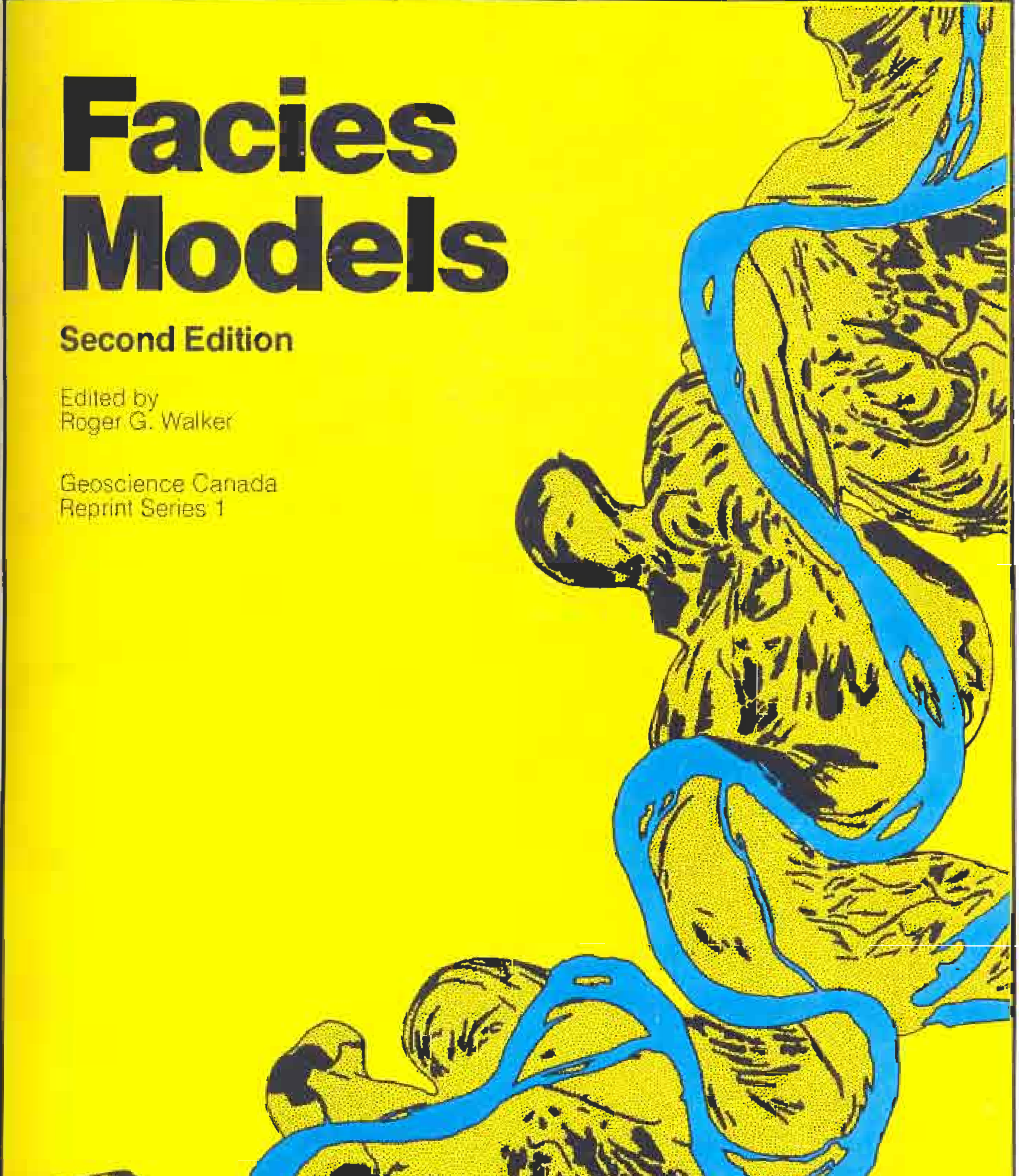
Facies Models

Second Edition

Edited by
Roger G. Walker

Geoscience Canada
Reprint Series 1



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Facies Models, Second Edition

Edited by

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A fully rewritten version of the first edition, with several new contributions. Most of the papers in the first edition appeared originally in Geoscience Canada, 1976-1979, published by the Geological Association of Canada.

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Introduction to Carbonate Facies Models

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INTRODUCTION

This paper is a general introduction to facies models in carbonate sedimentary rocks. Here I would like to set the stage for these papers by outlining the inherent differences between siliciclastic and carbonate deposits (Table 1) and discussing some of the attributes of carbonate sediments which are important to the formulation of facies models (Table 2).

CARBONATE SEDIMENTS ARE BORN, NOT MADE

This deceptively simple phrase encapsulates the main theme of the differences between the two sediment types. Siliciclastic sediments made primarily by the disintegration of parent rock are transported to the environment of deposition, and once there, the patterns of texture and fabric are impressed upon the sediment by the hydraulic regimen. The signature of siliciclastic facies is thus in sedimentary structures and grain size variations. Carbonate sediments, on the other hand, are born in or close to the environment of deposition. Thus, in addition to the purely physical sedi-

Table 1
Differences between siliciclastic and carbonate sediments.

Carbonate Sediments	**Siliciclastic Sediments**
The majority of sediments occur in shallow, tropical environments	Climate is no constraint, sediments occur worldwide and at all depths
The majority of sediments are marine	Sediments are both terrestrial and marine
The grain size of sediments generally reflects the size of organism skeletons and calcified hard parts	The grain size of sediments reflects the hydraulic energy in the environment
The presence of lime mud often indicates the prolific growth of organisms whose calcified portions are mud size crystallites	The presence of mud indicates settling out from suspension
Snallow water lime sand bodies result primarily from localized physicochemical or biological fixation of carbonate	Shallow water sand bodies result from the interaction of currents and waves
Localized buildups of sediments without accompanying change in hydraulic regimen alter the character of surrounding sedimentary environments	Changes in the sedimentary environments are generally brought about by widespread changes in the hydraulic regimen
Sediments are commonly cemented on the sea floor	Sediments remain unconsolidated in the environment of deposition and on the sea floor
Periodic exposure of sediments during deposition results in intensive diagenesis, especially cementation and recrystallization	Periodic exposure of sediments during deposition leaves deposits relatively unaffected
The signature of different sedimentary facies is obliterated during low grade metamorphism	The signature of sedimentary facies survives low-grade metamorphism

Table 2
The sedimentary aspect of modern carbonate producing and binding organisms and their counterparts in the fossil record.

Modern Organisms	**Ancient Counterpart**	**Sedimentary Aspect**
Corals	Archaeocyathans, Corals, Stromatoporoids, Bryozoans, Rudistid bivalves, Hydrozoans.	The large components often remain in place, forming reefs and mounds.
Bivalves	Bivalves, Brachiopods, Cephalopods, Trilobites and other arthropods.	Remain whole or break apart into several pieces to form sand and gravel-size particles.
Gastropods, Benthic Foraminifers	Gastropods, Tintinids, Tentaculitids, Salterellids, Benthic Foraminifers, Brachiopods.	Whole skeletons that form sand and gravel-size particles.
Planktonic forminifers	Planktonic foraminifers, Coccoliths (post-Jurassic).	Medium sand-size and smaller particles.
Encrusting foraminifers and coralline algae	Coralline algae, Phylloid algae, Renalcids, Encrusting Foraminifers.	Encrust on or inside hard sustrates, build up thick deposits or fall off upon death to form lime sand particles.
Codiacean algae-*Penicillus*	Codiacean algae-*Penicillus*-like forms.	Spontaneously disintegrate upon death to form lime mud.
Blue-green algae (Cyanobacteria)	Blue-green algae (especially in Pre-Ordovician).	Trap and bind fine-grained sediments to form mats and stromatolites.

mentary parameters used in the analysis of non-carbonate sediments, the composition of the sedimentary particles themselves is equally important in characterizing the depositional environment. The particles may either be precipitated out of seawater (e.g., ooids) or formed by organisms (e.g., corals and clams).

VARIATIONS OF CARBONATE PRODUCING ORGANISMS WITH TIME

To interpret ancient sedimentary sequences and construct facies models we rely heavily upon observations in modern environments of deposition. This approach works and is seen to work because the basic composition of most sedimentary particles has remained the same through time; a quartz sand grain or an ooid is generally the same in the Pleistocene, Permian or Precambrian. Because organisms have changed with time it is difficult, at first glance to compare modern and specific ancient carbonate facies.

The approximate diversity, abundance, and relative importance of the principal groups of calcareous marine organisms as sediment producers through the Phanerozoic are outlined in Figure 1. It appears from this diagram that there has been a gradual shift in major players through time. In spite of the variations shown in Figure 1, I think that carbonate secreting organisms in the rock record, when viewed as sediment producers, do have living equivalents in modern oceans, although they may not even be in the same phyla. This is because, despite the numerous groups of organisms with hard parts, there are only two ways in which these hard parts are arranged: 1) as whole, rigid skeletons (foraminifers, snails, corals), and 2) as numerous individual segments held together in life by soft organic matter (trilobites, clams, fish). Table 2 lists the more important carbonate producing and binding organisms their sedimentary aspect and their fossil equivalents.

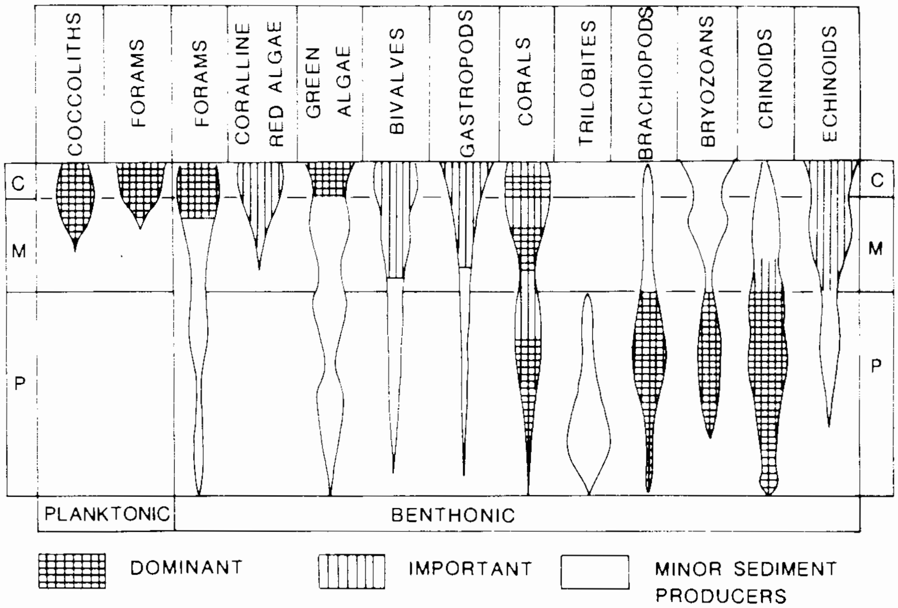


Figure 1
The approximate diversity, abundance and relative importance of various calcareous marine organisms as sediment producers (modified from Wilkinson, 1979). P = Paleozoic; M = Mesozoic; C = Cenozoic

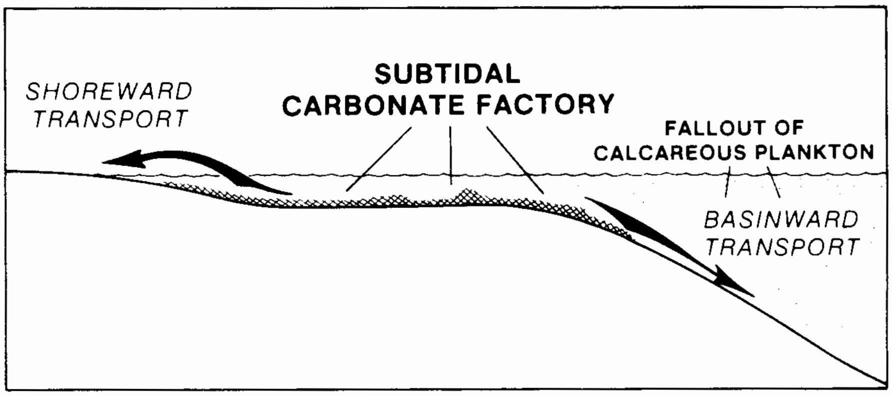


Figure 2
A sketch illustrating the main zones of carbonate accumulation, with most of the carbonate in water less than 30 metres deep.

ZONES OF CARBONATE ACCUMULATION

Because the precipitation of carbonate is easiest in warm, shallow seawater, most carbonate sedimentation takes place on continental shelves or banks in the tropics. Although most sediments produced in this 'carbonate factory' remain in the source area, some are transported basinward (Fig. 2). Thus, there are three different zones of accumulation: 1) the subtidal, open shelf and shelf margin, characterized by in-place accumulations of lime sands, lime muds and reefs; 2) the shoreline, where sediments are transported from the open shelf onto beaches and tidal flats; and 3) the slope and basin, where shelf-edge sediments are transported seaward, often by mass movements, and redeposited at depth. In the basins, especially in post-Jurassic time, the fallout of calcareous zooplankton and phytoplankton has also contributed significantly to carbonate sediments.

The characteristics of many carbonate depositional environments have been summarized and profusely illustrated in colour by different authors in the American Assoication of Petroleum Geologists Memoir 33 (Scholle *et al.*, 1983). The reader who wishes a detailed account of different carbonate sedimentary facies through time is referred to outstanding documentation by Wilson (1975).

As each of the three zones of accumulation have distinctive sedimentary environments and produce differing sedimentary facies, they will form a framework for the subsequent articles on carbonate facies models. The shoreline and slope-to-basin facies models are most like siliclastic facies models because sediments are transported from one area and deposited in another. At the other end of the spectrum reefs and reef-like deposits are the most unlike siliclastic facies as they are predominantly accumulations of biologically produced in-place carbonate.

REFERENCES

The reference list on this topic is relatively short because recently several excellent texts on carbonate sediments and facies have appeared. From these the reader can gain access easily to most of the pertinent literature on any specific aspect.

Bathurst, R.G.C., 1975. Carbonate sediments and their diagenesis. Developments in sedimentology No. 12. New York, Elsevier, 658 p.
This book is the most complete reference on the topic of carbonate deposition and diagenesis. Chapters 1 and 2 detail the petrography and occurrence of modern and ancient carbonate particles. Chapters 3 and 4 summarize several different and well-studied environments of carbonate deposition. The book does not cover ancient sedimentary rock sequences.

Flugel, E., 1982. Microfacies analysis of limestones. New York, Springer-Verlag, 633 p.
A thorough and well-illustrated documentation of microscopic aspects of carbonate rocks cast in terms of facies. An excellent reference work to compliment this series of articles.

Folk, R.L., and Robles, R., 1964. Carbonate sands of Isla Perez: Alacran Reef, Yucatan. Journal of Geology, v. 72, p. 255-292.
A classic study illustrating how two different skeletal organisms, corals and the codiacean alga *Halimeda*, break down under different conditions into specific grain sizes.

