where calcite spar cement is in optical continuity with a host grain, schematically shown here by twin planes going from cement into grain. (d) Poikilotopic spar, large crystals enveloping several grains.

referred to as sparite or calcite spar (see Fig. 4.35). Sparite possesses a number of features, which, taken together, allow its cement interpretation. These are:

- 1 its location between grains and skeletons, and within original cavities;
- 2 its generally clear nature, with few inclusions;
- 3 the presence of planar intercrystalline boundaries;
- 4 a drusy fabric, i.e. an increase in crystal size away from the substrate or cavity wall;
- 5 crystals with a preferred orientation of optic axes normal to the substrate.

The fabric characteristics of sparite 4 and 5 are reflections of the preferred growth direction of calcite, parallel to the *c*-axis. Calcite spar is invariably precipitated after the fibrous calcite described earlier, which is mostly a marine cement. In some cases, there is a layer of internal sediment between the two cement generations. Calcite cement may also take the form of large poikilotopic crystals, several millimetres to centimetres across (as in Plate 6c). The large crystals are the result of a low nucleation rate and slow growth, perhaps because pore fluids were only just saturated with re-

spect to CaCO_3 . Where echinoderm grains, and others composed of a single calcite crystal, are present in the limestone, then the sparite cement may precipitate syntaxially (in optical continuity) upon the grain to produce an overgrowth (see Plate 12c,d). Preferential cement growth upon such single crystal grains may envelop adjacent, small, polycrystalline grains.

Calcite-spar crystals are commonly delicately zoned as a result of subtle variations in Fe and Mn contents. The zonation can be revealed by staining with Alizarin Red S plus potassium ferricyanide, or by observing the luminescence (see Plate 13b,c and Section 4.1). Studying the zonation pattern in calcite spar in a limestone formation on a regional scale allows a cement stratigraphy to be erected from which the larger-scale hydrology of the basin can be reconstructed (e.g. papers in *Sedimentary Geology*, 65(3/4), 1989; Horbury & Robinson, 1993; Budd *et al.*, 1995).

Carbon and oxygen isotopes are being used increasingly in the study of carbonate cements. Fibrous calcite commonly has a marine signature in both $\delta^{13}\text{C}$ and $\delta^{18}\text{O}$, whereas calcite spar tends to show more negative $\delta^{18}\text{O}$, reflecting precipitation at a higher temperature during burial and/or precipitation from meteoric water (see papers in Schneidermann & Harris (1985) and Horbury & Robinson (1993); and Tucker & Wright, 1990). Fluid inclusions in calcite also provide information on the chemistry of pore fluids and the temperature of precipitation (see Emery & Robinson, 1993; Goldstein & Reynolds, 1994).

Origin of calcite spar. There has been much discussion of the environment of precipitation of sparite, the main cement in all medium- to coarse-grained limestones and the fill of most cavities in fine-grained limestones. Although drusy sparite is the typical cement of near-surface meteoric diagenesis, in many limestones there is evidence for sparite precipitation after compaction. The occurrence of broken and fractured fibrous cement fringes and micrite envelopes around formerly aragonite skeletal grains indicates mechanical compaction resulting from overburden pressure after aragonite dissolution and before sparite precipitation (see Plate 7c). In other limestones, sutured contacts between grains show that chemical compaction occurred before sparite precipitation (see Plate 13a). The pore waters for deep-burial cementation will be either connate (modified seawater buried with the sediments) or meteoric (or a mixture of the two). In

fact, in many sedimentary basins it does appear that pore waters are largely of meteoric origin, although salinity is much higher than that of near-surface fresh water. There are three possible sources for the CaCO_3 : (a) the pore water itself, (b) pressure dissolution within the limestones or at deeper levels and (c) dissolution of CaCO_3 , mainly skeletal aragonite, in calcareous shales interbedded with the limestones. For precipitation of calcite from trapped seawater the Mg/Ca ratio would have to be lowered; the adsorption of Mg onto clay minerals could have this effect. Pressure dissolution has been considered of major importance in cementation in view of the great quantities of CaCO_3 that clearly have been dissolved from pressure dissolution planes, and because of the vast quantities of CaCO_3 that are required from somewhere to produce the fully cemented limestone formations that are seen at the Earth's surface. As noted in Section 4.7.1, in some limestones there is evidence for calcite spar having been precipitated on the sea floor.

4.7.4 Neomorphism

Some diagenetic processes involve changes in the mineralogy and/or fabric of the sediment. For these processes of replacement, once loosely referred to as recrystallization, the term *neomorphism* is now used to include all transformations between one mineral and itself or a polymorph. There are two aspects to neomorphism: the wet polymorphic transformation of aragonite to calcite and the wet recrystallization of calcite to calcite. Both processes are wet because they take place in the presence of water, through dissolution–reprecipitation; dry, solid-state processes, such as the inversion of aragonite to calcite or recrystallization *sensu stricto* of calcite to calcite, are unlikely to occur in limestones, where diagenetic environments are always wet. Most neomorphism in limestones is of the aggrading type, leading to a coarser mosaic of crystals. Two common types are (a) microspar–pseudospar formation from micrite and (b) the calcitization of originally aragonitic skeletons, ooids and cements. Degrading neomorphism results in a finer mosaic of crystals.

A scheme for describing textures and fabrics of neomorphic limestones and dolomites, and other sediments such as evaporites that have been precipitated, crystallized or recrystallized, is given in Table 4.4.

Table 4.4 Terms for describing textures and fabrics of crystal mosaics in sedimentary rocks

For crystal shape:	anhedral—poor crystal shape subhedral—intermediate crystal shape euhedral—good crystal shape	
For equigranular mosaics:	xenotopic—majority of crystals anhedral hypidiotopic—majority of crystals subhedral idiotopic—majority of crystals euhedral	
For inequigranular mosaics:	porphyrotopic—where larger crystals (porphyrotopes) are enclosed in a finer-grained matrix poikilotopic—where larger crystals (poikilotopes) enclose smaller crystals	
Size-scale:	micrometre-sized	0–10 μm
	decimicrometre-sized	10–100 μm
	centimicrometre-sized	100–1000 μm
	millimetre-sized	1–10 mm
	centimetre-sized	10–100 mm

Microspar–pseudospar; aggrading neomorphism

It is not uncommon to find that in fine-grained limestones, the micritic matrix (less than 4 μm) has been locally or even totally replaced by microspar (crystal sizes between 4 and 10 μm) and pseudospar (10–50 μm) (see Plate 13e). This *neomorphic spar* can be recognized by:

- 1 irregular or curved intercrystalline boundaries, commonly with embayments (contrasting with the plane intercrystalline boundaries of sparite cement, Section 4.7.3);
- 2 very irregular crystal-size distribution and patchy development of coarse mosaic;
- 3 gradational boundaries to areas of neomorphic spar;
- 4 presence of skeletal grains floating in coarse spar.

Aggrading neomorphic textures are prominent in micritic limestones and they can give a mottled, almost brecciated appearance, of ‘clasts’ of neomorphic spar/microspar in a micritic matrix. Such pseudo-breccias are common in the British Carboniferous (Solomon, 1989). Neomorphism may have taken place within an original inhomogeneous sediment, such as that resulting from bioturbation. Aggrading neomorphism involves the growth of certain crystals

at the expense of others. It is likely that growth takes place in solution films and cavities between crystals, by syntaxial precipitation on pre-existing crystals. The CaCO_3 will be derived from dissolution of submicrometre-sized crystals and inflowing pore waters. Studies of fine-grained limestones by SEM have shown that much of the microspar is actually a cement rather than the result of aggrading neomorphism. Lasemi & Sandberg (1984) have found relics of aragonite in some ancient micrites and so were able to distinguish aragonite-dominated precursor muds (ADP) from calcite-dominated ones (CDP).

Calcitization of aragonite grains and cements

Bioclasts and ooids composed originally of aragonite are now mostly drusy sparite in limestones, through dissolution of the aragonite and later precipitation of calcite into the void (Section 4.7.1). Where calcite spar has not been precipitated, then biomoulds and oomoulds are present. In some cases, however, the grains have been replaced by calcite with no intervening void phase, a process referred to as *calcitization*. Where this has occurred, features to note are:

- 1 relics of the internal structure of the shell, preserved through inclusions of organic matter and minute crystals of aragonite;
- 2 an irregular mosaic of small and large calcite crystals, with wavy, curved or straight intercrystalline boundaries;
- 3 a brownish colour to the neomorphic spar, owing to residual organic matter, which imparts a pseudo-pleochroism to the crystals (see Plate 8a).

Aragonite cements also endure either dissolution or calcitization, but more tend to be calcitized (see Fig. 4.31).

Degrading neomorphism

Degrading neomorphism, whereby large crystals of CaCO_3 are replaced by smaller calcite crystals, a process of 'crystal diminution', is rare in limestones, and mostly has occurred through tectonic stress or very low-grade metamorphism. The process is most easily observed to have taken place in echinoderm grains. Micritization of skeletal grains by endolithic microbes is not a neomorphic process of course, but does result in a fine-grained mosaic.

4.7.5 Compaction

As discussed in Section 2.9.1, increasing overburden pressure leads to compaction in sediments and two categories are recognized: mechanical and chemical. Mechanical compaction may begin soon after deposition, whereas chemical compaction normally requires more than several hundred metres of burial.

Mechanical compaction in grainy sediments leads to a closer packing of the grains and a rotation of elongate bioclasts towards the plane of the bedding. As the overburden pressure increases, fracture of bioclasts may take place and micritic grains may become squashed and deformed. If there are early cements around grains, these can be spalled off, as can oolitic coatings of grains. Mechanical compaction may lead to collapse of micrite envelopes, if there has been an earlier phase of dissolution of the aragonitic bioclasts (see Plate 7c).

Lime muds generally suffer more compaction during very shallow burial as the sediments dewater. This may lead to burrows being compressed, especially if there is a high clay content in the sediment. Alternating clay-poor and clay-rich, fine-grained limestones, a common deep-water facies, usually show the differential effects of compaction well, with shell fracture and burrow flattening in the latter, but not the former. Compaction can lead to the formation of skeletal packstones from skeletal wackestones, as a result of the closer packing of grains.

Chemical compaction is the result of increased solubility at grain contacts and along sediment interfaces under an applied stress. Mostly this is the result of overburden but tectonic stresses also give rise to pressure-dissolution effects. Three common textures result from chemical compaction: fitted fabrics, stylolites and pressure-dissolution seams. In a grainstone with little or no early cement, sutured and concavo-convex contacts develop between grains (see Plate 13d). If the pressure dissolution between grains is intense, then a *fitted fabric* may be produced. This can be on a microscopic scale, as in oolitic and echinoderm grainstones, or on a macroscale, between intraclasts, fossils and early lithified diagenetic nodules and burrow fills in a muddy, compactible sediment. A *stylolite breccia* texture may be produced.