Ginsburg, R.N., and James, N.P., 1974. Holocene carbonate sediments of continental shelves. *In* Burke, C.A., and Drake, C.L., eds., The geology of continental margins. New York, Springer-Verlag, p. 137-157. A short article summarizing the sedimentology of eight different well-studied areas of carbonate sedimentation in the modern ocean.

Ginsburg, R.N., Lloyd, R.M., Stockman, K.W., and McCallum, J.S., 1963. Shallow water carbonate sediments. *In* Hill, M.N., ed., The sea, Vol. 3, p. 554-578.
The article illustrates how the architecture of modern marine carbonate skeletons governs the grain-size of the resultant sediments.

Ham, W.E., ed., 1962. Classification of carbonate rocks, a symposium. American Association of Petroleum Geologists, Memoir 1, 279 p.
This symposium contains several papers, notably those by W.E. Ham and L.C. Pray, M.W. Leighton and C. Pendexter, R.L. Folk, R.J. Dunham, which by attempting to classify sedimentary carbonates outline succinctly the important factors governing carbonate sedimentation.

Horowitz, A.S., and Potter, P.E., 1971. Introductory petrography of fossils. New York, Springer-Verlag, 302 p.
Chapter 2 is a concise introduction to carbonate sedimentology and the remainder of the book is devoted to the recognition of various skeletal particles in thin section.

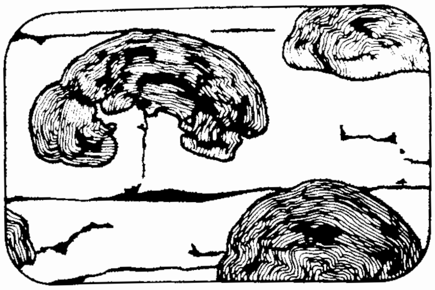
Miliman, J.D., 1974. Marine carbonates. New York, Springer-Verlag, 375 p.
This book is devoted wholly to modern carbonate sediments. The first half of the book is an exhaustive documentation of different carbonate particles; the second half is a discussion of modern environments of carbonate deposition - this book is most useful for the specialist.

Scholle, P.A., Bebout, D.G., and Moore, C.H., 1983. Carbonate depositional environments. American Association of Petroleum Geologists, Memoir 33, 708 p.
A superb coverage of all carbonate depositional environments, both modern and ancient, outlined in 12 sections with all illustrations in colour - the best overall coverage of this topic to be found anywhere.

Wilkinson, B.H., 1979. Biomineralization, paleoceanography and the evolution of calcareous marine organisms. Geology, v. 7, p. 524-528.
A short and useful article summarizing amongst other things the relative importance of various skeletal invertebrates as sediment producers through the Phanerozoic.

Wilson, J.L., 1975. Carbonate facies in geologic history. New York, Springer-Verlag, 471 p.
Chapters 1, 2, and 12 of this book are an excellent summary of the principles and stratigraphic aspects of carbonate sedimentation. The bulk of the text is a detailed review of carbonate sedimentary facies at different times in geologic history. This is the best single source book for ancient carbonates.

Wray, J.L., 1977. Calcareous algae. New York, Elsevier, 185 p.
The best single source for information or the sedimentology of various calcareous algae through geologic history.



Shallowing-Upward Sequences in Carbonates

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INTRODUCTION

Perhaps the most commonly encountered carbonates are laterally persistant, evenly bedded limestones and dolomites of apparent shallow water origin, as demonstrated by abundant fossil mud cracks and stromatolites. These deposits, which are usually found on the continents and in relatively undeformed portions of mountain belts, are not only important sources of paleontological and sedimentological information, but are also common host rocks for hydrocarbons and metallic ores (particularly lead and zinc). As such, it is critical that we be able to determine, as precisely as possible, the environment in which each of the interbedded sediments was deposited.

A quantum jump in our understanding of these deposits occured when modern carbonate tidal flats were examined in detail, notably by Robert Ginsburg and his colleagues in Florida and the Bahamas about 25 years ago. It was quickly realized that there was a host of sedimentary structures and textures on these flats that would allow a much more precise definition of environments of deposition than was possible before: these findings were quickly applied to fossil sequences (Fischer, 1964; Laporte, 1967; Aitken, 1966; Roehl, 1967). This application in turn generated two different lines of investigation: 1) description of other areas of modern tidal flat deposition, in particular the southern shore of the Persian Gulf where evaporites are common, and Shark Bay, Western Australia, where a great variety of modern stromatolites are forming; and 2) documentation of different styles of tidal flat deposits in the geologic record.

THE MODEL

Carbonate sediments characteristically accumulate at rates much greater than the rate of subsidence of the shelf or platform upon which they are deposited (Schlager, 1981). This is because carbonate sediments are produced mainly in the environment of deposition - especially in shallow water where conditions for the biological and physico-chemical fixation of carbonate are optimum. As a result, carbonate accumulations repeatedly build up to sea level and above, resulting in a characteristic sequence of deposits in which each unit is deposited in progressively shallower water. This *shallowing-upward sequence* commonly is repeated many times in a succession of shallow water deposits (Fig. 1).

Readers will recognize that such a shallowing upward sequence also may be termed a 'regressive sequence'. This term has led to much confusion in the past, because it has been used to describe deposits associated with a high rate of sediment production and accumulation under relatively static sea level - sea bottom conditions. I have, therefore, abandoned the term 'regressive' altogether in favor of a rock-descriptive term, albeit interpretive; the *shallowing-upward sequence.*

1) The Model as a Norm. The ideal carbonate shallowing-upward sequence comprises four units illustrated in Figure 2. The basal unit, which is generally thin, records the initial transgression over pre-existing deposits and so is commonly a high energy deposit. The bulk of the sequence, which may be of diverse lithologies, consists of normal marine carbonate, as discussed below. The upper part of the sequence consists of two units: the intertidal unit within the normal range of tides; the other a supratidal unit deposited in the area covered only by abnormal, windblown or storm tides. Each of these units exhibits the characteristic criteria of subaerial exposure.

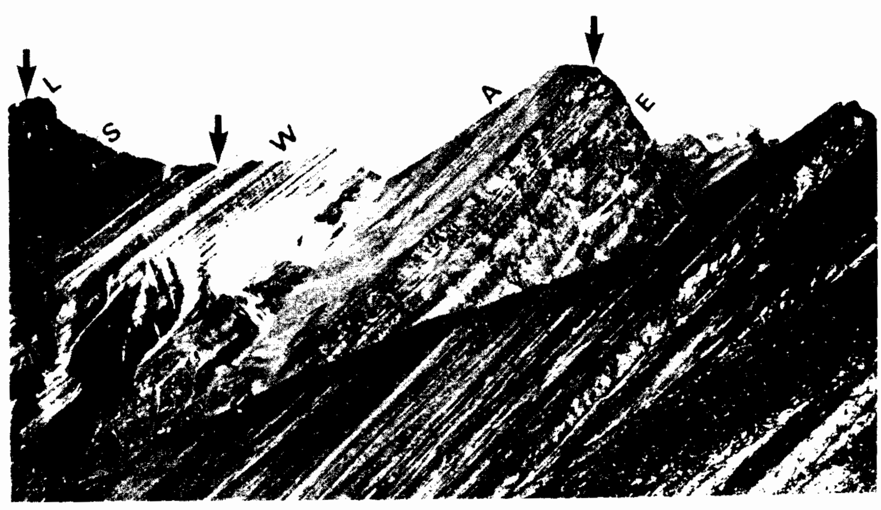

Figure 1
Bedded carbonates ranging in age from Middle to Late Cambrian near Fortress Lake, B.C. Arrows mark the top of large-scale shallowing-upward sequences (L - Lyell Fm., S - Sullivan Fm., W - Waterfowl Fm., A - Arctomys Fm., E - Eldon and Pica Fms.). Striping of the Waterfowl Fm. is caused by repetitive smaller scale shallowing-upward sequences between subtidal-intertidal limestones (dark) and supratidal dolomites (light). Photo courtesy J.D. Aitken.

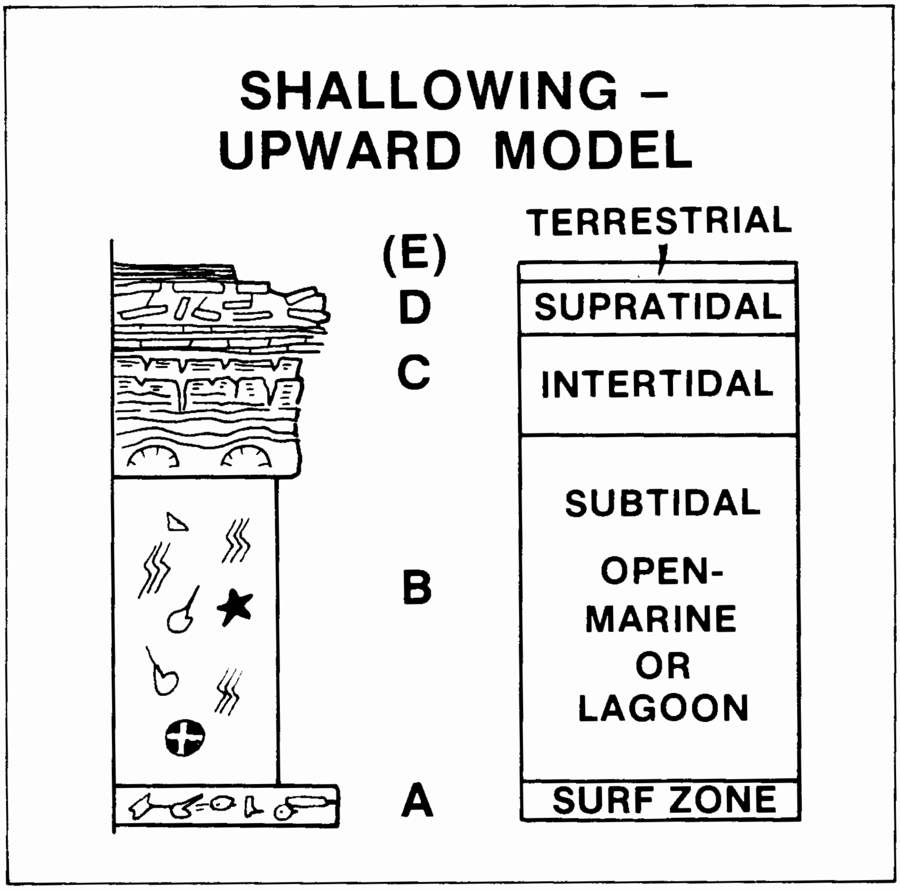

Figure 2
Five divisions of the shallowing-upward model for carbonates: A) lithoclast rich lime conglomerate or sand. B) fossiliferous limestone. C) stromatolitic, mud-cracked cryptalgal limestone or dolomite. D) well laminated dolomite or limestone, flat-pebble breccia. E) shale or calcrete, bracketed to emphasize that the unit is often missing - see text. Symbols used throughout are from Ginsburg (1975).

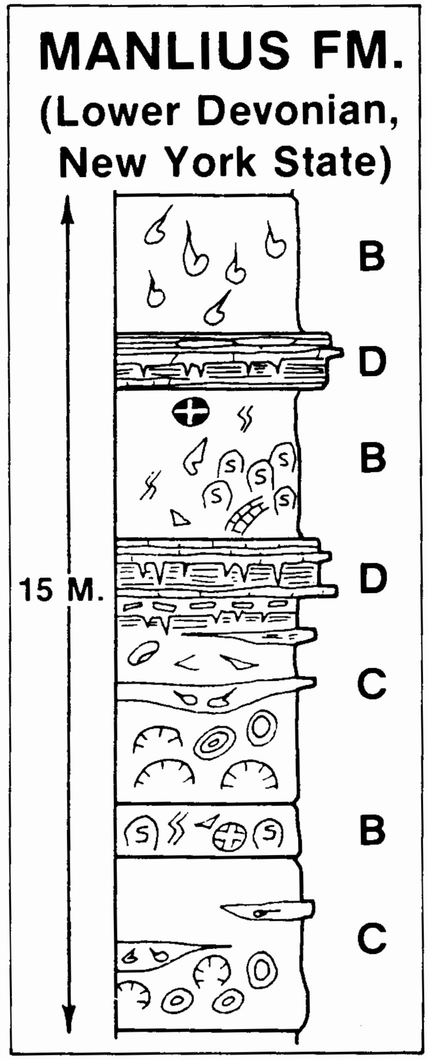

Figure 3
Actual sequence of several shallowing-upward sequences from the Manlius Fm., New York State (From Laporte, 1975).

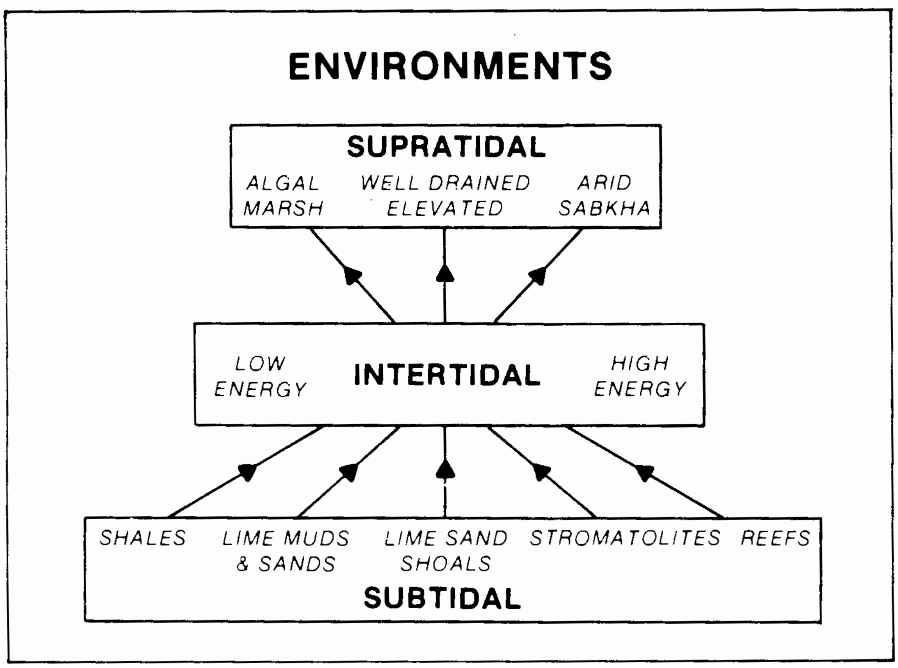

Figure 4
A flow diagram indicating the various possible environmental transitions present in a carbonate shallowing-upward sequence.

2) *The Model as a Predictor.* The thread that binds all such sequences together is the presence of the distinctive intertidal unit, which, once recognized, allows one to interpret the surrounding lithologies in some kind of logical sequence (Fig. 3), and thus predict what lithologies should occur in the rest of the succession under investigation.

First-order variation on the basic model revolves around the two main types of intertidal environment: 1) quiet, low-energy situations, commonly referred to as tidal flats, and 2) agitated, high-energy situations, or quite simply, beaches. Second-order variation involves the kind of subtidal units below and supratidal units above: the subtidal reflects the type of marine environment adjacent to the tidal flat and supratidal reflects the adjacent terrestrial environment, in particular the climate (Fig. 4).

For purposes of discussion I will begin with those sequences that contain low-energy intertidal units (tidal-flats) because they exhibit the greatest variety of distinctive features and consequently are well documented, both in modern and ancient settings. To place the observed features in context we should first examine modern carbonate tidal flats.

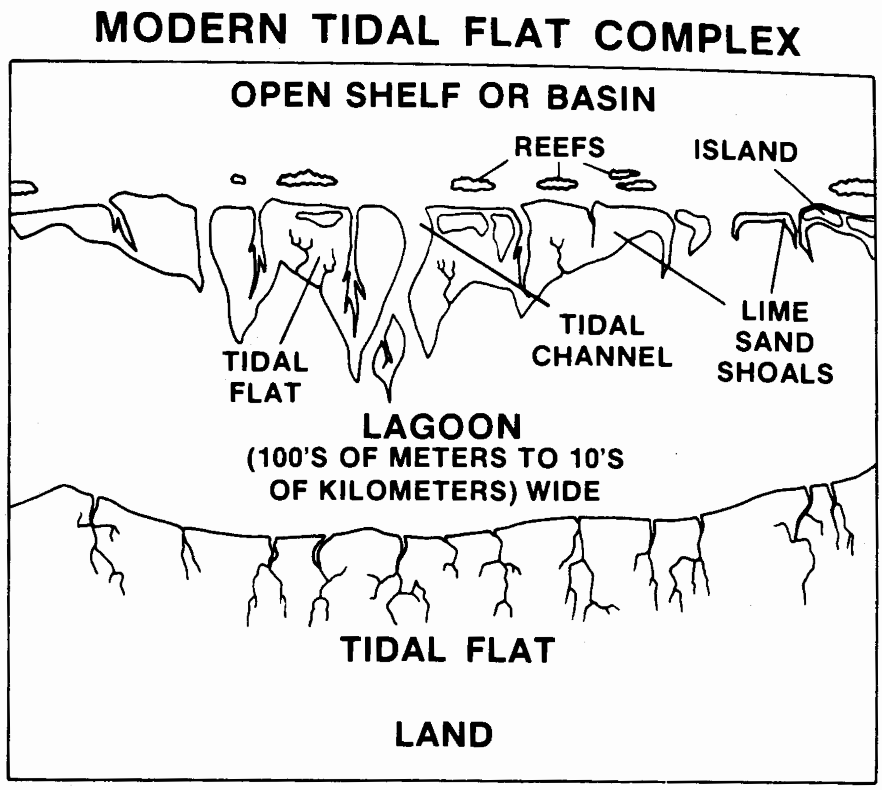

Figure 5
Plan view of the geometry of a modern tidal flat complex. Note that tidal flats can be present both adjacent to the land or in the lee of lime sand shoals.

SEQUENCES WITH A LOW-ENERGY INTERTIDAL UNIT

Modern Tidal Flats

The main elements of a modern carbonate tidal flat system as exemplified by the narrow shelf and embayments of Shark Bay, Western Australia, the southern coast of the Persian Gulf, and wide platform of the Bahama Banks are shown in Figure 5. The sedimentary features of these tidal flats are beautifully illustrated by Shinn (1983). A characteristic of most modern examples is that they occur in protected locations: protected that is from the open ocean waves and swells, yet still affected by tides and severe storms. This unique setting is commonly afforded by the presence of a semi-protective barrier composed of lime sand shoals, locally associated with reefs and/or islands. The barrier commonly is dissected by tidal channels through which flow high velocity tidal currents. A shallow muddy lagoon lies in the lee of this barrier. The lagoon may be enormous as in the case of the Bahamas, relatively narrow and elongate as in the Persian Gulf, or very small as in the pocket embayments of Shark Bay. In such an arrangement, tidal flats are present as: 1) small flats atop and on the lee side of the emergent sand shoals of the barrier, and 2) large flats along the shoreline of the shallow lagoon (Fig. 5). Thus tidal flats occur in association with two separate carbonate accumulations, high energy sand bodies and low energy lime muds. A third type, which is less common in modern situations, is the association with reefs, especially the interior of large reef complexes.

Intertidal Environments. The intertidal zone, especially along rocky coasts and beaches is commonly a gradual transition from sea to land without much noticeable variation. On wide, gradually sloping tidal flats this zone can be the familiar gradual transition or a complex area of many subenvironments. At one end of the spectrum the flats have few, very shallow, short tidal creeks (Fig. 6). At the other end of the spectrum the flats are dissected by many tidal creeks flanked by levees. Slight depressions between the creeks are occupied by tidal ponds (which fill and partially empty during each rise and fall of the tide) and the whole complex is fronted by small beach ridges or erosional steps (Fig. 6). Perhaps in this case it would be better to refer to the whole zone as the "pond and creek belt" because some of the areas are dry most of the time (levees and beaches) whereas others are continuously submerged (ponds and creeks). These complications have led some workers (e.g., Ginsburg and Hardie, 1975) to despair of conventional terms and instead to relate different zones to the per cent of time that they are exposed rather than to their position.

On some tidal flats where there are many tidal creeks and noticeable relief between levee and tidal pond (about 1 m), as in the Bahamas, the true intertidal zone which lies between the levee and pond may comprise only 60 to 70% of the intertidal environment. In other areas such as the Persian Gulf, where there are fewer creeks and less relief, almost the whole flat is truly intertidal. The most important point to grasp is that numerous environments may exist in very close proximity not only perpendicular to the shoreline but parallel to it as well, so that in the geologic record rapid, local lithological variations are to be expected, both vertically and laterally, rather than a smooth succession of progressively shallower environments.

The tidal flat wedge is built up of fine grained sediments brought onto the flats from the adjacent offshore marine zone by storms rather than by daily tides. Large storms such as hurricanes which flood the flat with sheets of water, white with suspended sediment, are particularly effective. Shinn *et al.* (1969) have suggested that the tidal flat is a river delta turned wrong-side out, with the sea as the "river" supplying sedi-

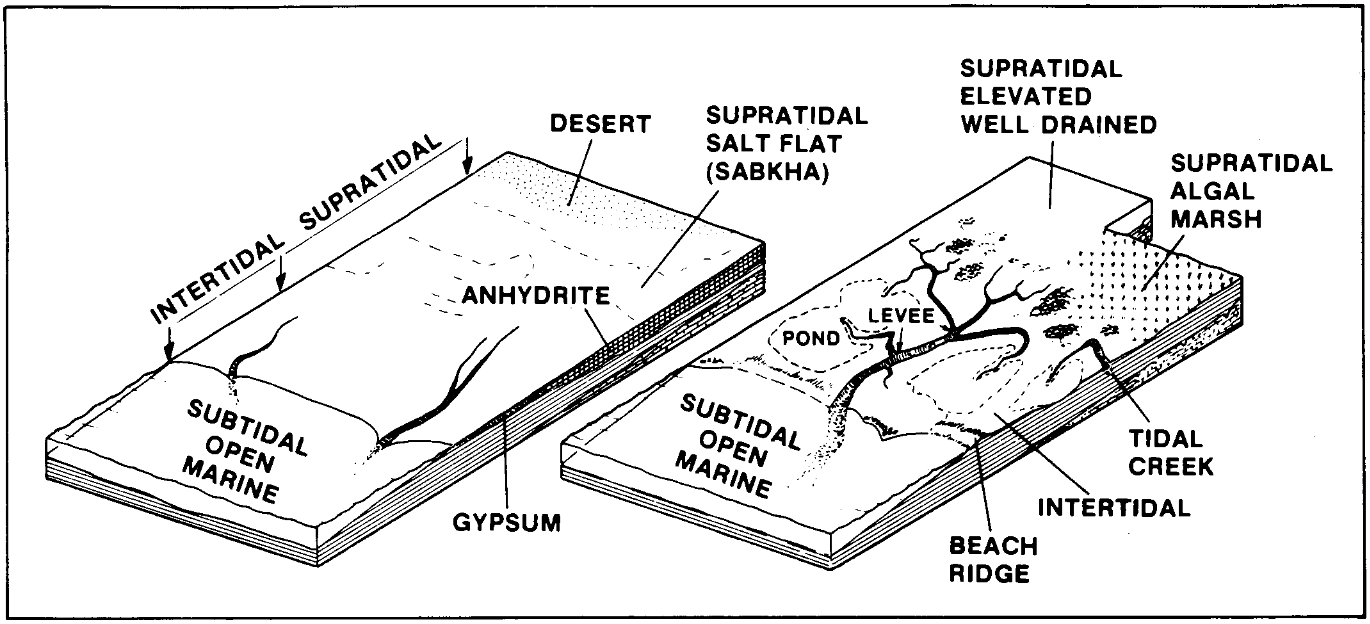

Figure 6
Block diagrams showing the major morphological elements of a tidal flat; left – a hypersaline tidal flat with few channels and bordering a very arid desert (similar to the modern Persian Gulf), right – a normal marine tidal flat with many channels and ponds and bordering an elevated well-drained area of low swamp algal marsh in a humid climate (similar to the modern Bahamas).

ment to the channeled flats as the "delta".

Sediments of the intertidal zone are characterized by three distinctive features, not found elsewhere: 1) algal mats, 2) irregular to even laminations (cryptalgal laminites), with fenestral porosity, and 3) desiccation features.

The algal mats are gelatinous to leathery sheets of blue-green algae which grow on top of the sediment surface. They are widely regarded as the signature of intertidal deposits. These mats are constructed solely or primarily by blue-green algae, which although photo-synthetic like other algal and higher plants, are prokaryotic and have much stronger affinities with bacteria than other eukaryotic forms. They are more correctly called cyanobacteria. It is probably more correct to refer to these mats as cyanobacterial mats or microbial mats (Bauld, 1981) but since the term algal mats is so entrenched in the literature I shall, for the time being, continue to use it in this paper. The mats may occur throughout the intertidal zone but their precise distribution is controlled by climate and the presence or absence of other organisms. The upper limit is controlled by climate; in arid areas they cannot grow above the high intertidal into the suprtidal zone, whereas in areas of high rainfall where the supratidal zone is moist or flooded for days at a time, mats are prolific. The lower limit is more variable and controlled by several factors. Garrett (1970) indicates that the main culprits are gastropods that eat algae. In areas of normal salinity, mats are prevented from developing below the middle intertidal zone because they are browsed by gastropods; in areas of hypersalinity (deadly for gastropods) mats grow down into the subtidal zone. In addition, algal mats will colonize only a temporarily or permanently stable bottom, and will not grow on shifting sand. Pratt (1982) points out that stromatolites are more common in post-Lower Ordovician rocks than generally realized, and that grazing by gastropods may be overemphasized as a limiting factor in their distribution. Rather it may be that substrate competition from various metazoans, together with increased rates of sedimentation during the Phanerozoic are more important controls.

Although the algal mats may themselves vanish with time, evidence of their presence during deposition remains because of the peculiar pores that they help to create, generally referred to as 'laminoid fenestrae'. These are irregular, elongate to mostly sub-horizontal sheet-like cavities (loferites or birds-eyes of some workers) with no obvious support and much larger than can be explained by grain packing. They are simply due to the fact that the mats are covered with sediment. The mats eventually rot away during burial, leaving voids as well as holes due to entrapped gas and mat shrinkage. Caution should be used when interpreting these structures, however, as similar features can be produced by submarine cementation of pellets, ooids and aggregate grains (Shinn, 1983).

Other sediments recording the presence of blue-green algal mats are the finely laminated carbonates (Fig. 7) ranging from stratiform and lightly crenulated to the familiar arched domes of stromatolites. These have been called cryptalgal (hidden, algal) laminations by Aitken (1967) in reference to the fact that the influence of algae in the rock-forming process is more commonly inferred than observed.

Lower Intertidal Zone. Much of the subtidal character remains evident in sediments from this part of the environment, and the deposits are commonly well burrowed and bioturbated. In hypersaline areas, however, the surface of the sediment is veneered with a thick algal mat, frequently broken into desiccation polygons. Beneath the mat, grains are

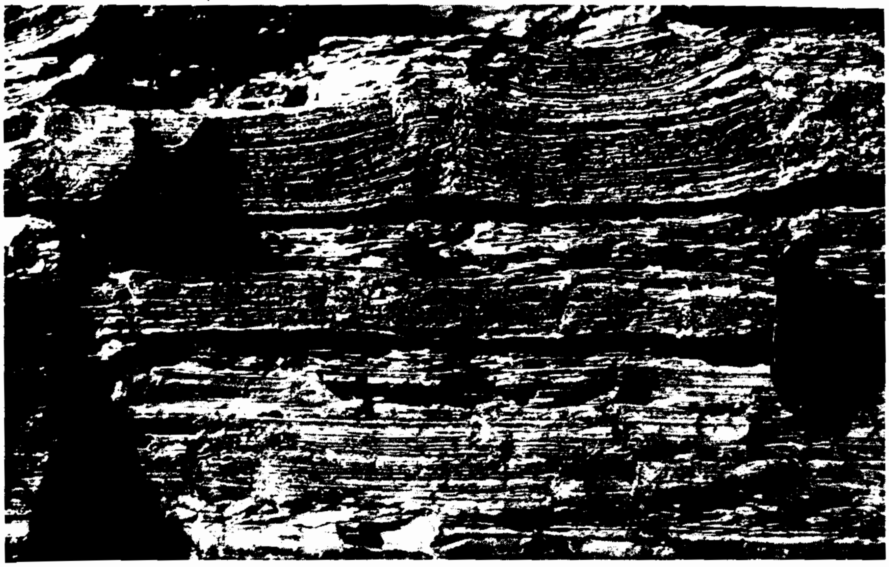

Figure 7
Cryptalgal laminites that have been mud-cracked. The intertidal unit of a shallowing-upward sequence in the Petit Jardin Fm. (Upper Cambrian) on the south shore of the Port-au-Port Peninsula, Nfld. (Photo courtesy R. Levesque).

Figure 8
A bedding plane of mud-cracked polygons with the edges of each polygon curled up, probably because the algal mats in the polygons shrivelled upon exposure and drying out. Near the top of a shallowing-upward sequence in the East Arm Fm., (Upper Cambrian), Bonne Bay, Nfld.

blackened due to reducing conditions and altered by boring algae to peloids of lime mud.