Stylolites are through-going sutured surfaces that cut grains, cement and matrix indiscriminately (see

Plate 14a; see also Fig. 9.10). Clay, iron minerals and organic matter, the insoluble residue from the limestone's dissolution, are usually concentrated along the stylolites. In more argillaceous limestones, *dissolution seams* are smooth, undulose and anastomosing horizons of insoluble residue. They generally pass around grains and early diagenetic nodules, and where abundant, the term *flaser limestone* is used.

Pressure dissolution is an important process accentuating bedding planes, particularly between more muddy and more grainy sediments, and it may also lead to the development of *pseudobedding planes* (see Section 4.6.1 and Simpson, 1985). Calculations have shown that considerable amounts of CaCO_3 can be liberated by pressure dissolution, so that this process is often cited as one of the main sources of CaCO_3 for limestone cementation, especially of late-diagenetic, burial calcite spar. Burial diagenesis is reviewed in Choquette & James (1987).

4.8 Dolomitization, dedolomitization and silicification

It is not uncommon to find that ancient limestones have been partially or even completely dolomitized. The dolomite so formed can be replaced by calcite in the process of dedolomitization (Section 4.8.3). Limestones also can be silicified to various degrees (Section 4.8.4). These diagenetic processes commonly result in an obliteration of sedimentary and petrographic details.

4.8.1 Dolomites

The mineral dolomite is a rhombohedral carbonate belonging to the $\bar{3}$ trigonal/hexagonal crystal system. Ideally, it consists of an equal number of Ca^{2+} and Mg^{2+} ions arranged into separate sheets with planes of CO_3^{2-} anions between. The well-ordered nature of the dolomite lattice results in a series of superstructure reflections on X-ray diffraction (XRD), which are not present in the structurally similar calcite. The height of the ordering d_{015} peak relative to the d_{110} peak gives a measure of the degree of ordering. Most modern dolomites have a low degree of ordering compared with older dolomites. The term protodolomite was introduced for Ca–Mg carbonates made in the laboratory with no, or very weak, ordering reflections. Many natural dolomites are also not stoichiometric

(Ca:Mg is 50:50), but have an excess of Ca^{2+} ions, up to Ca:Mg of 58:42. The Ca^{2+} substitution for Mg^{2+} increases the lattice spacing, and this also can be measured with XRD by the shift in the position of the d_{104} peak (see Hardy & Tucker (1988) for details of the technique). Iron substitution is common in dolomites, giving ferroan dolomite with a few mol.% FeCO_3 ; ankerite $\text{CaMg}_{0.5}\text{Fe}_{0.5}(\text{CO}_3)_2$ may be associated.

The replacement of CaCO_3 minerals by dolomite and the precipitation of dolomite cement may take place soon after the sediments have been deposited, i.e. penecontemporaneously and during early diagenesis, or a long time after deposition, usually after cementation, during burial. The term primary has often been applied to dolomite, implying a direct precipitate from sea or lake water. In fact the majority of dolomites have formed by replacement of pre-existing carbonate minerals, although dolomite cements are common. The word dolomite is used for both the mineral and the rock type; the term *dolostone* has been used for the latter. Carbonate rocks are divided on the basis of dolomite content into:

limestone	0–10% dolomite
dolomitic limestone	10–50% dolomite
calcitic dolomite	50–90% dolomite
dolomite (dolostone)	90–100% dolomite

To convey an indication of grain or crystal size in a dolomite, the terms *dolorudite*, *dolarenite*, *dolosparite* and *dolomicrite* can be used. In many cases, if the original structure has not been destroyed completely, the dolomites can be described in terms of Dunham's (or Folk's) classification (see Section 4.4), preceded by the word dolomitic, or prefixed by *dolo-*. For the description of textures and fabrics the scheme of Table 4.4 can be used.

Two common types of dolomite mosaic are *xenotopic* and *idiotopic*, the former of anhedral crystals with curved to serrated and irregular crystal boundaries, and the latter of euhedral rhombic crystals (see Fig. 4.36 and Plate 14a). The features to note when describing dolomite textures are shown in Table 4.5.

The preservation of the original limestone texture in a dolomite varies from completely fabric destructive with no obvious relics of the original sediment (see Plate 14a,c), to fabric retentive, with good to perfect preservation of original structure (see Fig. 4.3 and Plate 14b for examples). Dolomitization also may be fabric selective, only destructively replacing the matrix

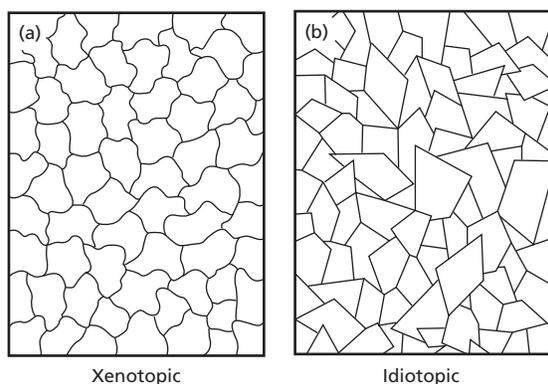


Fig. 4.36 Dolomite textures. (a) Xenotopic. (b) Idiotoxic.

Table 4.5 Terms for describing dolomite textures

Dolomite crystal size	unimodal or polymodal
Dolomite crystal shape	anhedral, subhedral, euhedral
Dolomite mosaic	xenotopic, hypidiotoxic, idiotoxic
Crystal type	limpid, rhombic, baroque
Dolomite cement	isopachous, drusy
CaCO ₃ grains	unreplaced or replaced or moulds; if replaced: partial or complete, mimic or non-mimic
CaCO ₃ matrix	Unreplaced or replaced
Void-filling dolomite	limpid, rhombic, drusy or baroque

and not the grains, or only replacing certain bioclasts. Dolomite rhombs may be distributed randomly through the limestone (as in Plate 13f). Original crystal/grain size and mineralogy are important controls. An original micritic sediment generally is dolomitized to a fine-grained mosaic, so that primary sedimentary structures are still preserved. Grains of high-Mg calcite, such as red algae, some foraminifers and echinoderms, can be dolomitized with little fabric alteration (*mimic replacement*). By way of contrast, aragonitic grains (e.g. molluscs) are either dolomitized with much fabric alteration, or the aragonite is dissolved out and the mould filled by dolomite cement (in a similar manner to aragonite altering to calcite). Low-Mg calcite grains may resist dolomitization or be dolomitized destructively. The timing of dolomitization is also a factor, because if it is late, during burial, then it is very likely that the original, mixed-mineralogy sediment has already stabilized to low-Mg calcite, so that the dolomite is fabric destructive.

Dolomite crystals are commonly zoned; in many cases the inner part is more cloudy (from fluid inclusions or calcite relics) and the outer part is clear. Dolomite cements, as opposed to replacements, occur in primary and secondary cavities in many limestones and dolomites. They vary from cavity linings of clear, 'limpid' rhombs to drusy mosaics similar to calcite spar (as in Fig. 4.5).

One type of dolomite, which may be a replacement or a cement, is *baroque* or *saddle dolomite* (see Plate 14c). It is also known as pearlspar. The crystals generally are large (many millimetres) and have conspicuous, curved crystal faces. In thin-section, they have curved cleavage and undulose extinction. They commonly contain inclusions (fluids or mineral relics) and many are ferroan. Baroque dolomite commonly is associated with sulphide mineralization, hydrothermal activity and also hydrocarbons. It often is considered typical of burial dolomitization and the characteristic lattice distortion is attributed to variations in the concentration of Ca²⁺ ions adsorbed onto the growing crystal surface (see review of Spötl & Pitman, 1998).

The distribution of dolomites in the stratigraphic record is not even, and it has been said frequently that dolomites increase in abundance back in time. Dolomites do appear to be more common than limestones in the Precambrian, and this has led to the suggestion that seawater had a different composition then, so that dolomite could be precipitated directly, or could replace CaCO₃ more easily. Alternative views are that dolomitization environments were more prevalent as a result of palaeogeographical and palaeoclimatic differences, or that simply through being older, the limestones have had more time in which to become dolomitized. A survey by Given & Wilkinson (1987) suggested two maxima of dolomite occurrence in the Phanerozoic (Jurassic–Cretaceous and mid-Palaeozoic), which broadly correspond to the highstands in the first-order global sea-level curve, indicating a role of geotectonics and hydrosphere–atmosphere chemistry in dolomitization. Sun (1994) suggested that the abundance of Phanerozoic dolomites coincided with periods of extensive peritidal deposition and large-scale evaporite basins. Burns *et al.* (2000) thought that periods of more extensive dolomitization correlated with decreased oxygen levels in the atmosphere and oceans, and that this fostered more active communities of anaerobic microbes, which promoted dolomite formation.

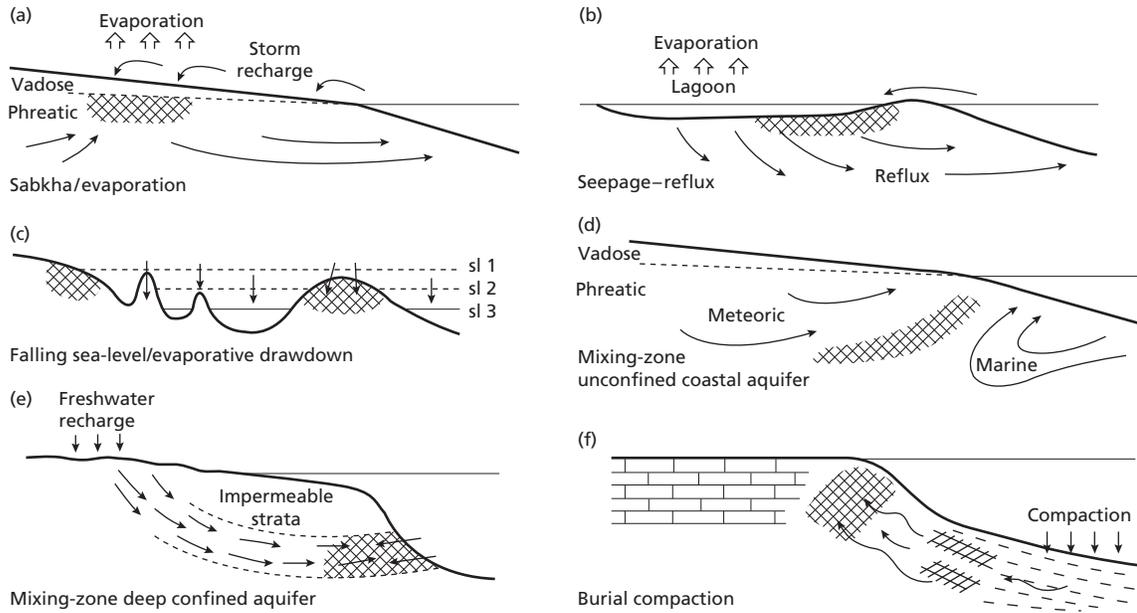


Fig. 4.37 Models of dolomitization, illustrating the variety of mechanisms for moving dolomitizing fluids through the sediments. For seawater dolomitizing models see Fig. 4.38.