Tidal ponds and the creeks that drain them on hypersaline tidal flats support the most prolific growth of algal mats anywhere on the flat. The algal mat flourishes in water depths greater than those in the immediate offshore area because of relatively elevated salinities in the ponds. On tidal flats where the salinity is closer to normal, marine tidal ponds are populated by a restricted but prolific fauna of foraminifers and gastropods and the gastropods prevent the growth of algal mats. Similarly, if tidal creeks are common in such areas, the channels are devoid of mats but do contain concentrations of the pond fauna that are washed out during total exchange and which may accumulate as bars of skeletal lime sand. As the channels migrate these skeletal sands commonly form a basal lag deposit.

Middle and Upper Intertidal Zone. Sediments here are commonly light-grey to light-brown (oxidizing conditions), have good fenestral porosity (the variable growth of algal mats), are graded (episodic storm deposition) and are broken into desiccation polygons (prolonged exposure). There is generally good growth of algal mats throughout, and in the lower parts thick leathery mats are separated into desiccation polygons a few centimetres to a metre in diameter with cracks filled by lime mud in the lower parts (Fig. 8). In the central parts, thinner leathery mats have surfaces that are puffed up into blisters and convoluted into crenulated forms. In the upper parts, shriveled, crinkled and split mats are found. Bedding generally is irregular, especially in the upper zones, with mats alternating with graded storm layers.

In some settings, sediment in the upper intertidal zone dries out to form chips of lime mud while in others the sediment below the mats is lithified to a depth of as much as 10 cm.

Although sediments commonly are laminated throughout the intertidal environment, they are also riddled with small-scale tubules produced by insects and worms, and with larger tubes produced by crabs and other crustaceans.

Sediments also may be penetrated by the prolific shallow roots systems of salt tolerant plants.

Supratidal Environment. In all situations (including channel levees) this area is characterized by long periods of exposure. This is reflected by the lithification of storm deposited sediments in the form of surface crusts several centimetres thick, and which in turn are fractured into irregular polygons. These polygons may be pushed up by the force of crystallization (or by plant roots) to form 'teepees', or dislodged completely to form pavements of flat-pebble breccia. Clasts are commonly cemented on modern tidal flats by cryptocrystalline aragonite or calcite, and characteristically contain considerable (25 to 50%) fine crystalline dolomite.

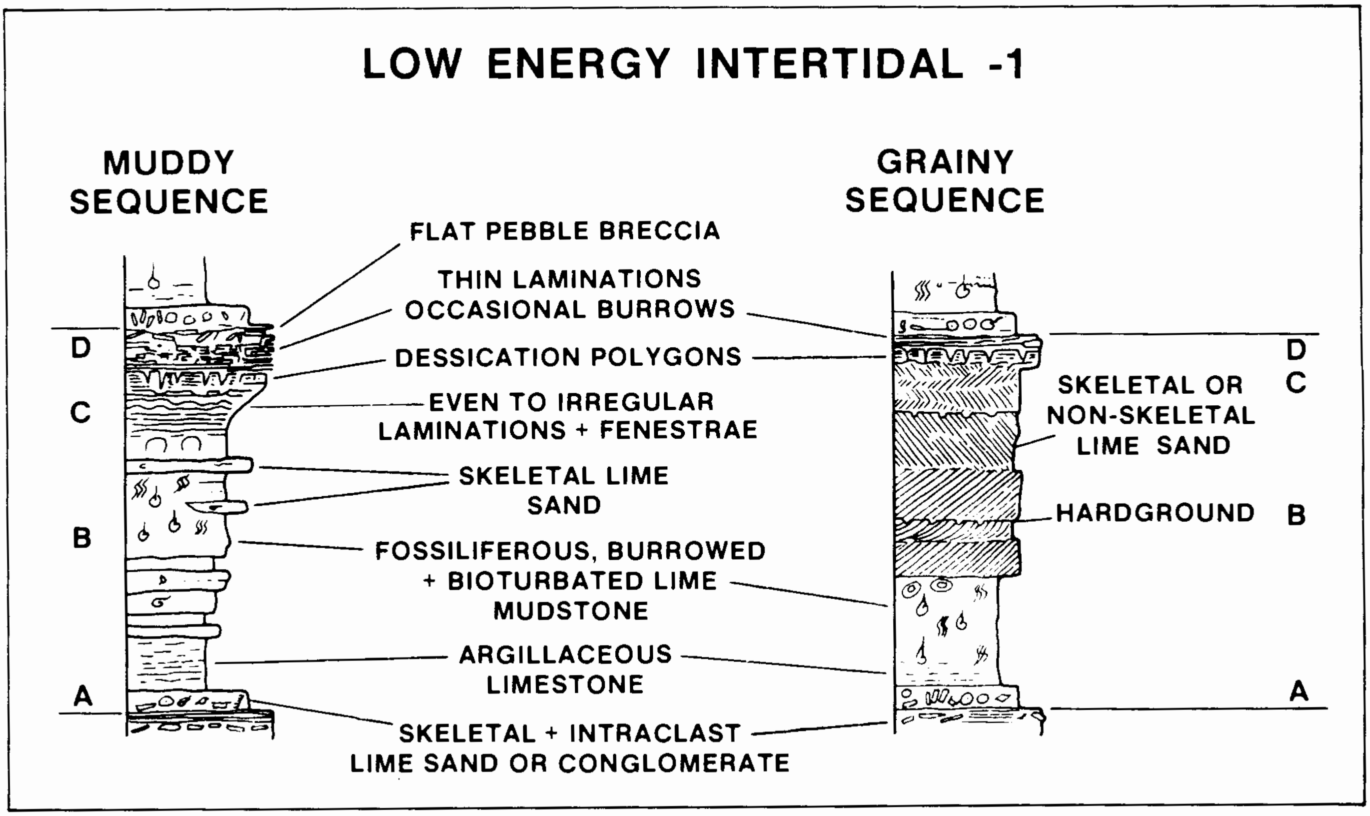

Figure 9
Two hypothetical sequences with a low energy tidal flat unit developed on a low energy subtidal unit (left) and a high energy lime sand unit (right).

If the creek levees in the intertidal zone have built up above normal high tide level, they consist of hard, finely to very finely laminated sediment, extremely regular and composed of alternating layers of sediment and thin algal mats with excellent fenestral porosity.

The landward parts of the supratidal zone may grade into various terrestrial environments, the end members of which are: 1) areas of elevated, pre-existing bedrock and no sedimentation in which the surface of the rock is characterized by intensive subaerial diagenesis, and the development of caliche (calcrete crusts); 2) areas of contemporaneous sedimentation which grade between: a) low-lying environments in regions of high rainfall occupied by algal marshes, b) low-lying environments in arid, desert regions, characterized by evaporite formation, and c) well-drained zones, often slightly elevated and with little deposition.

Algal marshes, flooded by fresh water during the rainy season, are an ideal environment for the growth of algal mats and these mats are periodically buried by layers of sediment swept in during particularly intense storms. The preserved record is therefore one of thick algal mats alternating with storm layers. With progressive aridity the supratidal zone dries out. If the chlorinity of the groundwaters remains constantly above 39 ‰ cementation, particularly by aragonite, is common. Cementation is most common if there is minor but consistent input of fresh water from inland to dilute the hypersaline groundwaters somewhat. If the chlorinity of the groundwaters remains constantly about 65 ‰ then authigenic evaporites precipitate within the sediment below ground level. In this setting (called a supratidal sabkha, or salt flat in the Middle East; see "Continental and Supratidal (Sabkha) Evaporites", this volume) dolomitizaiton is also common in the subsurface, saline brine pools occur at the surface, and terrigenous wind-blown sand is common in the sediment.

In relatively well-drained zones the supratidal environment is a deflation surface occasionally cut by the upper reaches of tidal creeks, sometimes damp from rising capillary waters and covered by a thin film of algal mat. Scoured and rippled sediment is common and clasts are sometimes encrusted with algae to form oncolites.

COMMON SEQUENCES WITH A LOW—ENERGY INTERTIDAL UNIT

Muddy and Grainy Sequences. These sequences developed either by progradation of the wide continental tidal flat or by shoaling the lime sand bodies that formed the barrier offshore (Fig. 9). The climate in the region of deposition was generally too wet or the ground-water table too low or diluted by fresh water to permit precipitation of evaporites.

The muddy sequences, those in which skeletal lime muds or muddy lime sands are the main subtidal unit, are well developed today in well-drained areas of Shark Bay where salinities are too high to permit development of a normal marine fauna as well as browsing of the algal mats by gastropods. Muddy sequences are also well developed in the tidal creek and pond belt of the Bahamas. These sequences are generally regarded as the 'classic' tidal flat sequences. The basal unit, if present, records the initial incursion of the sea onto land and as such is commonly coarse-grained, composed of clasts: all diagnostic of surf-zone deposition. The

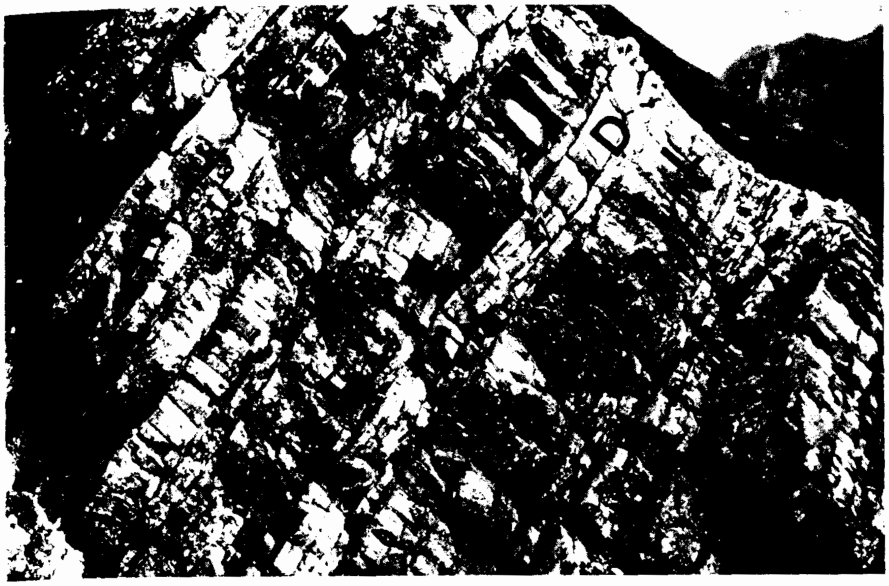

Figure 10
Shallowing-upward sequences comprising lower intertidal-subtidal limestones (L) overlain by supratidal dolomites (D – Cryptalgal laminites, sandy in part) in the Lyell Fm. at Takakkaw Falls, Yoho National Park, B.C. (Photo courtesy J.D. Aitken).

Figure 11
Numerous shallowing-upward sequences comprising thick subtidal oolite lime sands and thin intertidal-supratidal cryptalgal laminites with fenestrate porosity: Petite Jardin Fm., Port-au-Port Peninsula, Nfld.

subtidal unit is characteristically a bioturbated lime wackestone to packstone with a normal and diverse marine fauna, commonly containing stromatolites in deposits older than middle Paleozoic. In Precambrian and lower Paleozoic deposits the characteristic tidal flat features such as desiccation polygons, well-laminated sediments and fenestrae will occur at the base of the intertidal zones (Figs. 10 and 11). In deposits younger than middle Paleozoic, the prolific browsing and burrowing activity in the lower intertidal zone (unless the water mass was hypersaline) has homogenized the sediment, so that the signature of intertidal deposition is recorded only within the mid and upper intertidal sediments.

If the tidal flat was extensively channeled, the migration of channels back and forth may also have destroyed some of the subtidal character, forming instead a partial fining-upward sequence (much like that a river), with a basal skeletal lime sand.

Where fenestrae are present they show a zonation: horizontal to laminated in the lower intertidal environments (smooth mat), irregular and, in some cases, vertical in the middle and upper intertidal environments (pustular, shriveled and crinkled mats).

Desiccation polygons are most common near the top, apparently coincident with cementation. The supratidal zone is characterized by very evenly laminated deposits or flat pebble breccias.

Readers interested in the finer details of such sequences are referred to studies by Laporte (1967) and Fischer (1964), the latter outlining and documenting a similar facies sequence but in reverse order, forming a deepening-upward sequence.

A common early Paleozoic subtidal lithology in these sequences is alternating thin-bedded limestone and shale, forming ribbon to parted limestones. Sepkowski (1981) notes that flat-pebble conglomerates are conspicuous in these Cambrian and Ordovician sequences and suggests that they may be formed by early seafloor lithification of true carbonate followed by erosion and redeposition as storm deposits. Thus the presence of flat-pebble conglomerates alone need not indicate tidal flat deposition. Expansion of infauna in middle Ordovician time led to greater burrowing in the subtidal zone and so reduced the opportunity for early lithification in younger sediments.

Shoaling sequences also may be present off-shore from the low-energy tidal flat, on the lime-sand shoals. Here low energy tidal flats developed in the lee of the leading edge of the shoal once beach ridges were developed or currents had swept sand together to form islands. This will be reflected in the sequence as a sudden change from obvious high energy deposition to low energy intertidal deposition. The subtidal unit is generally well-sorted, oolitic, pelletoidal or skeletal lime sand (pelmatozoans are particularly common in the Paleozoic), with a few containing oncolites. Bedding is characteristically planar, with herringbone cross-laminations, large at the base and becoming smaller upwards, and individual bedding planes commonly covered with small-scale ripples. Early cementation is characteristic, and so deposits contain many intraclasts of cemented lime sand, and bored surfaces. Once the shoals, or parts of the shoals are inactive they may be burrowed and much of the original cross-bedding may be destroyed.

The intertidal to supratidal units are similar to those described above but are generally relatively thin. If the shoal is exposed for a long time caliche and soil profiles commonly develop, reflected by brown irregular laminations, breccias, and thin shale zones.

An excellent description of muddy and grainy sequences can be found in Demicco and Mitchell (1982).

Stromatolite and Reef Sequences. One common variation on the model is the development of shoaling-upward sequences in association with abundant stromatolites in the lower Paleozoic/-Precambrian and with reefs in the Phanerozoic in general.

In Shark Bay, Western Australia, where all environments are hypersaline and so stromatolites abound, the interrelationship between stromatolite morphology and environment has only recently been documented (Hoffman, 1976). In the intertidal zone columnar to club-shaped forms up to one metre high are found rimming headlands. In relatively high energy, exposed environments the relief of the columns is proportional to the intensity of wave action. The columnar forms grade laterally away from the headlands to the lower energy bights, where the stromatolites are more prolate and elongate oriented normal to the shoreline. In tidal pools digitate columnar structures abound.

These growth forms are the result of active sediment movement; algal mats only grow on stablized substrate, thus columns are nucleated upon pieces of rock, or cemented sediment; growth is localized there and does not occur on the surrounding shifting sands. Early lithification of the numerous superimposed layers of mat and sediment turns the structure into resistant limestone. Moving sand continuously scours the bases of the stromatolites. The mounds or pillars are largest in subtidal or lower intertidal environments and decrease in synoptic relief upwards, finally merging with stratiform mats in upper intertidal zones, above the zone of active sediment movement.