4.8.2 The origin of dolomites and dolomitization models

There is still much debate and argument over the origin of dolomite, particularly concerning the pervasive dolomitization of extensive limestone platforms. One problem with dolomite is that it is difficult to manufacture in the laboratory under sedimentary–diagenetic conditions using natural waters. It is thus difficult to unravel the chemical controls on dolomite precipitation. Seawater is supersaturated with respect to dolomite, but with its highly ordered structure, dolomite appears to be inhibited from direct precipitation by various kinetic factors (the high ionic strength of sea-water, Mg^{2+} hydration and low CO_3^{2-} activity) so that aragonite and high-Mg calcite, with their simpler structure, are precipitated instead. The major considerations in the dolomitization of a limestone are the source of the Mg^{2+} and the mechanism by which the dolomitization fluids are moved through the rocks (see reviews of Tucker & Wright, 1990; Purser *et al.*, 1994). Models for dolomitization are shown in Figs 4.37 & 4.38.

At the present time, dolomites are being precipitated in intertidal–supratidal sediments of the Bahamas, Florida and Trucial Coast (Arabian Gulf). The dolomite itself, poorly ordered and calcium-rich, consists mostly of 1–5 μm rhombs, occurring within the sediments or forming hard surface crusts. There is much evidence in the form of dolomitized gastropods and pellets for replacement, but direct precipitation also takes place (Lasemi *et al.*, 1989). In some of these modern peritidal occurrences, it appears that the dolomite is being precipitated from seawater,

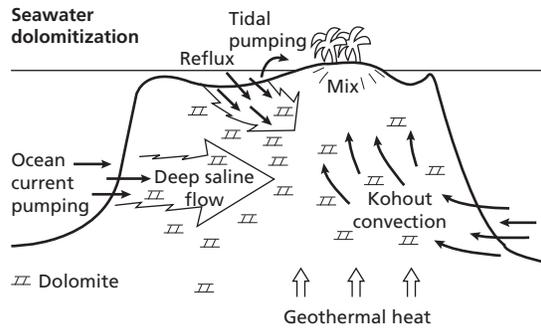


Fig. 4.38 Models for seawater dolomitization of limestones, all basically different ways of pumping seawater through a carbonate platform.

evaporated to a greater or lesser extent. The Mg/Ca ratio of the pore fluids is raised by the precipitation of aragonite and gypsum–anhydrite, and this promotes dolomite precipitation. Reduced SO_4^{2-} also helps. The Coorong in South Australia is a series of coastal lakes where dolomite is being precipitated, but not in association with evaporites. The dolomite forms where Mg^{2+} -rich continental ground water discharges into shallow lakes and is evaporated. Dolomite does precipitate directly from saline lake waters; this occurs in Victoria, Australia (Deckker & Last, 1989), where Mg/Ca ratios are high from weathering of basalts nearby. In the shallow-marine environment, dolomite is forming in Baffin Bay, a shallow, periodically hypersaline lagoon with restricted connection to the Gulf of Mexico and in a lagoon near Kuwait.

In the geological record, many fine-grained dolomites with evidence of peritidal deposition (planar stromatolites, birdseyes, polygonal cracks, tepees and intraclasts) have probably formed through evaporitic dolomitization and direct dolomite precipitation. Such dolomites generally show good preservation of the sedimentary structures, as in Plate 10e.

A model for dolomitizing shallow subtidal and reef limestones is *seepage-reflux*, whereby high Mg/Ca ratio fluids are generated in lagoons and beneath tidal flats and sabkhas (Section 5.2.1) by evaporation and these descend into the subsurface through density contrasts with marine pore waters. Unfortunately, there are no large-scale modern examples, but the model is applied frequently to dolomite formations closely associated with evaporites, such as those in the Upper Permian (Zechstein) basin of northwest Europe. Dolomitization of platforms and reefs around evaporite basins could well occur during drawdown (Fig. 4.37c, see Kendall, 1989).

Another model that has been very popular, especially for relatively early, near-surface dolomitization of limestone where there are no associated evaporites, is the *meteoric-marine, mixing-zone* model. As noted above, seawater is supersaturated with respect to dolomite but because of kinetic factors, dolomite normally does not precipitate unless the Mg/Ca ratio is raised (as in the evaporative model), and dolomite is forced out of solution. Therefore, it has been argued, dolomite precipitation is more likely to take place from dilute solutions (where there would be fewer interfering ions present) and at slow crystallization

rates. Calculations have suggested that the mixing of meteoric ground waters with up to 30% sea water would cause undersaturation with respect to calcite but increasing saturation for dolomite. Dolomitization has thus been predicted for seawater–meteoric-water mixing zones, where salinities are reduced but Mg/Ca ratios are maintained. In the mixing model, the Mg^{2+} is derived from seawater, the Mg/Ca ratio is maintained above unity and ground-water movement pumps the dolomitizing fluid through the limestones. Although a most attractive hypothesis and applied to many pervasively dolomitized limestones without associated evaporites in the 1970s and 1980s, the mixing-zone model has come in for much criticism and is now abandoned. In essence, the very slow rate of dolomite precipitation (because of the ordered nature of the crystals) relative to calcite dissolution would mean that dolomite could not precipitate in significant quantities in a mixing zone. See Machel & Mountjoy (1986) and Hardie (1987) and reviews in Tucker & Wright (1990) and Purser *et al.* (1994).

Much dolomite is precipitated during the burial of a limestone formation, but there has been much debate as to whether this could result in the pervasive dolomitization of whole carbonate platforms. The driving force in this model is considered to be compaction of basinal mudrocks and the expulsion of Mg^{2+} -rich fluids into adjacent platform-margin limestones. The Mg^{2+} is thought to be derived from clay minerals, marine pore waters and high-Mg calcite, but mass-balance calculations have suggested that there is insufficient Mg^{2+} for large-scale dolomitization. The higher temperature of the burial environment should mean that some of the kinetic obstacles to dolomite precipitation are overcome. However, another problem appears to be that there is not a continuous source of fluid in the burial environment. Hydrothermal fluids may circulate, especially through fractures and faults and lead to local veins of baroque dolomite. Scattered rhombs, dolomite crystals along stylolites and pressure-dissolution seams, and late cavity-filling dolomite cements are common forms of burial dolomite in many limestones. Burial dolomites generally are coarse and fabric destructive; baroque dolomite is an especially common form. Burial dolomites have been well described from the Cambrian Bonneterre of Missouri, the Upper Devonian of Alberta (e.g. Dix, 1993; Drivet & Mountjoy, 1997), the Carboniferous of England and the Triassic of the

Dolomites (see papers in Purser *et al.*, 1994; Kupecz *et al.*, 1997; and Yoo *et al.*, 2000).

In recent years, attention has focused on the possibility that sea water alone, perhaps with a little modification, is the dolomitizing fluid, and a number of mechanisms have been put forward to drive seawater through the sediment (see Fig. 4.38). Discoveries of modern dolomite in the marine environment are significant here. Beneath Enewetak atoll in the Pacific, Eocene strata at a depth of 1250–1400 m are being dolomitized. As a result of the cooler water at this depth, seawater is just undersaturated with respect to calcite, but still supersaturated with regard to dolomite. Seawater is pumped through the atoll by oceanic tides and upwards by thermal convection as a result of the high heat flow out of underlying volcanic basement. Within the Bahama Platform, there also is a strong circulation of sea water as a result of ocean currents such as the Gulf Stream impinging on the Bahama escarpment. This *ocean current/tidal pumping* could play a major role in dolomitization because it drives huge volumes of seawater through carbonate platforms. The high geothermal gradient beneath carbonate platforms is thought to generate a large-scale convection system, which draws cold seawater into the platform. Diagenetic changes, including dolomite precipitation, in periplatform ooze on the Bahama slopes have been attributed to this *Kohout convection*. Lagoonal or platform-interior seawater is frequently a little more saline (40–45‰) than open-ocean water (35‰) and this is sufficient to cause *reflux* into the underlying sediments. Restudy of tidal-flat dolomites in the Florida Keys has suggested that they are forming through *tidal pumping* of Florida Bay water through the sediments, rather than through simple evaporation of seawater. Experiments have suggested that SO_4^{2-} in seawater is an important kinetic inhibitor of dolomitization. The SO_4^{2-} is reduced by microbial activity and this takes place in sediments containing organic matter. Dolomite being precipitated from marine pore waters within anoxic pelagic sediments in the Gulf of California and off Peru, and in lagoonal muds of Brazil, could well be the result of microbial mediation and sulphate reduction. Indeed, there is a growing body of evidence that microbes play a major role in dolomite precipitation generally; see review of Burns *et al.* (2000), and Teal *et al.* (2000). One other mechanism that will move seawater through sediments is relative sea-level change. There also is much fluid

movement within the meteoric–marine mixing zone, which will generate pore-water circulation within the adjacent marine phreatic zone. Thus, now, seawater dolomitization is very popular; it provides the necessary Mg ions and there are several powerful processes for circulating seawater through a carbonate platform.

The origin of dolomites remains an enigma but careful field, petrographic and geochemical study does permit an elimination of some dolomitization models and implication of others. An analysis of the stable isotope and trace-element geochemistry of dolomites, as well as fluid inclusion data, can help to distinguish different generations and to make inferences on the nature of the pore waters involved: hypersaline/marine/mixed/connate. $^{87/86}\text{Sr}$ isotope data can be useful too, to identify the origin of the fluids, and even date the time of dolomitization.

An important consequence of many styles of dolomitization is that porosity is increased (see Plate 14a,b). Dolomite has a more compact crystal structure than calcite so that theoretically the complete dolomitization of a limestone results in a porosity increase of 13%, as long as there is no subsequent compaction or cementation. In addition, dissolution of relict calcite in dolomitized limestone or increasing calcite dissolution relative to dolomite precipitation during the dolomitization process creates extra porosity. Thus, dolomitization is important for hydrocarbon reservoir potential. Many oilfields of western Canada, for example, are in dolomitized Devonian reef limestones (see papers in Roehl & Choquette (1985) for case histories). Many papers on dolomite are contained in Shukla & Baker (1988) and Purser *et al.* (1994).

4.8.3 Dedolomitization

Dolomite may be replaced by calcite to produce limestone again. This calcitization process is referred to as dedolomitization and predominantly takes place through contact with meteoric waters. Calcite replacement of dolomite commonly is associated with the dissolution of gypsum–anhydrite, a near-surface phenomenon as well (Section 5.5). Burial dedolomitization also may occur. Recognition of ‘dedolomites’ is similar to that of replaced evaporites, a question of noting dolomite crystal shapes (rhombohedra) occupied by calcite (pseudomorphs), or calcite crystals with replacement fabrics (see neomorphic spar,

Section 4.7.4) containing small relict inclusions of dolomite (see Plate 14d). In some cases the original limestone texture is partially regenerated on dedolomitization; in other instances layers and concretions of fibrous calcite randomly and completely replace the dolomite. The dissolution of dolomite rhombs may lead to a mouldic porosity (see Plate 14e).

4.8.4 Silicification

Silicification, like dolomitization, can take place during early or late diagenesis. It takes the form of selective replacement of fossils or the development of chert nodules and layers (Section 9.4). Silica also occurs as a cement in some limestones. The main types of diagenetic silica in limestones are: (a) euhedral quartz crystals, (b) microquartz, (c) megaquartz and (d) chalcedonic quartz. They are described in Section 9.2 and shown in Figs 9.1, 9.2 & 9.11. Both length-fast and length-slow chalcedonic quartz occur and the latter may indicate the former presence of evaporites (Section 5.5). Sponge spicules are the main source of silica, together with diatoms and radiolarians (Section 9.4).

4.9 Porosity in carbonate sediments

The porosity of carbonate sediments shortly after deposition is very high: sand-sized sediments around 50%, lime mud around 80%. Porosity is lost or reduced through cementation, compaction and pressure dissolution, and gained through dissolution, dolomitization and tectonic fracturing.

Porosity in limestones can be divided into two main types: primary (depositional) and secondary (diagenetic–tectonic). Three common types of primary porosity are:

- 1 framework porosity, formed by rigid carbonate skeletons such as corals, stromatoporoids and algae, especially in reef environments;
- 2 interparticle porosity in carbonate sands, dependent on grain-size distribution and shape;
- 3 porosity in carbonate muds provided by fenestrae (birdseyes) and stromatactis (Section 4.6.3).

Secondary porosity includes:

- 1 moulds, vugs and caverns formed by dissolution of grains and rock, commonly through leaching by meteoric ground waters, but also by basinal (connate) waters;

- 2 intercrystalline porosity produced through dolomitization;

- 3 fracture porosity, formed through tectonic pressures, and through collapse and brecciation of limestone as a result of dissolution, such as of interbedded evaporites, or the limestone itself in karstification.

Primary porosity, and also secondary, is commonly facies controlled. Certain facies, such as reefs, fore-reefs and oolites have high primary porosities, whereas others have low porosities, lagoonal micrites and outer-ramp carbonates, for example, unless affected by the diagenetic–tectonic processes leading to porosity creation. Studies of carbonate facies distributions, cementation patterns and diagenesis, in particular dolomitization, coupled with porosity–permeability measurements are thus all required to detect any reservoir potential. Examples of carbonate hydrocarbon reservoirs are: the Upper Jurassic Arab Formation of Saudi Arabia, with a primary intergranular and dolomite porosity; Middle and Upper Devonian reef and fore-reef limestones of western Canada and the Ordovician Trenton Limestone of northeastern USA, both with a porosity, at least in part, as a result of dolomitization; the Ellenburger of Oklahoma and Texas, with karstic porosity, and the Upper Cretaceous Chalk of the North Sea and the Tertiary Asmari Limestone of Iran, both with fracture porosity. Carbonate reservoirs are described in Roehl & Choquette (1985), Kupecz *et al.* (1997) and Harris *et al.* (1999), and porosity is reviewed by Moore (1989) and Lucia (1995, 1999).

4.10 Carbonate depositional environments and facies

4.10.1 Non-marine carbonate sediments

Lacustrine limestones

Lacustrine carbonates are of three principal types: (a) inorganic precipitates, (b) algal/microbial sediments and (c) skeletal sands. Inorganic precipitation, producing lime muds, mostly takes place through evaporation, but CO₂ loss, as a result of plant photosynthesis or pressure–temperature changes, and mixing of fresh stream or spring water with saline lake water, also causes carbonate precipitation. Tufa mounds are spectacularly developed in some lakes (e.g. Mono Lake, California) as a result of sublacustrine

trine springs. Precipitation in shallow, agitated zones may produce ooids, as in the Great Salt Lake, Utah and Pyramid Lake, Nevada. The mineralogy of the carbonate mud and ooids precipitated depends largely on the Mg/Ca ratio of the water. Aragonite, calcite (high and low Mg) and dolomite (as mud not ooids) may all be precipitated.

Lime muds also may be produced through the activities of algae, cyanobacteria and microbes, and from phytoplankton blooms. The main role of cyanobacteria, however, is in the formation of stromatolites, common in modern lakes (e.g. Great Salt Lake) and in many ancient lake formations (e.g. the Green River Formation, Wyoming and Utah and the Pliocene Ridge Basin of California). Oncoids also occur, those from Lake Constance, Switzerland, being especially well known. Skeletal sands contain fragments of calcareous algae, such as *Chara*, as well as bivalves and gastropods.

Lacustrine carbonates are arranged in a similar facies pattern to their marine counterparts. Stromatolite 'reefs' and ooid shoals occur in more agitated, shallow waters, with lime muds occurring shoreward on littoral flats and in protected bays, and in the central deeper parts of lakes. One characteristic feature of lake-basin deposits is a rhythmic lamination, consisting of carbonate–organic matter couplets, often interpreted as seasonal in origin.

Two broad types of lake can be distinguished:

- 1 hydrologically open—these have an outlet and so are relatively stable;
- 2 hydrologically closed—these have no outflow and so are subject to rapid changes in lake level through fluctuations in rainfall and outflow.

Evaporation may exceed inflow so that a saline lake develops, where evaporites may be precipitated (see Section 5.4.2). Perennial and ephemeral closed lakes are recognized. In the geological record, both open and closed-lake carbonates occur in the Green River Formation, western USA, and perennial–ephemeral saline-lake carbonates, mostly dolomites, occur in the Cambrian Officer Basin, South Australia (Southgate *et al.*, 1989). Stromatolites, laminites, fish beds and dolomites were formed in the thermally stratified Devonian Orcadian lake of northeast Scotland (Trewin & Davidson, 1999).

Shallow-water, muddy lacustrine facies are commonly modified by pedogenesis and the term *palustrine* is used for these deposits, which usually are

nodular and mottled. They are widely developed in the Tertiary of the Mediterranean area, Cretaceous of Spain (Platt, 1989), and the Devonian Catskills of New York (Dunagan & Driese, 1999). The Everglades of Florida are a good modern analogue (Platt & Wright, 1992).

Reviews and descriptions of lake carbonates can be found in Matter & Tucker (1978), Dean & Fouch (1983), Anadon *et al.* (1991) and Talbot & Allen (1996).

Calcrete or caliche

In many parts of the world where rainfall is between 200 and 600 mm year⁻¹, and evaporation exceeds this precipitation, calcareous soils are formed. They are typically seen in river floodplain sediments but they also develop in other continental sediments (aeolian, lacustrine and colluvial deposits), and in marine sediments too, should they become subaerially exposed. Many terms are applied to these pedogenic carbonates but calcrete and caliche, the latter chiefly in the USA, are widely used. Calcrete occurs in several forms, from nodules to continuous layers, with massive, laminated and pisolitic textures. Many calcretes form in the upper vadose zone through a *per descensum* process of dissolution of carbonate particles in the upper A horizon of the soil profile and reprecipitation in the lower B horizon (see Fig. 3.7). They also may form within or just below the capillary-rise zone, where they are referred to as phreatic or *ground-water calcretes*, well known from southwest Australia. Some of these are composed of dolomite (*dolocrete*; e.g. Colson & Cojan, 1996). Calcretes develop in time from scattered to packed nodules (Fig. 4.39), to a massive limestone layer. The amount of time involved varies but is of the order of several to tens of thousands of years.

The characteristic fabric of calcretes is a fine-grained equigranular calcite mosaic with floating quartz grains through displacive growth (Fig. 4.40). Grains and pebbles may be split in this displacive process. Replacement of some grains also takes place. Circumgranular cracks and spar-filled veins also are common. Many calcretes possess spar-filled tubules, formerly occupied by rootlets, and lacey, micritic 'septa' within these, formed by calcification of fungal filaments, give the so-called alveolar texture. Rhizcretions (rootlet encrustations) and microbial grain coatings are common in some calcretes. Pisoids

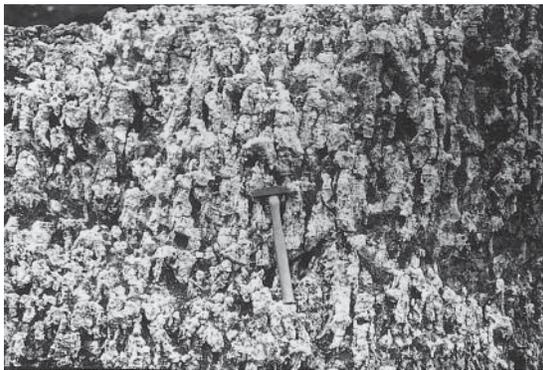


Fig. 4.39 Calcrete, consisting of closely packed elongate nodules that have grown in a river floodplain sediment. Old Red Sandstone, Devonian. Gloucestershire, England.

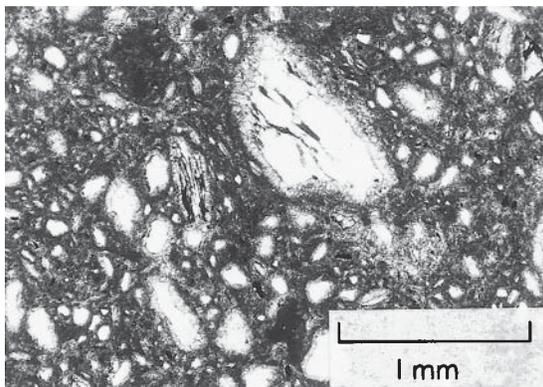


Fig. 4.40 Photomicrograph of modern calcrete showing dense micritic calcite, which has displaced the quartz grains so that the latter are now not in contact. There is also an envelope of coarser calcite crystals around each quartz grain; this is a typical feature of calcretes. Plane-polarized light. Recent calcrete. Almeria, Spain.

(‘vadoids’) form in some calcareous soils from calcification of fungal–bacterial clusters. Some of these show evidence for *in situ* growth: dominant downward-directed laminae, reverse-grading and a fitted, almost polygonal arrangement of pisoids. Others accrete during imperceptible downslope movement of the soil. Laminated crusts are associated with calcretes and pisoids and some form from calcification of root mats. Black pebbles may be present (Section 4.3.1). Reviews on calcretes have been presented by Wright & Tucker (1991) and Retallack (1997). Specific descriptions include Williams & Krause (1998) for Devonian cal-

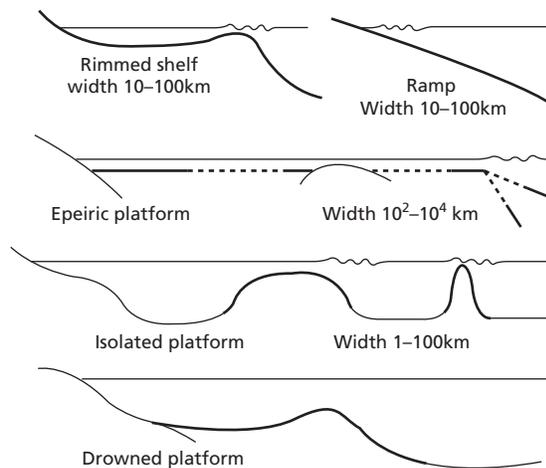


Fig. 4.41 Different types of carbonate platform.

cretes in floodplain mudrocks, and Beier (1987) for modern Bahamian calcretes.