The resulting model sequence, summarized in Figure 12, is integrated from the Shark Bay example and the summary sequence of 200 or more shoaling sequences present in the Rocknest Formation of middle Precambrian age near Great Slave Lake (Hoffman, 1976). In the intertidal zone deposits reflect higher energy than normal, indicating a more exposed shoreline. These sediments underlie and surround the domal (Fig. 13) to columnar stromatolites, which in turn grade up into more stratiform stromatolites, and finally into very evenly bedded structures. The supratidal unit of this sequence will be characterized by both desiccation polygons and flat-pebble breccias as well as occurrences of delicate branching stromatolites (Fig. 14), formed in supratidal ponds. Care should be taken in delineating this sequence because stromatolites that are similar to those in the intertidal zone also occur in the subtidal (Playford and Cockbain, 1976).

Shallowing-upward sequences are also common as the last stage of sedimentation in large bioherms, as numerous successions within the large back-reef or lagoonal areas of reef complexes, and as 'caps' on widespread biostromes. In this type of sequence the shoaling upward is first reflected in the subtidal unit itself, generally as a transition from large massive hemispherical colonial metazoans of the reef facies, to the more delicate, stick-like forms that are common in the shallow protected

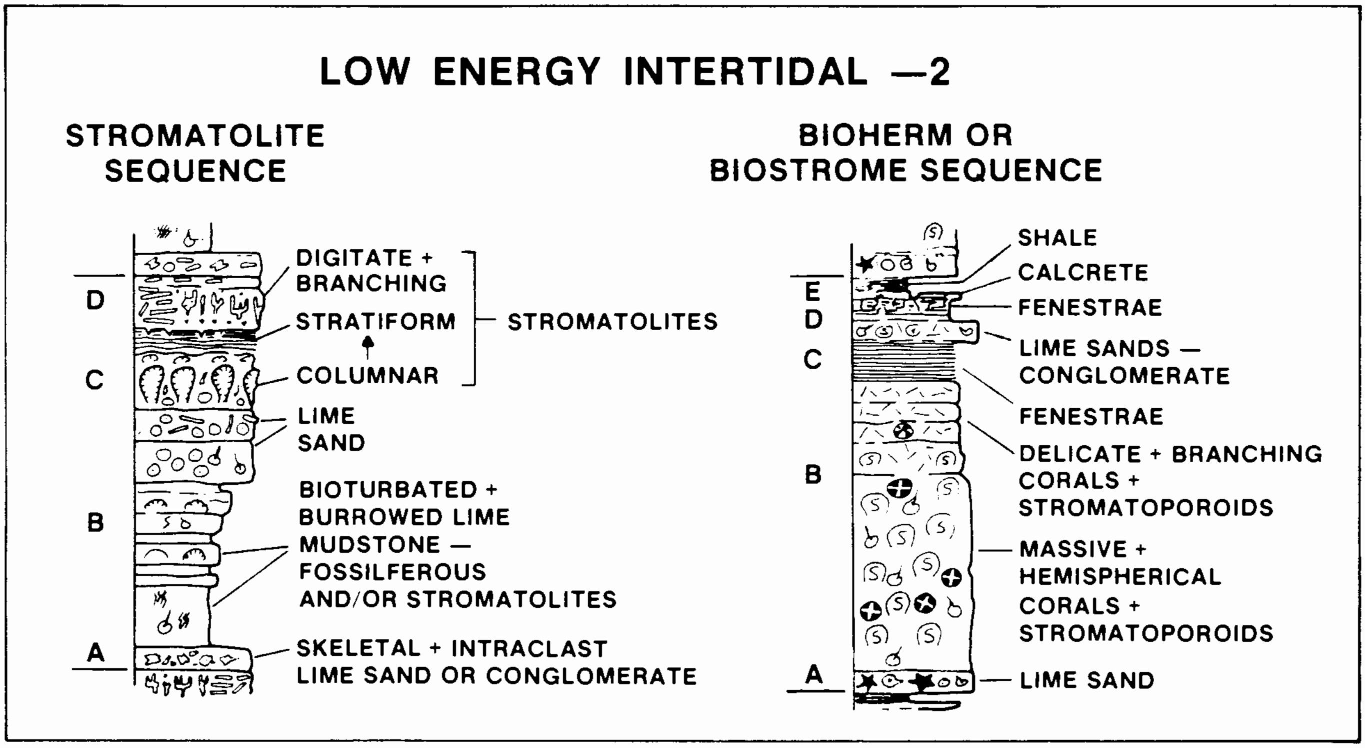

Figure 12
Two hypothetical sequences with a low energy intertidal unit developed in conjunction with stromatolites (left) and on top of a skeletal metazoan bioherm or biostrome (right).

Figure 13
A columnar to club-shaped stromatolite of Late Cambrian age from the Petite Jardin Formation, Western Newfoundland.

Figure 14
Digitate stromatolites from a shallowing-upward sequence of Late Cambrian age Western Newfoundland.

locations. These stick-like skeletons may be swept together on beaches at the edge of the tidal flat. As a result, the intertidal unit commonly contains a conglomerate within it, or at the base. The upper part of the sequence is otherwise similar to the others described. For a more detailed description of "reefy" sequences see studies by Havard and Oldershaw (1976), Read (1973), and Wong and Oldershaw (1980).

Carbonate-Evaporite Sequences (also see Kendall's paper in this volume). The other major variation on the model proposed at the beginning of this article is at the opposite end of the environmental spectrum, in the supratidal zone, in this case emergent in a very arid environment and flushed by hypersaline groundwaters. The hypersalinity of the groundwaters and attendant high evaporation results in the formation of authigenetic evaporites. This in turn raises the Mg^{++}/Ca^{++} ratio of the groundwaters and induces dolomitization of the sediment. The processes occur within the sediment, above the water table in the intertidal zone, and both above as well as below the water table in the supratidal zone. If the water compositions are barely within the field of gypsum precipitation, and there are fluctuations due to brackish flow of groundwater from the mainland, evaporites will occur in the form of isolated masses or crystals in the upper part of the sequence. If the groundwater compositions are continously well within the field of gypsum precipitation, growth of evaporite minerals takes place: 1) as a mush of gypsum crystals in the intertidal zone or as layers of anhydrite nodules, 2) as complex masses with a characteristic chickenwire texture, and 3) as layers contorted into enterolithic (intestine-like) shapes (Fig. 15). The important point, which is often ignored, is the growth of the evaporites within the sediment, as a diagenetic overprint on depositional facies of various environments. As evaporite growth is porphyroblastic, the host sediment commonly is displaced to intercrystalline areas and earlier fabrics are destroyed. Accompanying dolomitization is commonly intense with sediments of the intertidal and much of the subtidal zones affected.

Evaporites, however, are very soluble when exposed to percolating meteoric waters of low salinity and have a tendency to vanish from the record. Dissolution of the evaporites affects the sequences in several ways, but the most important is the formation of collapse breccias (Fig. 16). This collapse occurs when the evaporites dissolve leaving no support for the overlying sediments which subside into the void created by evaporite removal. Thus the top of the sequence is a breccia of marine limestone from the overlying sequence with a mixture of terrigenous sand, if a terrigenous facies capped the original sequence (Fig. 15). Isolated anhydrite crystals in lower parts of the sequence may be leached out, forming vugs which may be subsequently filled with quartz or chalcedony (usually length-slow). The dolomite, at least in the

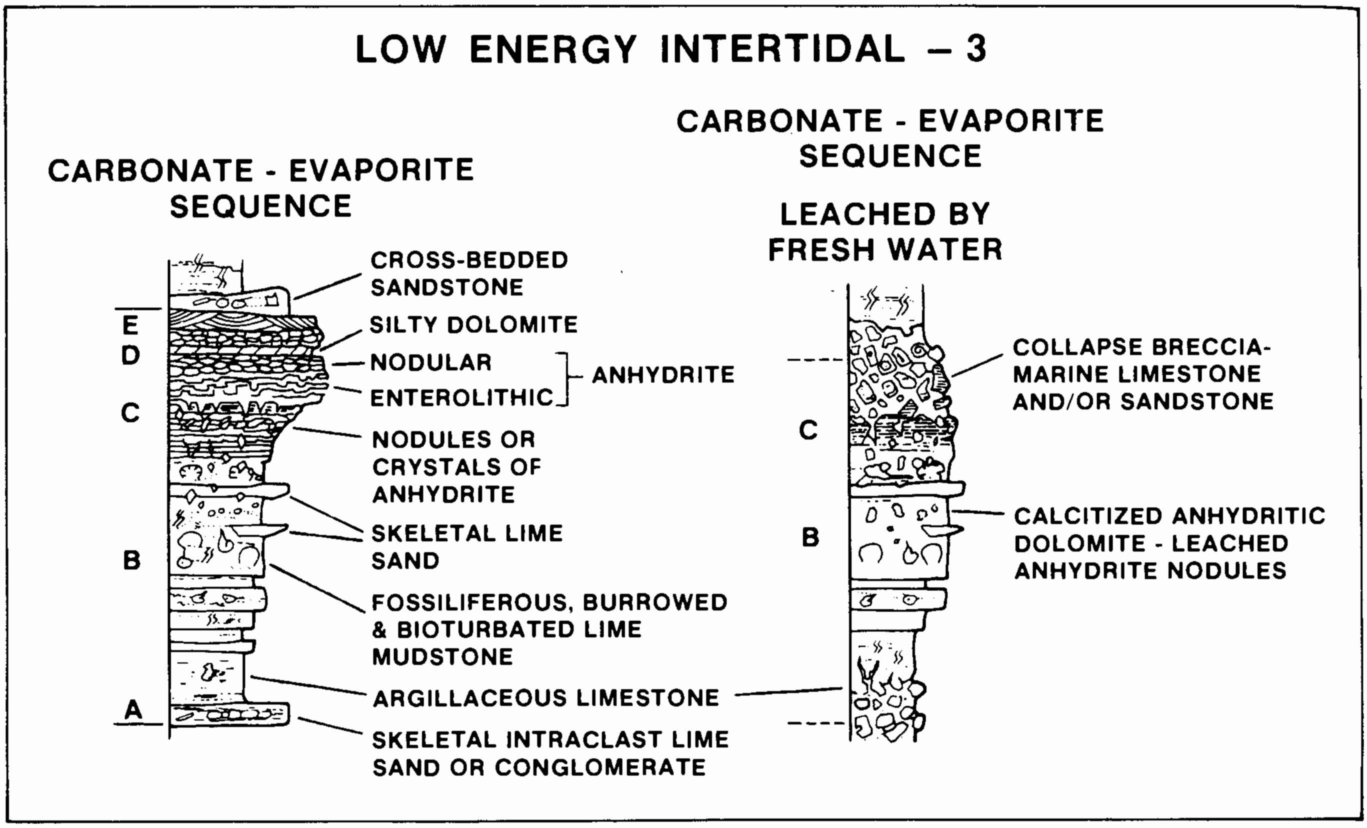

Figure 15
Two hypothetical sequences with a low-energy intertidal unit and a supratidal unit developed under arid conditions; on the right the evaporties have been dissolved by percolating fresh waters.

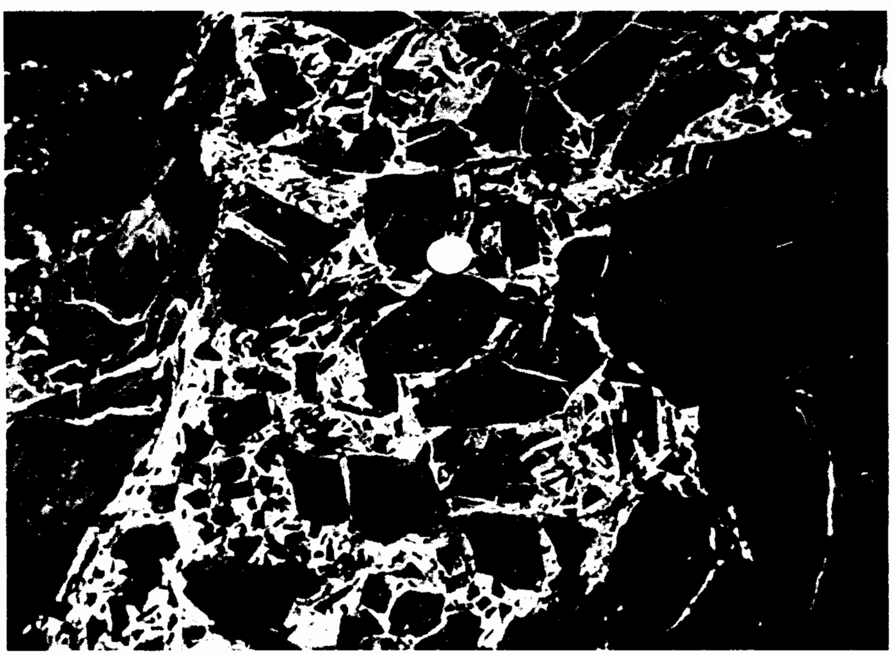

Figure 16
A collapse breccia of subtidal lime mudstone clasts in white calcite: caused by the solution of anhydrite at the top of a shallowing-upward sequence in the Shunda Fm. (Mississippian) at Cadomin, Alberta (Photo courtesy R.W. Macqueen).

upper part, is commonly altered to calcite, in the reverse of the dolomitization process (so-called "dedolomitization").

SEQUENCES WITH A HIGH ENERGY INTERTIDAL UNIT

In contrast to the low-energy intertidal (the tidal flat) the higher energy beach zone is not commonly recognized in the rock record. This may be partly because it resembles many subtidal grainstone deposits and hence is not obviously distinctive. Also, it is relatively narrow compared to the tidal flat, and has a lower preservation potential. Finally, the beach deposits lack the distinctive sedimentary features of the tidal flat. These very reasons illustrate the value of the concept of a shoaling-upward sequence as a guide. Once the potential for such a sequence is recognized in the geologic record, then one can concentrate on the search for subtle features that characterize beach deposition, which otherwise might go unnoticed.

Modern Carbonate Beaches

The sedimentology of carbonate beaches is nicely illustrated by Inden and Moore (1983). The beach is characterized by two zones: 1) the lower foreshore, usually below the zone of wave swash, and 2) the upper foreshore, the zone of wave swash. Sediments of the lower foreshore are coarse grained, poorly sorted, have a matrix of lime mud (if it is available), and are characterized by small and large-scale festoon cross-bedding, oriented parallel to the shoreline and generally attributed to longshore drift. The upper foreshore comprises thick-bedded, internally laminated, very well-sorted lime sands and gravels in planar cross-bedded accretionary beds that dip gently seaward (generally less than 15°). Sediments in the upper foreshore zone may have many open-space structures, the equivalent of the fenestrae of muddy intertidal sediment called keystone vugs (Dunham, 1969) or microcaverns (Purser, 1972). These are due to gas escape, and in the geological record are partly to completely filled with cement.

As on the tidal flat, periodic exposure of beach deposits leads to cementation and partial subaerial diagenesis. The textures thus created are difficult to recognize in the field but are important keys to recognizing the beach environment. The two most important of these diagenetic phenomena are beachrock and calcrete.

Shallowing-Upward Sequence With a High-Energy Intertidal Unit

The lower two units of this type of sequence are similar to those described in the preceding sections on sequences with low-energy intertidal units (Fig. 17). In this sequence, however, characteristic subtidal carbonates grade upward into coarse-grained lime sands with all the characteristics of the lower and upper foreshore described above (Fig. 9). The supratidal unit may be present in the form of a thin shale (soil), but more commonly the supratidal environment is represented not by a deposit but by intensive diagenesis of the upper unit (cementation, dissolution, calcrete formation and microkarst). This is in many ways similar to the diagenetic overprint of other facies by supratidal evaporite formation.

Beachrock is composed of seaward-dipping beds of lime sand and gravel that are generally cross-laminated and occur in the lower intertidal to middle intertidal environment. It is formed by the precipitation of carbonate cement out of seawater or mixed seawater and rainwater. The beds of limestone may be up to one metre thick, are commonly jointed at right angles to the beach, and

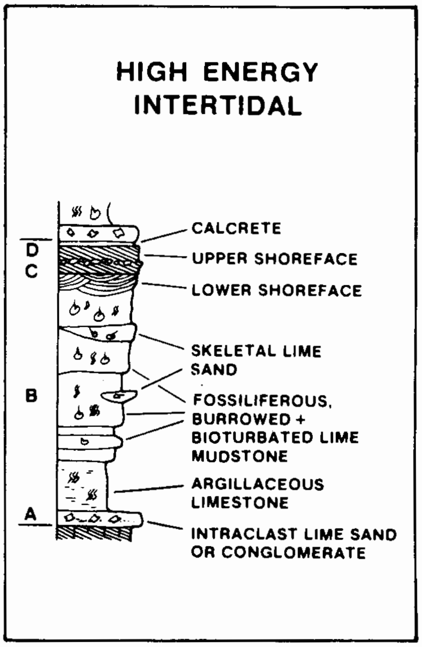

Figure 17
A hypothetical sequence with a high-energy intertidal unit: a beach, developed, in this case adjacent to a low energy subtidal environment.

are encrusted and/or bored by numerous intertidal organisms. Lithification disappears seaward and rarely extends higher than the intertidal zone. The partly cemented beds may be broken up and redeposited as conglomerates, made up of cemented sand clasts. In the upper parts of the intertidal zone cementation takes place in intergranular voids partly filled with air: the cements, as a result, are often stalactitic (more extensively developed on the undersides of grains).

If exposed for long periods of time and if located in an environment where there is at least periodic rainfall, the lime sands will begin to undergo subaerial diagenesis (see Bathurst, 1975, for an extended discussion of subaerial diagenesis). In addition the upper metre or so of such subaerially exposed deposits develop calcrete or caliche horizons which have many features that closely resemble those produced by laminar to laterally-linked stromatolites and oncolites. These features are discussed in detail by James (1972) and Read (1976).

The supratidal unit in these sequences may be any of the ones described above, although calcrete (caliche) is very common. Beaches may act as small barriers protecting supratidal ponds and flats so that the cap in such sequences will be thin beds of lime mud (often dolomitized) with all of the associated supratidal features. One variation not found elsewhere occurs where the high energy surf zone of the overlying sequence erodes the top of the sequence down to the cemented portions, resulting in truncation layers or hardgrounds that separate sequences.

CYCLICITY

The shallowing-upward facies model is demonstrably one of the most useful concepts a sedimentologist can have when working with platform carbonates. In the rock record these sequences occur on two scales; *small-scale*, or those less than a metre to a few 10s of metres in thickness at most, and *large-scale* or those many tens to hundreds of metres in thickness, and often likened to the Grand Cycles of Aitken (1967). In spite of the model's obvious utility, the precise mechanisms by which numerous sequences are generated remains obscure. Since the first edition of this paper (James, 1979) attention has shifted from the details of the model itself to the possible mechanisms of this accretion.

On the surface the explanation seems simple enough. Because the rate of carbonate deposition exceeds the rate of platform subsidence or sea level rise, sediments will rapidly accrete to sea level. This is fine for a single shallowing-upward sequence, but how are repeated sequences formed and how do they exist over vast carbonate platforms?

Small-Scale Sequences

Current thinking on how repeated small-scale sequences are formed is succinctly summarized by Wilkinson (1982), who points out that there are currently two end-member models, both of which generate virtually identical sequences (Fig. 18).

The *eustatic model* is one in which the rate of carbonate sedimentation is constant but the rate of subsidence or the absolute position of sea level are not constant and change in a non-uniform or periodic fashion. During periods of stability or slowly rising sea level (Figs. 18 and 2) the whole sequence progrades out across the shelf yielding a typical shallowing-upward or regressive sequence. This pattern is interrupted by a sudden and rapid sea level rise, flooding the platform (Fig. 18-3) and resulting in a short period of arrested- or

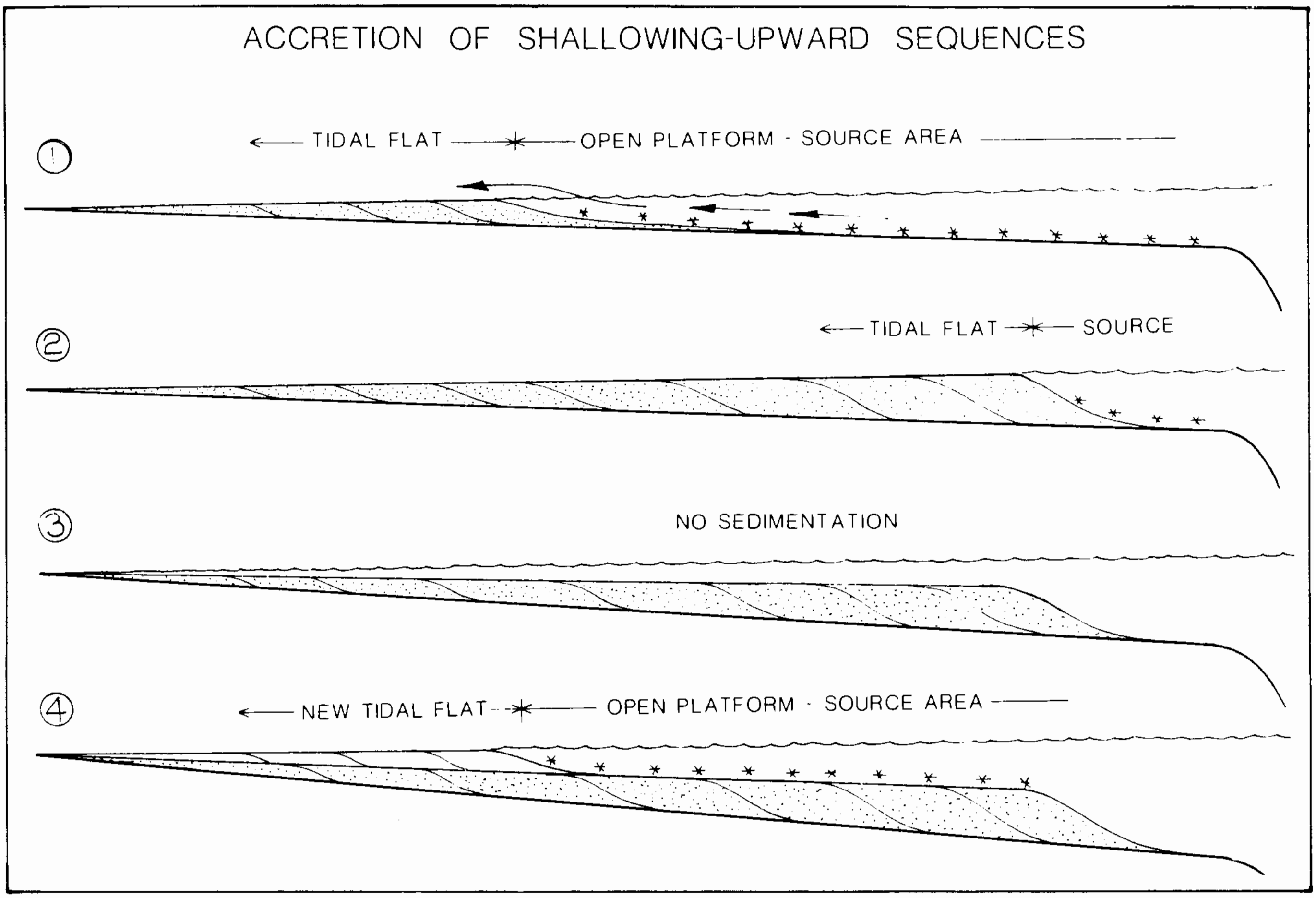

Figure 18
A sketch illustrating how two shallowing-upward sequences can be produced by progradation of a tidal flat wedge. These general conditions apply in the case of both eustatic and autocyclic models.

non-deposition. Sea level remains relatively stationary in this new position for a time (Fig. 18-4) and progradation begins again with a new shallowing-upward sequence forming over the old one. One variant of this model, calling for sudden platform-wide shifts in sea level has been formalized as "Punctuated Aggradational Cycles" by Anderson and Goodwin (1980).

While this model may be attractive as an explanation for large-scale shallowing-upward sequences it is less compelling for small-scale cycles, because each cycle must record either a sudden eustatic change in sea level or a tectonic event. If it is the cause of small-scale cycles then the problem becomes one of small-scale cyclicity on a global scale (Schwarzacher and Fischer, 1982) which is beyond the scope of this paper.

In the alternative *autocyclic model* the control is intrinsic and lies in the rate of carbonate sedimentation as controlled by source area (Fig. 18). The model was first proposed by Ginsburg (1971) and has been used for many years as a model in field seminars at the University of Miami. Similar schemes have subsequenly been suggested by Matti and McKee (1976), Mossop (1979) and Wilkinson (1982).

As in the eustatic model, sedimentation is envisaged as taking place on a gently inclined shelf under conditions of a gradually subsiding shelf or slowly rising sea level or some combination of both (Fig. 18-1). The source area of the sediments for the prograding wedge is the large subtidal area (Fig. 18-2), which is gradually reduced in size with seaward progradation. Eventually the situation is reached in which the source area is too small or too deep to provide sediments for the prograding wedge (Fig. 18-3), and so sedimentation stops. Relative sea level will continue to rise, however, and soon (Fig. 18-4) the whole platform will once again be subtidal and deep enough for sediment production and the cycle will begin again.

The second and related problem concerns precisely how tidal flats prograded over the vast areas of epeiric or epicontinental seas. The stratigraphic record illustrates that two situations are common. In one, individual component lithologies of remarkably uniform thickness occur over wide geographical areas. For these sequences Ginsburg (1982) has proposed either continuously prograding tidal-flat wedges which would create time transgressive units and leave large areas exposed to prolonged meteoric diagenesis, or repeated deposition of thin areally-extensive single-event units by wind-driven "tides".

In other examples individual lithologies cannot be correlated between adjacent wells only kilometres apart (e.g., Wong and Oldershaw, 1980) or transitions from supratidal to intertidal to subtidal facies can be walked out in a

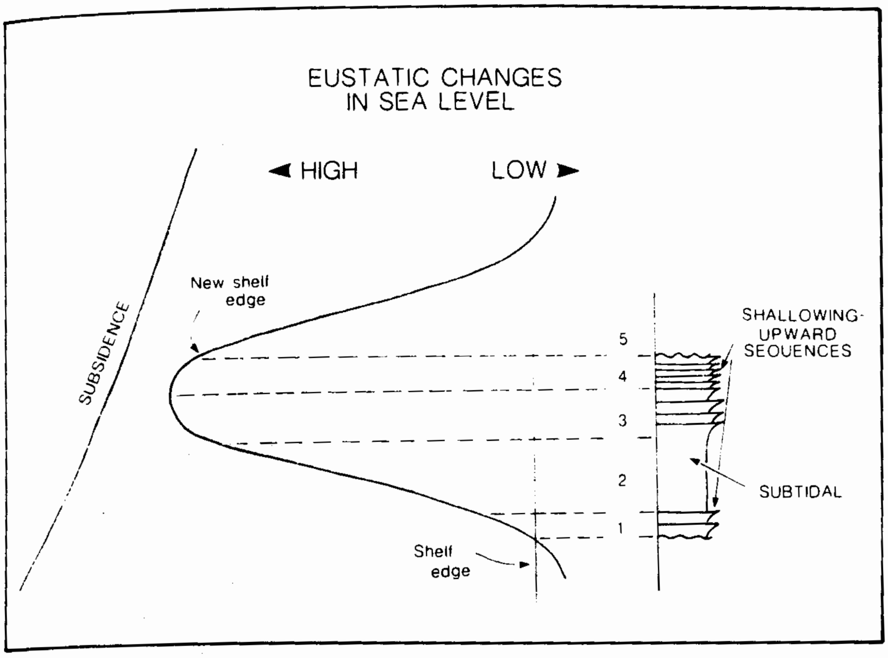

Figure 19
A diagram illustrating how a large-scale shallowing-upward sequence is produced under conditions of slow platform subsidence and a uniform rise and fall in eustatic sea level.

single bed over distances of only a few kilometres (e.g., Pratt and James, in press). The most appealing explanations for these carbonates is deposition on a platform dotted by a mosaic of exposed banks or islands separated by subtidal areas, with the whole complex shifting both laterally and vertically in response to hydrodynamic conditions through time.

This "island" model has an additional attraction in that a complex facies mosaic of numerous shallowing-upward sequences can be formed everywhere on a subsiding shelf at the same time, and is thus equally applicable to sequences formed under conditions of relatively uniform sea level or under conditions of sporadically changing sea level.

Large-Scale Sequences

These larger-scale packages, the upper parts of which are characterized by many small-scale sequences, are more complex, often involving an interplay of eustatic sea-level fluctuaitons, tectonics and platform geometry (see Aitken, 1978).

An example of how one such large-scale sequence might be developed is outlined in Figure 19. In this case, I have used the concepts formulated by the Exxon Seismic Stratigraphy Group (Vail pers. commun., 1984). Here subsidence is viewed as constant, with the rise and fall of sea level more or less symmetrical.

(1) As sea level begins to rise slowly over the shelf edge the platform is flooded. During this initial stage carbonate accretion can outpace relative sea level rise and shallowing-upward sequences develop.

(2) During the prolonged period of relatively rapid sea level rise platform subsidence is ongoing and the tendency is to maintain deep water over the shelf. Depending on the rates of each, either subtidal conditions may develop, or if both are slow a few thick shallowing-upward sequences may form.

(3) The rate of sea level rise slows, and shallowing-upward sequences develop easily.