4.10.2 Marine carbonates and carbonate platforms

Thick successions of mostly shallow-marine limestone are a feature of the geological record. *Carbonate platform* is a general term widely used for these and they typically develop along passive continental margins, in intracratonic basins, failed rifts and foreland basins. Five major categories of carbonate platform are recognized: the rimmed shelf, ramp, epeiric, isolated and drowned platforms (Fig. 4.41). Each platform type has a particular pattern of facies and facies succession. The *rimmed shelf* (Fig. 4.42) is a shallow-water platform with a distinct break-of-slope into deeper water. Reefs and carbonate sand bodies occur along the high-energy shelf margin, restricting the circulation in the shelf lagoon behind to a greater or lesser extent. Along the shoreline, depending on the energy level and tidal range, tidal flats or a beach-barrier complex will be present. Debris from the rimmed-shelf margin is shed onto the adjacent slope and into the basin. Modern rimmed shelves occur off South Florida, Belize and Queensland (Great Barrier Reef). The *carbonate ramp* (Fig. 4.43) is a gently sloping surface with a generally high-energy, inner-ramp shoreline passing offshore to a quiet, deeper-water outer ramp, affected periodically by storms. Along the shoreline there may be a beach-barrier–tidal-delta complex, with lagoons and tidal

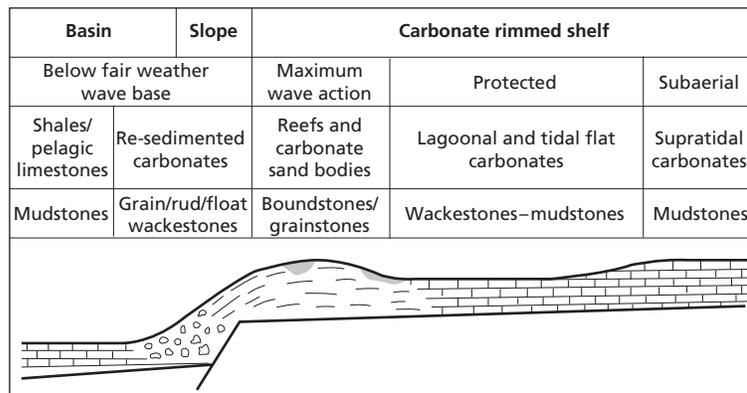


Fig. 4.42 General facies model and limestone types for a rimmed shelf.

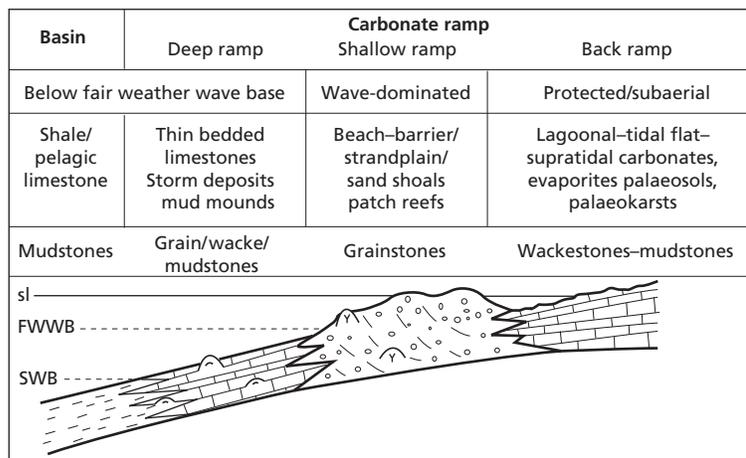


Fig. 4.43 General facies model and limestone types for a carbonate ramp.

flats behind, or a beach-ridge/strandplain system. Large reefs generally are not present on ramps, but patch reefs may occur on the inner ramp, and pinnacle reefs and mud-mounds in deeper water on the outer ramp. In effect, the carbonate ramp is equivalent to a siliciclastic open shelf. Modern carbonate ramps are located off the Yucatan Coast of Mexico, the Trucial Coast of the Arabian Gulf and Shark Bay of Western Australia. The *epeiric platform* is a very extensive (100–10 000 km across), relatively flat cratonic area covered by shallow sea. Along the margin, there may be a gentle (ramp-like) or steep (shelf-like) slope into the adjoining basin, but the margin is not a major feature of the platform. Within the platform itself, there may be deeper-water basins, surrounded by ramps and rimmed shelves. Epeiric platforms are dominated by low-energy, shallow subtidal–intertidal sediments.

Storms and tidal currents may be important. There are no good modern examples of epeiric platforms of the size that existed in the past, but the interior of the Great Bahama Bank and Florida Bay may be close analogues.

Isolated or detached platforms are surrounded by deep water and so are very much affected by prevailing wind and storm directions. They vary in size from a few kilometres to a few hundred kilometres across. The Bahamas is a large, modern example; smaller ones occur off the Belize shelf and in the Red Sea. Detached platforms consist of pure carbonate sediment because terrigenous material, apart from wind-blown dust, is excluded. The margins of these platforms may be shelf or ramp in character. *Drowned platforms* are ones that have suffered a relatively rapid sea-level rise so that deeper-water facies are deposited over shallow-water

facies. Many pelagic limestones were deposited in these situations.

The type of carbonate platform developed is determined largely by tectonics and relative sea-level change, although one type commonly evolves into another. For discussions, reviews and case histories of carbonate platforms see Wilson (1975), Crevello *et al.* (1989), Tucker & Wright (1990), Tucker *et al.* (1990), Loucks & Sarg (1993), Wright & Burchette (1996), Harris *et al.* (1999) and Insalaco *et al.* (2000). The carbonate sediments of the major depositional environments are now briefly described.

4.10.3 Intertidal–supratidal carbonates

Tidal flats are areas regularly to rarely covered by water, dominated by weak currents and wave action. They are developed extensively upon epeiric platforms and they occur along the shorelines of low-energy shelves and ramps, typically behind beach-barriers and around lagoons. Tidal-flat carbonates are dominantly lime mudstones, commonly peloidal, although local lenses of coarser sediment (grainstone) may represent tidal-channel fills. Fenestrae (Section 4.6.3) are the characteristic structure, giving rise to the distinctive birdseye limestone (Fig. 4.24). The fauna may be restricted in diversity; gastropods in particular may abound, together with ostracods, foraminifers and bivalves. Thin, coarse layers of subtidal skeletal grains may occur, transported onto the tidal flat by storms. Microbial mats and stromatolites (Section 4.3.3) are typical of tidal-flat deposits. Many are simple planar varieties, showing desiccation cracks and laminoid fenestrae (e.g. Figs 4.12 & 4.25). Small domes may develop and in higher-energy areas, columnar stromatolites. Bioturbation is common, and rootlets may occur. Syndimentary cementation of tidal-flat sediments can produce surface crusts, which may expand to form tepee structures (see Fig. 4.22) and megapolygons (Section 4.6.1). Crusts may break up to give intraclasts, reworked into edgewise conglomerates and flakestones (Fig. 4.5). Penecontemporaneous dolomitization may take place, giving fine-grained dolomite mosaics (Section 4.8.1). In arid climatic areas, the evaporite minerals gypsum–anhydrite, and possibly halite, will develop in the sediment. They may be preserved as pseudomorphs (Section 5.5). Slight uplift and contact with meteoric waters may result in palaeokarstic surfaces, laminated crusts, calcretes,

vadose pisoids and black pebbles. Tidal-flat sediments commonly occur at the top of shallowing-upward cycles (see Section 4.10.9 and Fig. 4.54). Modern carbonate tidal flats and their ancient equivalents have been described by Ginsburg (1975), Hardie (1977, 1986), Shinn (1983), Tucker & Wright (1990) and Demicco & Hardie (1995).

4.10.4 Lagoonal limestones

Lagoons are subtidal areas located behind barriers, which may be reefs or carbonate sand shoals. They occur on rimmed shelves (shelf lagoon) and along the inner part of carbonate ramps. Protected, lagoonal-type environments are widely developed on epeiric platforms too. Organisms living in lagoons, and therefore the sediments accumulating in these predominantly quiet-water areas, depend largely on the degree of restriction and permanence of the barrier. Lagoons may be normal in terms of salinity, as in the lagoons of atolls; brackish where there is much freshwater runoff, as in the inner part of Florida Bay, for example; or hypersaline to a greater or lesser extent, such as the inner part of the Great Bahama Bank, Shark Bay (Western Australia), and the lagoons of the Trucial Coast (Arabian Gulf). The sediments are variable in grain size, although many are carbonate muds, rich in peloids. Aggregates are common. Towards the barrier the muds pass into coarser sediments, and coarse skeletal debris may be derived from small, coral patch reefs, which commonly grow in lagoons of normal salinity. The lagoon floor is dominated by molluscs, green algae and foraminifers. The green algae in particular are a major sediment contributor (Section 4.3.3), and microbes play a significant role in skeletal breakdown and production of micritized grains. Surficial microbial mats and sea-grasses may cover the lagoon floor, as in the Bahamas and Shark Bay (Section 4.3.3). Bioturbation is intensive, largely through the activities of crustaceans and bivalves. Sedimentary structures may be poorly developed, although thin, vaguely graded beds of coarser grains and shell lags may be formed through periodic storm reworking.

Lagoonal limestones are common in the geological record, particularly in back-reef and back-barrier situations passing shorewards into tidal-flat facies. Many are wackestones and mudstones, in some cases with fossils preserved in growth position. Examples are the *Amphipora* wackestones of Devonian back-

reefs in western Canada, Europe and Australia, the calcisphere-rich *porcellanites* of the Carboniferous, and the thick-shelled megalodont bivalve facies of Triassic back-reefs.

4.10.5 Intertidal–subtidal carbonate sand bodies

These sand bodies occur in areas of high tidal current and wave activity and include barriers, beaches, the shoreface and tidal deltas along ramp shorelines, and shoals and banks along exposed, rimmed-shelf margins. Depths of deposition are mostly less than 5–10 m. The sediments are carbonate sands and grainstones composed of ooids and rounded and sorted skeletal grains. The latter are fragments of normal-marine organisms: corals, bivalves, foraminifers and algae, and echinoderms and brachiopods in the Mesozoic–Palaeozoic. Sedimentary structures are ubiquitous, chiefly cross-bedding on all scales, perhaps with herringbone cross-bedding through tidal-current reversals (Fig. 2.23), reactivation surfaces, HCS+SCS, also planar bedding, scours and channels (see Section 2.3).