(4) Sea level falls slowly, but subsidence is also continuing, so the net effect is a still stand or close to it. Numerous, thin, shallowing-upward sequences develop with long periods of subaerial diagenesis between.

(5) Sea level begins to fall rapidly, outpacing subsidence and quickly dropping below the shelf edge, which is now higher because of the intervening carbonate accretion; the whole platform is exposed.

The result of this sequence of events is a large-scale shallowing-upward sequence, possibly with a few cycles at the base but mainly a lower half of subtidal sediments and an upper half of numerous shallowing-upward sequences. Obviously this is one sequence formed in response to a given set of variables but other sequences can be easily generated using similar principles. What is important is that a large-scale asymmetrical cycle can be produced by a uniform rise and fall of sea level.

SUMMARY

In the past there has been a natural tendency to use obvious sedimentary structures (e.g., mud cracks, stromatolites) to infer that parts of a carbonate sedimentary sequence had been periodically exposed. Individual structures, however, often have counterparts in other sedimentary environments (e.g., syneresis cracks, subtidal stromatolites) resulting, in many cases, in questionable paleoenvironmental interpretations. With all the data now available on carbonate strandline deposition we can frequently use what have become natural associations of sedimentary features in a vertical succession and define, with precision, specific strandline facies and their interrelationships.

While this is true for low-energy shoreline sequences, it is much less so for high-energy shoreline sequences. To bring all aspects of this type of facies model to comparable levels of understanding much more data is needed on exposed or high-energy intertidal environments, not from the modern, but from the rock record. In addition, the time is ripe to test whether or not the diagenetic features which result from periodic subaerial exposure (cementation, microkarst, calcrete) can be commonly recognized in ancient sequences.

In conclusion, the shallowing-upward sequence is one of the commonest, and with the wealth of sedimentary features, one of the easiest models to apply to the carbonate facies spectrum. As a *norm* this model is constructed from a synthesis of over 40 well-documented fossil examples, but our understanding of the meaning of most diagnositc sedimentary features comes from half a dozen modern settings. Carbonate platform deposition, being what it is, a rapidly accreting system; once a shallowing-upward sequence has been recognized,

one can *predict* that other similar sequences will be present vertically. If the sequences recur in an orderly fashion (are cyclic) then similar packages should be present laterally as well. Alternatively, if parts of the model are stacked in a less regular fashion, reflecting deposition as a complex of islands and banks on an open shelf, then specific laterally equivalent facies are less predictable. Recognition of the style of the model, which is in turn dependent upon the overall climatic (humid versus arid) and oceanographic (normal versus hypersaline) conditions, then allows the model to be both a *basis for physical and chemical interpretation* as well as a *guide* to the interpretation of other shallow water carbonates of the platform succession.

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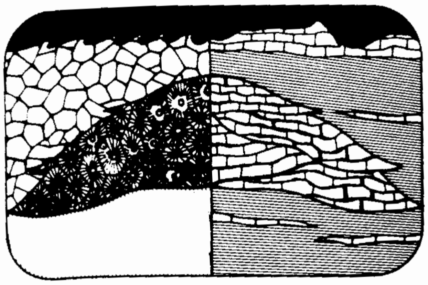

Reefs

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INTRODUCTION

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

Because they are built by organisms, fossil reefs (Fig. 1) are storehouses of paleontological information and modern reefs are natural laboratories for the study of benthic marine ecology. Also, fossil reefs buried in the subsurface contain a disproportionately large amount of our oil and gas reserves compared to other types of sedimentary deposits. For these reasons, reefs have been studied in detail by paleontologists and sedimentologists, perhaps more intensely than any other single sedimentary deposit, yet from two very different viewpoints. This paper is an integration of these two viewpoints. I shall concentrate less on the familiar trinity of back-reef, reef, and fore-reef, but more on the complex facies of the reef proper.

Since the first edition of Facies Models, there has been much new information on both the sedimentology and paleontology of reefs. The model itself has been presented elsewhere (James, 1983) and amplified using numerous examples from the modern and fossil record. In this present version the model remains unchanged but many of the underlying concepts and implications that flow from it have been revised and/or enlarged.

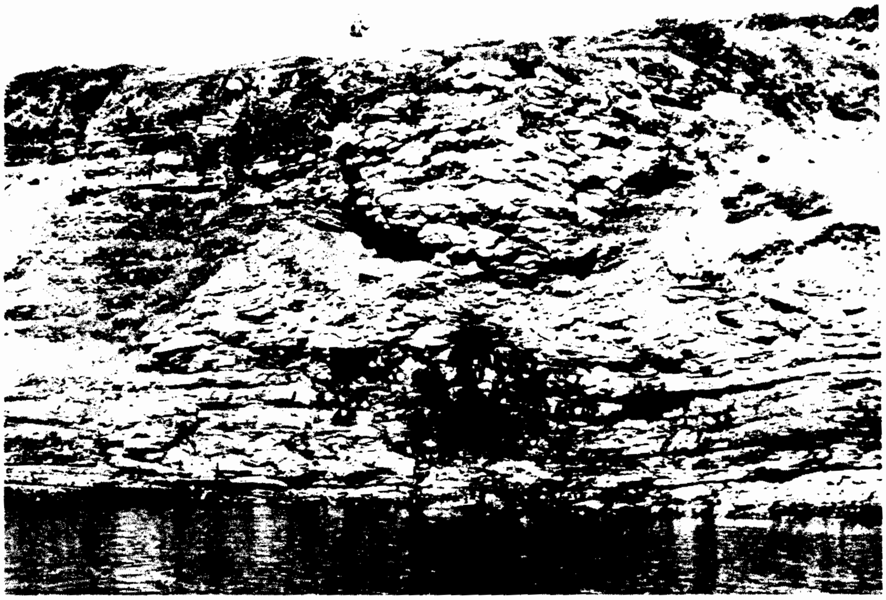


Figure 1
A patch reef of Lower Cambrian age exposed in sea cliffs along the northern shore of the Strait of Belle Isle, Southern Labrador.

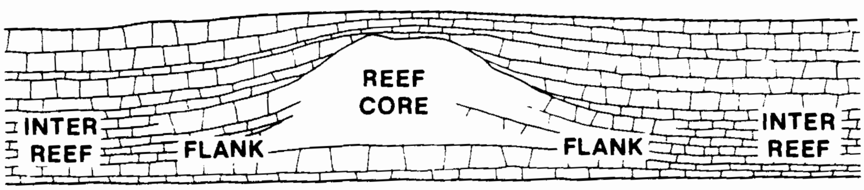


Figure 2
A sketch illustrating the three major reef facies in cross-section.

THE ORGANISM-SEDIMENT MOSAIC

Reefs can generally be subdivided into three facies (Fig. 2).

1) *Reef-core facies* - massive, unbedded, frequently nodular and lenticular carbonate comprising skeletons of reef-building organisms and a matrix of lime mud.

2) *Reef-flank facies* - bedded lime conglomerates and lime sands of reef-derived material, dipping and thinning away from the core.

3) *Inter-reef facies* - normal shallow-water, subtidal limestone, unrelated to reef formation, or fine-grained siliciclastic sediments.

A useful, non-generic term for such a structure is "bioherm" - for discussion of this and other reef terminology the interested reader is referred to papers by Dunham (1970), Heckel (1974), Longman (1981), and James (1983).

Reef facies are best differentiated on the basis of several independent criteria including: 1) the relationship between, and relative abundance of large skeletons and sediments, i.e., the type of reef limestone, 2) the diversity of reef-building species, and 3) the growth form of the reef builders.

Types of Reef Limestone

The present state of any thriving reef is a delicate balance between the upward

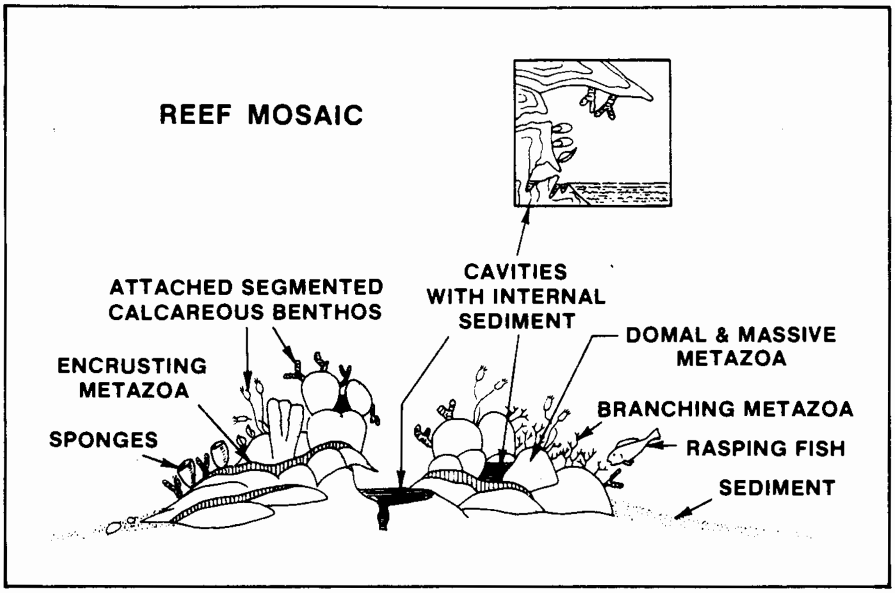


Figure 3
A sketch illustrating the different aspects of the organism/sediment mosaic that comprises a reef.

growth of large skeletal metazoans, the continuing destruction of these same organisms by a host of rasping, boring, and grazing organisms, and the prolific sediment production by rapidly growing, shortlived, attached calcareous benthos (Fig. 3).

The large skeletal metazoans (e.g., corals) generally remain in place after death, except when they are so weakened by bio-eroders that they are toppled by storms. The irregular shape and growth habit of these reef-builders result in the formation of roofed-over cavities inside the reef that may be inhabited by smaller, attached calcareous benthos. These cavities may also be partly or completely filled with fine-grained "internal" sediment. Encrusting organisms grow over dead surfaces and aid in stabilizing the structure. Branching reef-builders frequently remain in place, but just as commonly are fragmented into sticks and rods by storms to form skeletal conglomerates around the reef.

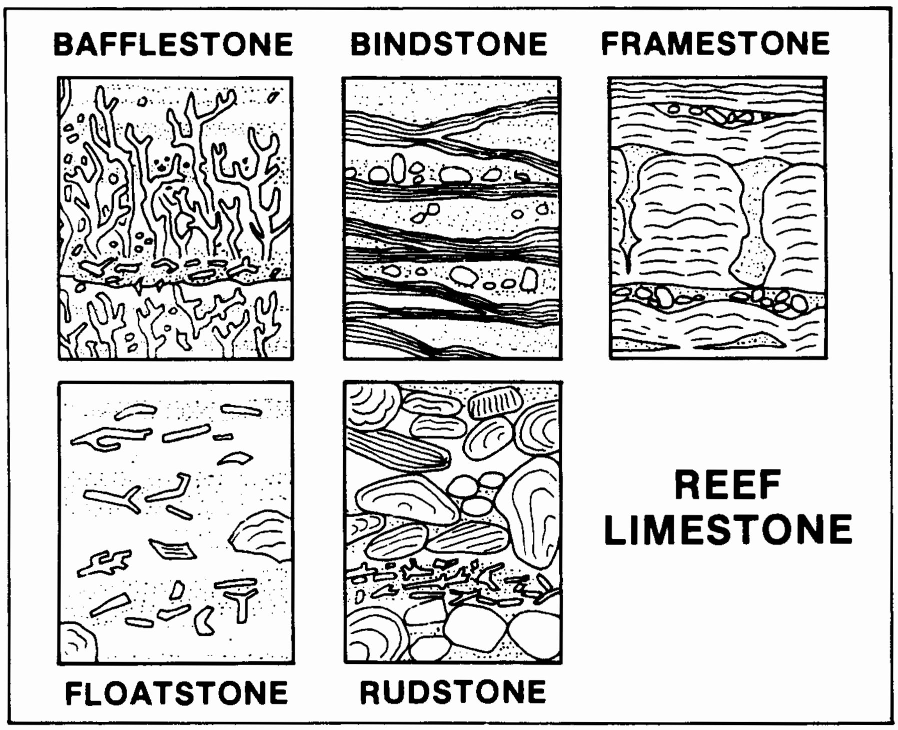


Figure 4
An interpretative sketch of the different types of reef limestone recognized by Embry and Klovan (1971). Autochthonous reef limestones are in the upper row, while allochthonous reef sediments are in the lower row. .

Most reef sediment is produced by the post-mortem disintegration of organisms that are segmented (crinoids, calcareous green algae) or non-segmented (bivalves, brachiopods, foraminifers). These organisms grow in the many nooks and crannies between larger skeletal metazoa. The remainder of the sediment is produced by various taxa that erode the reef: boring organisms (worms, sponges, bivalves) produce lime mud and rasping organisms that graze the surface of the reef (echinoids, fish) produce copious quantities of carbonate sand, and silt. This material is deposited around the reefs as an apron of sediment, and it also filters into the growth cavities to form internal sediment, which is characteristically geopetal.

Many different classifications have been proposed for the resulting reef carbonates, but the most descriptive and widely accepted is a modification of Dunham's (1970) classification of lime sand mud-rocks proposed by Embry and Klovan (1971) (Fig. 4). They recognize two kinds of reef limestone, allochthonous and autochthonous. The allochthonous limestones are the same as the finer grained sediments, but with two categories added to encompass large particles. If more than 10% of the particles in the rock are larger than 2 mm and they are matrix supported it is a *Floatstone*; if the rock is clast supported it is a *Rudstone*. The autochthonous limestones are more interpretative; *Framestones* contain in-place, massive fossils that formed the supporting framework; *Bindstones* contain in-place, tabular, or lamellar fossils that encrusted or bound the sediment together during deposition; *Bafflestones* contain in-place, stalked fossils that trapped sediment by baffling.

Many reefs also appear to be preferential sites for precipitation of synsedimentary cement (James and Choquette, 1983), and are hard limestone just below the growing surface. The abundance of early cement in many fossil reefs has led some workers to view these buildups as "cementation reefs" rather than biological - sedimentological structures.

Diversity Amongst Reef-Building Metazoans

Relative abundance of different organisms is one of the easiest observations that can be made on a fossil reef, and so potentially one of the most useful. Although intuitively there should be a simple relationship between diversity and environment, recent thinking suggests that this relationship is complex and not straightforward. It seems that diverse faunas (Fig. 5) probably develop when conditions for growth are optimum, i.e., nutrients are in adequate supply and daily chemical and physical stresses are low, *but* when the community is not able to reach competitive equilibrium because of frequent population reduction. If not disturbed, a community will reach competitive equilibrium where a few species dominate, i.e., low diversity. The relative diversity in such settings is probably a complex interplay between frequency of disturbance, population growth rate and nutrient supply. Many of the above concepts have been derived from the study of coral reefs (Connell, 1978) where high diversity is the result of growth in an environment which is relatively nutrient-poor and subject to periodic catastrophic disturbance by tropical storms. (Woodley *et al.*, 1982). An implication of this concept is that we should expect patchiness and evidence of extensive fragmentation and debris formation in the most diverse of fossil reef communities.