Carbonate shoreline sands are similar to their siliciclastic equivalents (see Section 2.11.5), in terms of their facies succession and sedimentary structures. The northeast Yucatan coast of Mexico is a modern example of a carbonate strandplain and the Trucial Coast of the Arabian Gulf is a barrier–lagoon–tidal-delta system. The Jurassic Lincolnshire Limestone of eastern England was deposited in such a barrier-island–lagoon system and the Jurassic Great Oolite of southern England was also deposited along an inner-ramp shoreline. Oolites in the South Wales Carboniferous were deposited in strandplains and barriers of a carbonate ramp (Burchette *et al.*, 1990). The Jurassic Smackover Formation of the Gulf Coast Rim was deposited in similar inner-ramp environments, as were Mississippian grainstones in the Illinois Basin (see papers in Wright & Burchette (1998)).

Shelf-margin sand bodies are best developed along windward shelf-margins, where skeletal debris also may be derived from nearby reefs. The shoals are elongate parallel to the shelf-break and traversed by tidal channels that have lobes of sediment (spillovers) located at their ends. The shoals are covered in sand-waves and dunes that mostly are orientated lagoonwards. Lily Bank on the northeast margin of the Little

Bahama Bank is a good example of this sand-body type. Through time, windward sand shoals may prograde into the lagoon to generate a muddy to grainy, coarsening-upward unit. Continued growth of a sand shoal may lead to the development of a sand flat, with islands. This has happened at Joulter's Cay in the Bahamas over the past 5000 years. In areas where tidal currents are strong, sand bodies are ridges orientated more normal to the shelf break, with grass-covered muddy sands in the depressions between. Such oolite ridges are prominent on the platform margins at the heads of the deep channels cutting into the Bahama Banks (e.g. Schooner and Exuma Cays, see Gonzalez & Eberli, 1997). Lime sand also accumulates along more leeward shelf-margins, and here off-shelf transport of sediment to the adjoining foreslope is a major process in platform progradation. This has occurred along the western side of the Great Bahama Bank, generating clinofolds (see Section 4.10.8 and Eberli & Ginsburg, 1989).

Occurring in high-energy locations, where water is continuously pumped through the sands, these grainstones commonly have marine cements, and hardgrounds may be formed. Ancient examples of shelf-margin, carbonate sand bodies include Zechstein oolites in northeast England (Kaldi, 1986). See papers in Harris (1984), the review of Handford (1988) and Mutti *et al.* (1996).

Oolitic and skeletal grainstones are important hydrocarbon reservoirs. The Jurassic Smackover Formation of the Gulf Coast Rim has substantial oil reserves. The type of porosity varies considerably over the region, from primary intergranular to dissolutional to intercrystalline where dolomitized. The original mineralogy of the ooids, aragonite to the north (landwards) and calcite to the south (basinwards) is a major factor in porosity evolution (Heydari & Moore, 1994).

Carbonate sands are also deposited in deeper waters on carbonate ramps by the action of storms. The sediments are grainstones and packstones, especially where depths are close to wave-base, passing into wackestones and lime mudstones in deeper water. Storm waves and currents generate HCS in above storm-wave-base lime sands, and shell lags and graded beds below storm wave-base (see Section 2.3.2). Bioturbation is widespread. Common skeletal components are molluscs, chiefly bivalves, foraminifers and coralline algae, and brachiopods and echinoderm

debris in the Mesozoic and Palaeozoic. Terrigenous silt and clay are an important constituent of many outer-ramp and deep-shelf limestones. Storm-deposited limestones ('tempestites') are described by Aigner (1984) from the Triassic in Germany, Handford (1986) from the Mississippian of Arkansas and Jennette & Pryor (1993) from the Ordovician of Ohio.

4.10.6 Reefs and carbonate build ups

Although coral reefs are one of the most familiar and most studied of modern carbonate environments, there are many other types of reef developing at the present time and preserved in the geological record. Such carbonate *build ups*, a term widely used for locally formed limestone bodies that had original topographic relief, are a common feature of many carbonate formations going back to the Precambrian. The literature on modern and ancient reefs is vast; reviews and compilations include Toomey (1981), James (1983), Fagerstrom (1987), Geldsetzer (1989), Loucks & Sarg (1993), Monty *et al.* (1995), Harris *et al.* (1999) and Wood (1999).

The term reef itself is best restricted to a carbonate build up that possesses(ed) a wave-resistant framework constructed by organisms, but to be clear the

term ecological reef or organic framework reef can be used for this. Specific types of such ecological reef, shown in Fig. 4.44, are *patch reef*, small and circular in shape; *pinnacle reef*, conical; *barrier reef*, separated from the coast by a lagoon; *fringing reef*, attached to the coast; and *atoll*, enclosing a lagoon. Other terms frequently used are *bioherm* for local *in situ* organic growth with or without framework; *biostrome* for laterally extensive *in situ* growth with or without framework; *organic bank* or *loose skeletal build up* for an accumulation of mostly skeletal sediment, chiefly through trapping or baffling; *mud-mound* or *mud bank* (formerly reef knoll) for an accumulation of mostly lime mud (micrite), probably by trapping and baffling.

Many different organisms can be and have been involved in the construction of reefs. At the present time, the main reef builders are corals and coralline algae; others of limited importance are sponges, serpulids, oysters and vermetid gastropods. In the past, practically all invertebrate groups have at one time or another contributed to reef growth. Special mention can be made of cyanobacteria and microbes (formerly blue-green algae) constructing stromatolite bioherms/biostromes in the Precambrian and Cambrian (locally in younger rocks too), stromatoporoids in the Ordovician to Devonian, rugose corals in the Silurian to

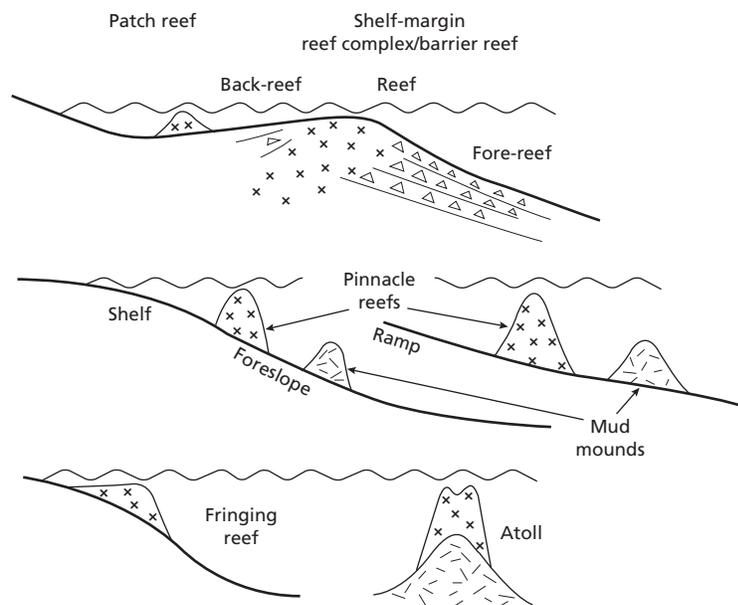


Fig. 4.44 The main types of reef and their location.

Carboniferous, scleractinian corals from the Triassic to Recent, phylloid algae in the Carboniferous and Permian, sponges in the Triassic–Jurassic and rudistid bivalves in the Cretaceous (see Kiessling *et al.* (1999) for a review).

Organisms in reefs take three roles: (i) the frame-builders, those providing a skeletal framework (corals at the present time); (ii) the frame-binders and encrusters, organisms which consolidate the framework, such as calcareous algae and bryozoans; and (iii) the reef-users, such as boring bivalves and sponges, predatory fish and echinoderms. With many ancient reefs it is clear that there was no true solid framework, but much *in situ* organic growth. This last feature gives rise

to the two typical features of reef limestones, a massive appearance with no stratification (Fig. 4.45) and the presence of organisms in growth position (Fig. 4.46). Many reef limestones would be classified as boundstones, with the terms framestone, bindstone and bafflestone used to describe specific types (see Section 4.4); some reef-rocks do not have an obvious reef structure on the hand-specimen or thin-section scale and are a wackestone, lime mudstone or even a grainstone. Primary cavities are a feature of many reefs, although they are commonly filled with skeletal debris and cement. Syndimentary cementation is a feature of many modern and ancient reefs (Section 4.7.1).

There are many factors controlling the growth of



Fig. 4.45 Devonian reef limestones in the Canning Basin, Western Australia. The reef consists of massive unbedded limestone (central part of cliff); in front of the reef (to the left) occurs a talus slope of reef debris, these fore-reef limestones have an original dip; behind the reef (to the right) are flat-bedded limestones of the back-reef lagoon.

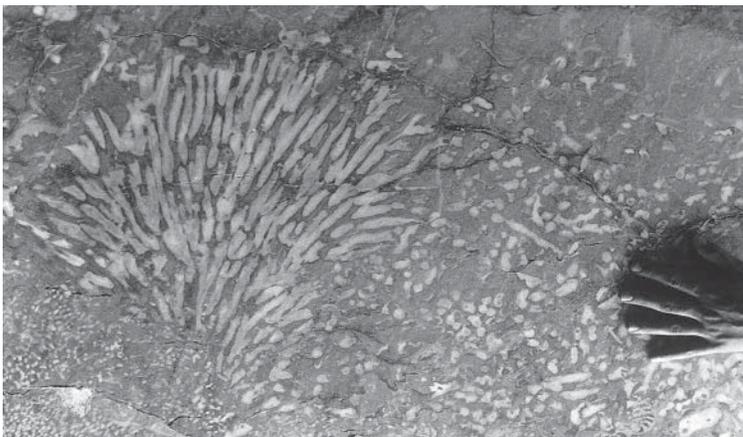


Fig. 4.46 *In situ* coral colony (*Thecosmilia*) from Triassic reef. Adnet, Austria.

modern coral reefs and it is likely that these same factors exerted an influence on coral and other reefs in the past. For coral-reef growth, these factors are:

- 1 water temperature—optimum growth occurs around 25°C;
- 2 water depth—most growth takes place within 10 m of the surface;
- 3 salinity—corals cannot tolerate great fluctuations;
- 4 turbidity and wave action—coral growth is favoured by intense wave action and an absence of terrigenous silt and clay.

The majority of reefs occur along shelf margins, an agitated zone where waves and currents of the open sea first impinge on the sea floor. Smaller patch reefs develop in open lagoons behind shelf-margin reefs, on carbonate ramps and epeiric platforms. Reefs, usually atolls, are also developed on submerged volcanic islands within the ocean basins. The configuration and morphology of some present-day reefs is a reflection of karstic dissolution of earlier reefs during glacial low-stands of sea-level.

Many modern reefs along shelf margins show a characteristic threefold division into fore-reef (reef front/slope), the reef itself (reef-crest, reef-flat) and

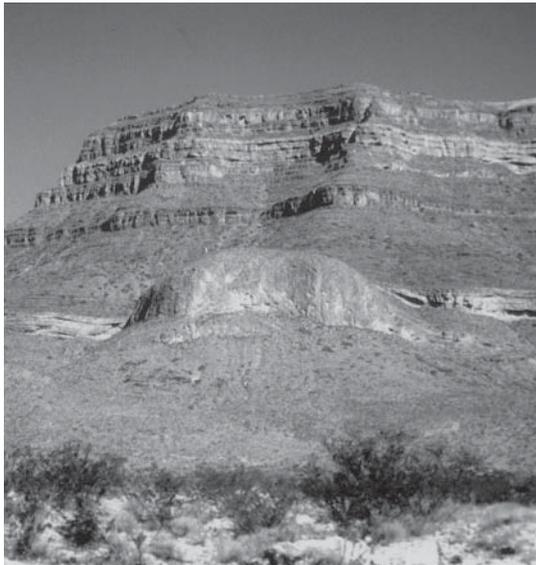


Fig. 4.47 Lower Carboniferous mud-mound (muleshoe mound), Sacramento Mountains, New Mexico. Photograph courtesy of Dave Hunt.

back-reef. The reef front is a steep slope, vertical in places, with organisms constructing reef in the upper part, passing down to a talus slope of coarse reef debris. Reef-derived carbonate breccias, debrites and turbidites may be present in the adjoining basin. A prominent system of surge channels gives a spur and groove morphology along the reef front, extending up to the reef flat, in some cases. The reef crest, covered by no more than 1–2 m of water, is the site of prolific organic growth, of corals and algae on modern reefs. Behind the crest is the reef-flat, a pavement of mostly dead coral. The back-reef area consists of reef debris adjacent to the reef-flat, passing shoreward to a quiet-water lagoon, where there may be patch reefs. A broadly similar facies pattern is seen in many ancient build ups (Fig. 4.45). Classic examples include the Permian Capitan Reef in Texas (see Saller *et al.*, 2000), the Devonian reefs of western Canada, Europe and Australia, and the Triassic reefs of the Northern Calcareous Alps (Austria) and Dolomites (Italy).

Mud-mounds are massive accumulations of lime mudstone and wackestone, tens to hundreds of metres across, which pass laterally into well-bedded limestone. The mud-mounds contain only scattered fossils; those of some possible significance are crinoids, bryozoans, sponges and algae. Much of the mud is peloidal. Mud-mounds typically occur in deeper water on carbonate ramps and shelf slopes. The best developed mud-mounds occur in the Palaeozoic, in the Carboniferous of northwest Europe, where some are referred to as Waulsortian reefs, the Carboniferous of New Mexico and Montana, and the Devonian of New York State and Belgium (Fig. 4.47).

The origin of mud-mounds is still a problem. Processes that could be involved are (a) the precipitation of lime mud by microbes, (b) entrapment of mud through the baffling action of bryozoans and crinoids and (c) concentration of mud into mounds by currents. Comparisons have been made with the mud banks of Florida and Shark Bay, where lime mud and skeletal debris is baffled and trapped by sea-grass and algae (Bosence, 1995). The sediment itself is derived from green-algal disintegration (Section 4.3.4) and the breakdown of larger skeletal grains. Many papers on mud-mounds can be found in Monty *et al.* (1995). Perhaps the most spectacular examples occur in the Algerian Sahara (Wendt *et al.*, 1997).

Reefs are important hydrocarbon reservoirs with the most porous facies generally occurring in the upper

fore-reef and reef-framework facies. However, marine cementation is most prevalent in this zone and so porosities can be reduced from high primary values. The talus at the toe of reef slopes also makes good reservoirs. Many reefs are dolomitized, and also this enhances their reservoir qualities. Examples include the Devonian reefs of western Canada, such as the Leduc and Golden Spike, and the Cretaceous rudist reefs of the Gulf Coast Rim, Mexico and the Middle East. Rudist bivalves have very porous skeletons but much of this intraskeletal porosity is not connected. The best reservoirs are thus located in rudist debris beds, rather than in the reefs themselves (which tend to be muddy) or in rudist limestones that have suffered much mechanical compaction during burial.

4.10.7 Pelagic limestones

Where water depth is too great for benthic organisms to flourish, in excess of some 50–100 m, then carbonate sediments composed of pelagic organisms will accumulate in the absence of clay. The maximum depth of accumulation is controlled by the rate of carbonate dissolution. In low latitudes, the ocean is saturated with respect to CaCO_3 in the upper few hundred metres and below this it becomes undersaturated, first with respect to aragonite, and then calcite. Below a few hundred metres CaCO_3 begins to dissolve, but it is not until greater depths that the rate of CaCO_3 dissolution increases substantially (this depth is the *lysocline*) (see Fig. 4.48). The depth at which the rate of dissolution is balanced by the rate of supply is known as the *carbonate compensation depth* (the CCD). This depth varies in the oceans, its position being controlled by calcareous plankton productivity, which itself depends largely on nutrient supply, and water temperature. In the tropical regions of the world's oceans, the CCD for calcite is between 4500 and 5000 m; the CCD for aragonite is about 2000 m less. The CCD shallows into higher latitudes and seawater is undersaturated in respect of CaCO_3 in temperate and polar waters. Calcareous oozes can accumulate on the sea floor, which is shallower than the CCD; siliceous oozes and red clays are present below this depth. Fluctuations in the CCD back into the Cenozoic and Mesozoic are now well documented.

Modern pelagic carbonates are composed of pteropods (aragonitic), coccoliths (Fig. 4.9) and foraminifers (both calcitic), and are found on outer

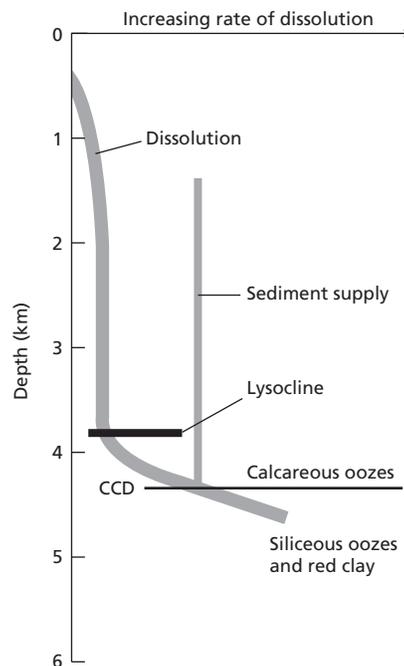


Fig. 4.48 Carbonate dissolution in the ocean. Below the carbonate compensation depth (CCD), carbonate sediment does not accumulate.

continental shelves, continental slopes and ocean floors starved of terrigenous clay, and on submarine rises, drowned reefs and volcanoes (seamounts and guyots) rising from the ocean floor. Ancient pelagic limestones occur in the Mesozoic of the Alpine region, the Ammonitico Rosso, Maiolica, Biancone and Scaglia, for example, and in the Devonian and Carboniferous of Hercynian Europe, the Cephalopodenkalk and Griotte. Characteristic features of pelagic limestones, apart from a dominantly pelagic fauna (e.g. Fig. 4.17), are their condensed nature and evidence for syndimentary cementation in the form of hardgrounds, lithoclasts, sheet cracks and neptunian dykes. Many pelagic limestones are nodular (Fig. 4.49) and some contain ferromanganese nodules and crusts (Section 6.7). Resedimentation of pelagic sediment is common too, particularly in slumps and debrites.

The Cretaceous chalks of northwest Europe and southern USA are composed largely of coccoliths and are thus pelagic limestones. Deposition took place at depths of around 50–150 m, so that there is a signifi-

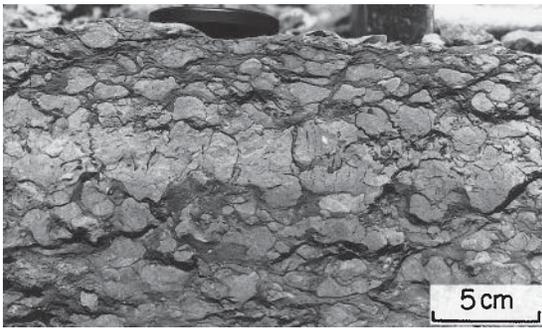


Fig. 4.49 Pelagic limestone: a micritic, nodular limestone that contains a pelagic fauna. Pressure dissolution effects are common and result in the clay seams (flasers) between nodules. Devonian. Montagne Noire, France.

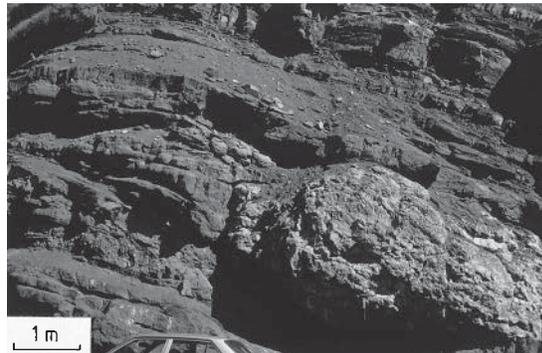


Fig. 4.50 Block of reef-rock in deep-water hemipelagic mudstones, which also contain limestone turbidites of shallow-water skeletal debris. Triassic. The Dolomites, Italy.

cant benthic macrofauna of echinoids, bivalves and brachiopods. Hardgrounds are common within the chalks and these are mineralized with phosphate and glauconite. In the Chalk of the North Sea, resedimentation is widespread, with major slides and slumps, debrites and turbidites, all greatly expanding the thickness of the formation. The North Sea Chalk is a major hydrocarbon reservoir and depositional as well as diagenetic factors were important in the acquisition of reservoir qualities. The most porous horizons are commonly those that were resedimented, rather than the pelagic chalks. Fracturing of the chalk through up-doming of Zechstein evaporites has greatly improved the permeability. Early oil entry and overpressuring inhibited burial compaction and cementation. See Glennie (1998) for more information on North Sea petroleum geology.

Papers on pelagic limestones are contained in Hsü & Jenkyns (1974); also see the reviews of Scholle *et al.* (1983), Tucker & Wright (1990) and Clari & Martire (1996).

4.10.8 Resedimented deep-water limestones

Shallow-water carbonate sediment is transported into the deep sea by the same processes as discussed for siliciclastics (see Section 2.11.7): slides, slumps, debris flows, turbidity currents and modified grain flows. In the slope area between shelf and basin, slump-folded limestones and slides are common, especially in pelagic-hemipelagic slope facies. Megabreccias, including very large blocks of shallow-water limestone,

are common off major shelf-margin reefs (e.g. Fig. 4.50), and many have formed through sea-level falls (Spence & Tucker, 1997). Carbonate debrites are very variable in texture, commonly forming a spectrum from unsorted and chaotic breccias, to inversely and normally graded breccias, perhaps with associated graded grainstones (A to E in Fig. 2.78). Such beds are prominent in the Devonian of the Canadian Rockies and Cambro-Ordovician Cow Head Group of Newfoundland. In the basins themselves, limestone turbidites are usually interbedded with hemipelagic dark shales (Fig. 4.51). Sole structures, graded and planar bedding and cross-lamination are all developed in limestone turbidites, just as in siliciclastic examples, although in many, good Bouma divisions (see Section 2.11.7) are not present.

Carbonate turbidite basins are usually fed from a line source, i.e. the whole platform margin, rather than one (or several) point sources(s) as in the case of many siliciclastic turbidite basins. Thus the submarine fan model is not appropriate for most limestone turbidite formations, and in fact very few ancient, carbonate submarine fans have been described. Resedimented limestones are usually ascribed to either the slope apron (see Fig. 4.52), or base-of-slope apron models, depending on whether the resedimented limestones interfinger with the shallow-water shelf limestones (the first model) or whether the upper slope is bypassed and sediment accumulates at the toe of the slope (the second model). Ancient examples of slope aprons are well developed in the Triassic of the Dolomites (Bosellini, 1984), the Permian Capitan Reef of Texas



Fig. 4.51 Limestone turbidites composed of crinoidal and other skeletal grains interbedded with hemipelagic mudrocks. Succession is inverted. This is the typical appearance of turbidites (both carbonate and siliciclastic) in laterally continuous beds with interbedded deep-water mudrocks. Devonian. Cornwall, England.

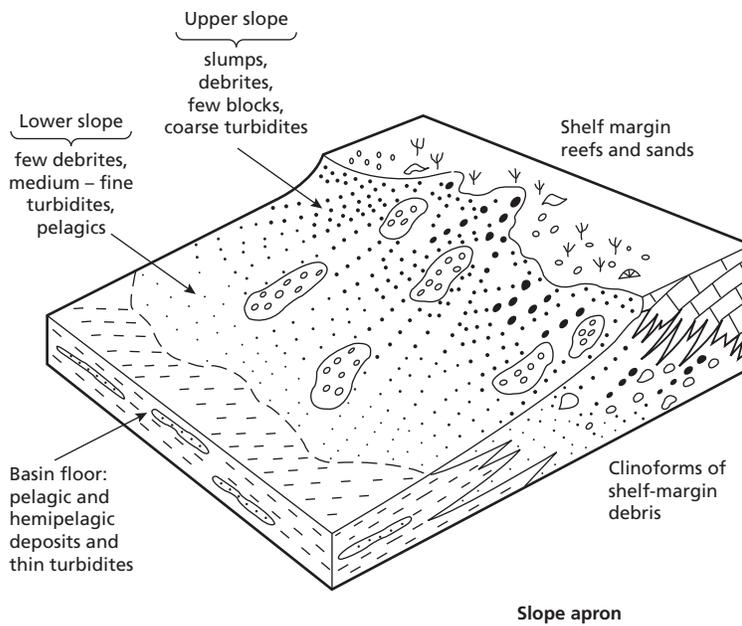


Fig. 4.52 Facies model for a carbonate slope apron where slope facies interdigitate with platform margin facies. In the base-of-slope apron model, not shown here, sediment mostly bypasses the upper slope and is deposited at the toe of the slope. From Tucker & Wright (1990).

and the Devonian Canning Basin reefs of Western Australia (Fig. 4.45), where large-scale, quite steeply dipping sheets of shallow-water debris extend up to the platform margin. These megascale cross-beds are called *clinoforms*, a term coming from seismic stratigraphy for dipping reflectors. Base-of-slope apron deposits are known from the Lower Palaeozoic of the Appalachians and Western Cordillera in North

America (Mullins & Cook, 1986). Resedimented Cretaceous rudist limestones are important oil reservoirs in the Golden Lane Atoll of Mexico (Enos & Stephens, 1993). They were deposited in a base-of-slope apron with debris coming from shelf-margin rudist reefs. Basin-margin, resedimented carbonates are commonly the sites of syngenetic mineralization. Hydrothermal fluids ascending along faults have

precipitated base-metal sulphides in the porous limestones in a number of cases (see Anderson & McQueen, 1982).

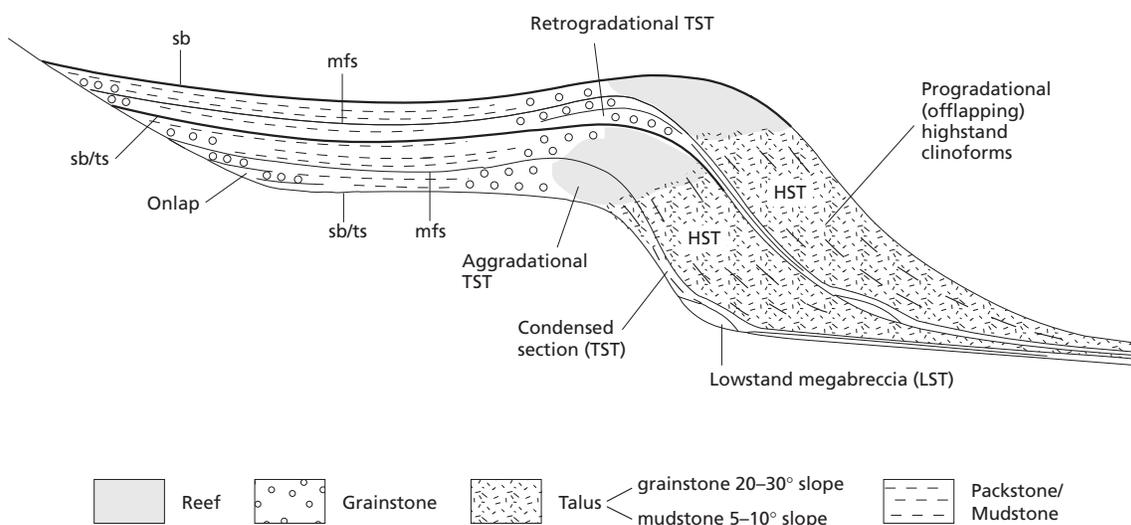
4.10.9 Carbonate sequences

Carbonate sedimentation responds readily to relative changes in sea-level and within many carbonate formations, large-scale and small-scale cycles can be identified. As noted in Chapter 1, relative sea-level changes operate on many scales. For carbonates, the long time-scale second- (10^7 years) and third- (10^6 years) order sea-level changes are responsible for the development of the carbonate platforms themselves and the deposition of hundreds of metres of limestone. Within these platforms, depositional sequences can be identified, by recognizing the key surfaces, and their component systems tracts, from the large-scale arrangement of facies packages, in terms of their onlap (retrogradation), aggradation and offlap (progradation) or from the facies successions, reflecting relative sea-level changes. Rimmed shelf margins, where sedimentation rates are high, are strongly affected by relative sea-level changes and stationary (aggradational), offlapping, onlapping, back-stepping, drowned and emergent types have all been described. In the concepts of sequence stratigraphy, most carbonate deposition takes place within the transgressive and highstand systems tracts, especially the latter when large quantities of sediment are deposited in the basin (*highstand shedding*) and

clinoforms are well developed. During sea-level lowstands, carbonate platforms are usually emergent and so subject to extensive meteoric diagenesis and karstification. However, collapse of the margin may take place at this time to generate megabreccias. Figure 4.53 shows a sequence stratigraphic model for a rimmed shelf. For papers on carbonate sequence stratigraphy see Loucks & Sarg (1993) and Harris *et al.* (1999).

Metre-scale cycles (or parasequences) are common within platform carbonate sequences. These are the result of depositional processes and fourth- and fifth-order relative sea-level changes, operating on a 10^4 – 10^5 years scale. Many of these cycles display a shallowing-upward trend. There are many types but a common one consists of shallow subtidal sediments passing up into tidal-flat facies, with evidence of emergence at the top (Fig. 4.54). The latter may take the form of a palaeokarstic surface or palaeosol. Mixed siliciclastic–carbonate cycles (see Fig. 2.68) and evaporite–carbonate cycles (see Figs 5.16 & 5.17) also occur. The parasequences may be repeated many times

Fig. 4.53 Sequence stratigraphic model for a rimmed shelf. There is much possible variation in the response of carbonate sedimentation to relative sea-level change. Here, the first sequence shows the case of reefal sedimentation keeping up with the sea-level rise during the TST so that the sediments aggrade, whereas in the second sequence the sea-level rise exceeds carbonate production in the TST and the shelf-margin sands back-step, i.e., retrograde. Abbreviations as for Fig. 2.86.



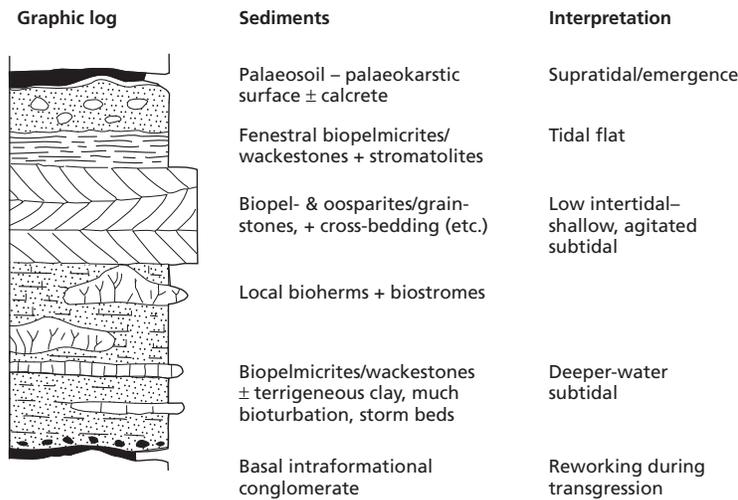


Fig. 4.54 Shallowing-upward limestone cycle (a parasequence): there are many variations on this general theme, depending largely on the energy level of the shoreline and on the climate. Such cycles are typically 1–10 m thick.

in a succession, and they can show systematic variations in thickness and character upwards. Upward-increasing thickness and proportion of subtidal facies in each successive cycle generally indicate an overall transgressive trend, and upward-decreasing thickness of cycles and increasing proportion of tidal-flat facies reflect an overall regressive trend. Many parasequences have formed through tidal-flat progradation and the lateral, seaward migration of facies belts. However, there has been much argument over the causes of the repetition of the cycles. Purely sedimentary processes have been invoked in a migrating tidal-island model and loss of carbonate source area. A tectonic mechanism of periodic fault movement or jerky subsidence has also been suggested (e.g. Satterley, 1996). However, it is currently very popular to invoke sea-level changes though ocean-water volume changes, especially via ice-caps, brought about by orbital forcing in the Milankovitch band (20 000–400 000 years). Statistical analysis and computer modelling of carbonate cycles have lent support to an astronomical control. See the reviews and papers by Tucker & Wright (1990), Wilkinson *et al.* (1997), Lehrmann & Goldhammer (1999) and Lehrmann *et al.* (2000).

Relative sea-level changes are a major factor in the diagenesis of limestones as well as their sedimentation. Long periods of relative sea-level rise lead to the pumping of seawater through carbonate sediments, and this can result in extensive marine cementation and dolomitization. Extended periods of relatively low

sea-level allow meteoric water to penetrate a carbonate formation, and, under a humid climate, this may generate dissolutional porosity, surface and sub-surface karst, calcite cementation and mixing-zone-related dolomitization. If the climate is more arid, then evaporite precipitation and reflux dolomitization are likely early diagenetic processes. See the reviews of Read & Horbury (1993) and Tucker (1993).

Further reading

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