Low diversity reef communities are of at least 3 types: 1) those at competitive equilibrium; 2) new communities (those that have moved into a new environment; and 3) those communities subject to severe and continuing chemical and physical stress. Among the factors most likely to stress modern and fossil reef-building communities are: 1) temperature and salinity fluctuations - most modern and likely most ancient reef-builders grow or grew best in tropical sea water of normal salinity; 2) intense waves and swell - the skeletons of most reef-builders will be broken or toppled by strong wave surge; 3) low light penetration - in modern reef-building organisms rapid calcification takes place because light dependent symbionts take over some of the bodily functions of the host; and 4) heavy sedimentation - all reef-builders are sedentary filter-feeders or micro-predators and water filled with fine-grained sediments would clog the feeding apparatus.

Figure 5
A shallow-water (1 m deep) living reef, composed of branching (left), foliose (centre) and hemispherical (centre) corals, off Gouldin Cay, Bahamas.

The Growth Form of Reef-Building Metazoans

The relationship between organism shape and environment is one of the oldest and most controversial topics in biology and paleobiology. In terms of reef-building metazoans, observations from the rock record (Figs. 6 and 7) of the interrelationship between organisms and surrounding sediments, combined with studies of modern coral distribution on tropical reefs, allow us to make some generalizations about form and environment that are useful in reef facies analysis (Fig. 8).

The limitations of applying these concepts directly to fossil reefs has recently been emphasized by Stearn (1982), who pointed out that no general patterns are applicable to all reefs, and that variations in shape are the result of the interaction between environmental factors with the genetically dicatated growth pattern of the organism. Observations on growth form must be used in conjunction with other parameters, and are most useful in providing additional information when dealing with low diversity communities.

THE SPECTRUM OF REEF TYPES

Reefs can develop just about anywhere in the carbonate facies spectrum. As isolated structures they are dispersed across shallow carbonate platforms but they also grow, outpacing subsidence, in slope and basinal settings. As more contiguous elements they commonly form long, sinuous barriers along the margins of the same platforms, close to land along the edge of narrow shelves or as halos around positive structural elements.

Modern reefs are best developed and most successful on the windward sides of shelves, islands, platforms and atolls where wind and swell are consistent and onshore. The asymmetry of many ancient reefs and distribution of sediment facies suggests that this was so in the past as well. The reason for the preferential development of reefs on the windward side is by no means established but sedimentation is likely the most important. Shallow water reef-building species characteristically produce abundant fine sediment, yet the major reef-builders, because they are filter feeders and micropredators, are

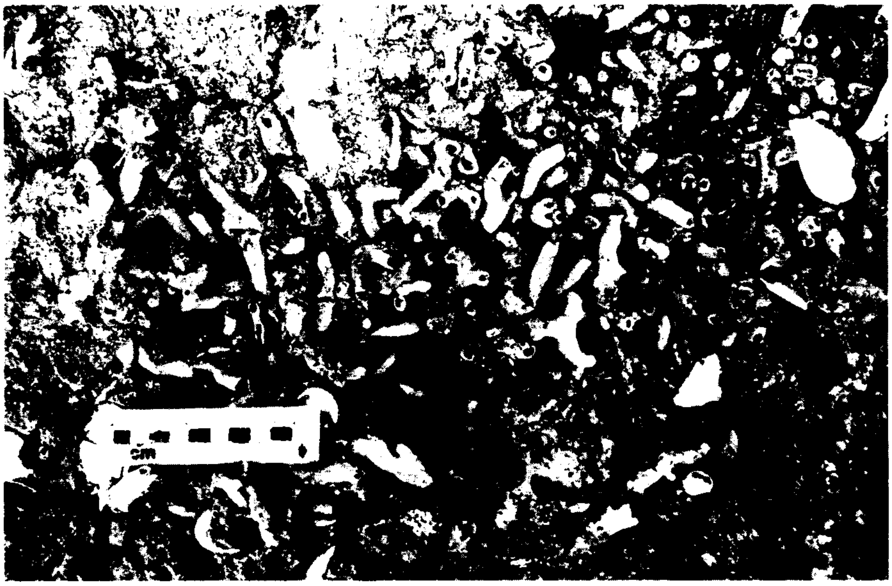


Figure 6
An accumulation of branching corals (Porites porites) *and bivalves in a late Pleistocene reef, Barbados, W.I.*

Figure 7
A small patch of domal shaped corals (Diploria *sp. in cross-section on a cliff exposure of Late Pleistocene reef limestone, Barbados, W.I.*

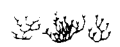
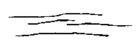
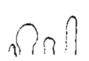
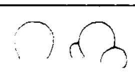
GROWTH FORM AND ENVIRONMENT OF REEF BUILDING SKELETAL METAZOA			
GROWTH FORM		ENVIRONMENT	
		Wave Energy	Sedimentation
	Delicate, branching	low	high
	Thin, delicate, plate-like	low	low
	Globular, bulbous, columnar	moderate	high
	Robust, dendroid, branching	mod-high	moderate
	Hemispherical, domal irregular, massive	mod-high	low
	Encrusting	intense	low
	Tabular	moderate	low

Figure 8
The growth form of reef-building metazoans and the types of environments in which they most commonly occur. From James (1983).

intolerant of fine sediment. The open ocean and windward locations are the only places in which fine sediment is continuously swept away.

Reefs form a natural breakwater when they grow into the zone of onshore waves and swells and create a relatively quiet environment in the lee of the reef crest. Commonly, this restriction significantly changes water circulation on the shelf, platform, or lagoon behind the reef. In such a marginal location, the symmetrical reef facies model comprising a reef-core facies surrounded on all sides by reef-flank facies is no longer discernable. Instead facies are more asymmetrically distributed with the reef-core facies flanked on the windward side by the fore-reef facies and on the leeward side by the platform facies (often called the back-reef facies).

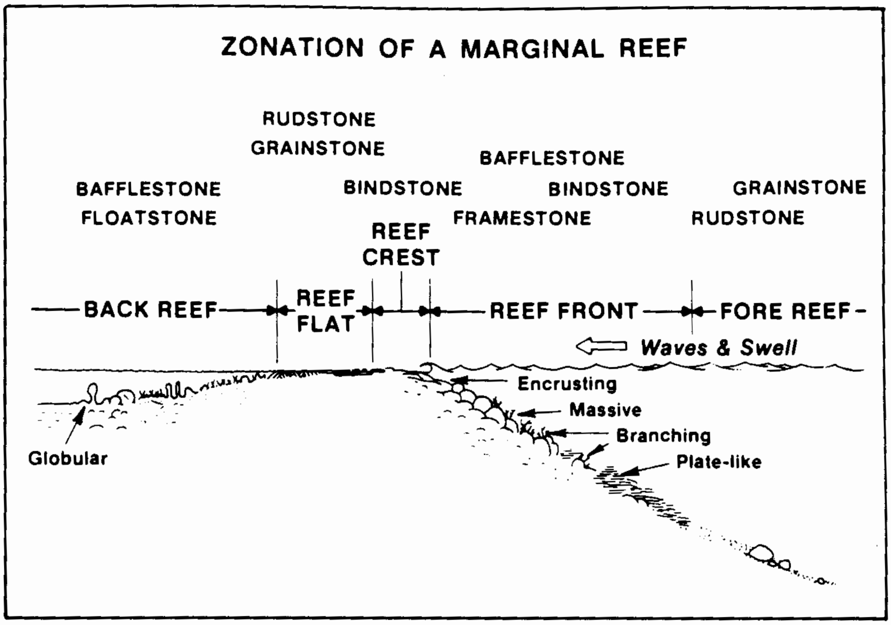


Figure 9
Cross-section through a hypothetical, zoned, marginal reef illustrating the different reef zones, spectrum of different limestones produced in each zone, and environment of different reef-building forms.

High-Energy Reefs

In these high energy settings the reef is distinctly zoned (Fig. 9)

Reef Crest Zone. This is the highest part of the reef at any stage in its growth, and if in shallow water, it is that part of the reef top that receives most of the wind and wave energy. Composition of the reef crest depends upon the degree of wind strength and swell. In areas where wind and swell are intense, only those organisms that can encrust, generally in sheet-like forms, are able to survive. When wave and swell intensity are only moderate to strong, encrusting forms still dominate but are commonly also bladed or possess short, stubby branches. In localities where wave energy is moderate, hemispherical to massive forms occur, with scattered clumps of branching reef-builders. The community is still of low diversity. The lithologies formed in these three cases would range from bindstones to framestones.

Reef Front Zone. This zone extends from the surf zone to an indeterminate depth, commonly less than 100 metres, where the zone of abundant skeletal growth grades into sediments of the fore-reef zone. Direct analogy between modern reefs, especially Caribbean reefs, and ancient reefs is difficult because today the sea floor from the surf zone to a depth of about 12 metres is commonly dominated by the robust branching form *Acropora palmata*, a species which developed only recently in the late Cenozoic. Such branching forms are rarely found in ancient reefs. Instead, the most abundant forms are massive, laminar to hemispherical skeletons, forming framestones and sometimes bindstones.

The main part of this zone supports a diverse fauna with reef-builders ranging in shape from hemispherical to branching to columnar to dendroid to sheet-like. Accessory organisms and various niche dwellers such as brachipods, bivalves, coralline algae, crinoids, and green segmented calcareous algae (Halimeda), are common. On modern reefs where the reef-builders are corals, this zone commonly extends to a depth of 30 metres or so. The most common rock type formed in this zone would still be framestone, but the variety of growth forms also leads to the formation of many bindstones and bafflestones.

Below about 30 metres wave intensity is almost non-existent and light is attenuated. The response of many reef-building metazoans is to increase their surface area, by having only a small basal attachment and a large but delicate plate-like shape. Rock types from this zone look like bindstones, but binding plays no role in the formation of these rocks and perhaps another term is needed.

The deepest zone of growth of coral and green calcareous algae on modern coral reefs is about 70 metres. The lower limit may depend upon many factors, perhaps one of the most important being sedimentation, especially in shale basins which border many reefs. This lower limit should therefore be used with caution in the interpretation of fossil reefs.

Sediments on the reef front are of two types: 1) internal sediments within the reef structure, generally lime mud giving the rocks a lime mudstone to wackestone matrix, and 2) coarse sands and gravels in channels running seaward between the reefs. These latter deposits have rarely been recognized in ancient reefs.

As a result of numerous observations on modern reefs it appears that most of the sediment generated on the upper part of the reef front and on the reef crest is transported episodically by storms up and over the top and accumulates in the lee of the reef crest. Sediments on the intermediate and lower regions of the reef front, however, are transported down to the fore-reef zone only when it is channelled by way of passes through the reef.

Reef Flat Zone. The reef flat varies from a pavement of cemented, large skeletal debris with scattered rubble and coralline algae nodules in areas of intense waves and swell, to shoals of well-washed lime sand in areas of moderate wave energy. Sand shoals may also be present in the lee of the reef pavement. Vagaries of wave refraction may sweep the sands into cays and islands. These obstructions in turn create small protected environments very near the reef crest. Water over this zone is shallow (only a few metres deep at most) and scattered clumps of reef-building metazoans are common. The resulting rock types range from clean skeletal lime grainstones to rudstones.

Back Reef Zone. In the lee of the reef flat, conditions are relatively tranquil and much of the mud formed on the reef front comes out of suspension. This, coupled with the prolific growth of mud and sand-producing bottom fauna such as crinoids, calcareous green algae, brachiopods, and ostracodes, commonly results in mud-rich lithologies. The two most common growth habits of reef-builders in these environments are stubby, dendroid forms, often bushy and knobby, and/or large globular forms that extend above the substrate to withstand both frequent agitation and quiet muddy periods.

The rock types characteristic of this environment are bafflestones or floatstones to occasional framestones with a skeletal wackestone to packstone matrix. In some reefs there are beds of nothing but dis-articulated branches in lime mud (e.g., *Amphipora* limestones of the Upper Devonian), but there is little evidence of much transport.

Fore-Reef Facies

This facies consists of thin to thick and massively bedded skeletal lime grainstones to lime packstones which are composed of whole or fragmented skeletal debris, blocks of reef limestones and skeletons of reef-builders. These grade basinward into shales or lime muds. In contrast to the reef facies, the beds are rarely dolomitized.

It should be remembered that this high-energy zonation, although most commonly observed on platform margin reefs, is also developed in on-platform isolated reefs and in reefs from slope or basinal settings that rise into the zone of breaking waves.

Low Energy Reefs

As wave energy decreases, so the style of reef growth changes; distinctive zonation becomes less noticable, shape of the reef building organisms is different and relative diversity decreases. The relationship between wave energy and reef type has been summarized for modern reefs by Geister (1980) and for fossil reefs by Wilson (1975) and is shown in Figure 10. Lower energy situations are also characterized by sluggish water exchange and so these changes may be amplified by variations in water salinity and nutrient content.

As a result most isolated reefs or patch reefs on carbonate platforms are poorly zoned and more like the ideal reef in Figure 1. They are generally circular to elliptical to irregular in plan and may be large enough to enclose a lagoon themsleves. Each reef is zoned with respect to depth, similar to the reef front in higher energy reefs.

Finally, in some settings, reefs as described above do not occur. These settings include the very inner parts of shallow platforms, the deeper (many 10s of metres) lagoons, and those platforms or parts of platforms covered by water of elevated or depressed salinity. There are, however, accumulations of carbonate sand and/or mud, built by organisms that are very close to reefs in composition. In modern seas these range from sea-grass banks to skeletal sand shoals, often bound by algae, to broad banks of corals and algae. Although not common today, these structures are an important part of the fossil carbonate spectrum and must be integrated into the reef model.

Reef Mounds

Many Phanerozoic carbonate sequences contain structures that some workers call reefs, some call mounds, and some call banks. They lack many of the characteristics we ascribe to reefs yet were clearly rich in skeletal organisms and had relief above the sea floor. The origin of these structures, which I have called *reef mounds* (Fig. 11), has probably caused more discussion than any other topic in the literature on reefs (Heckel, 1974).

Reef mounds are, as the name suggests, flat lenses to steep conical piles with slopes of up to 40° consisting of poorly sorted bioclastic lime mud with minor amounts of organic boundstone. With this composition they clearly formed in quiet water environments and from the rock record appear to occur in preferred locations: 1) arranged just downslope on gently-dipping platform margins (Fig. 12); 2) in deep basins; and 3) spread widely in tranquil reef lagoons or wide shelf areas. When viewed in section, reef mounds display a similar facies sequence in each case (Wilson, 1975) (Fig. 11).

Stage 1. Basal bioclastic lime mudstone to wackestone pile - muddy sediment with much bioclastic debris but no baf-

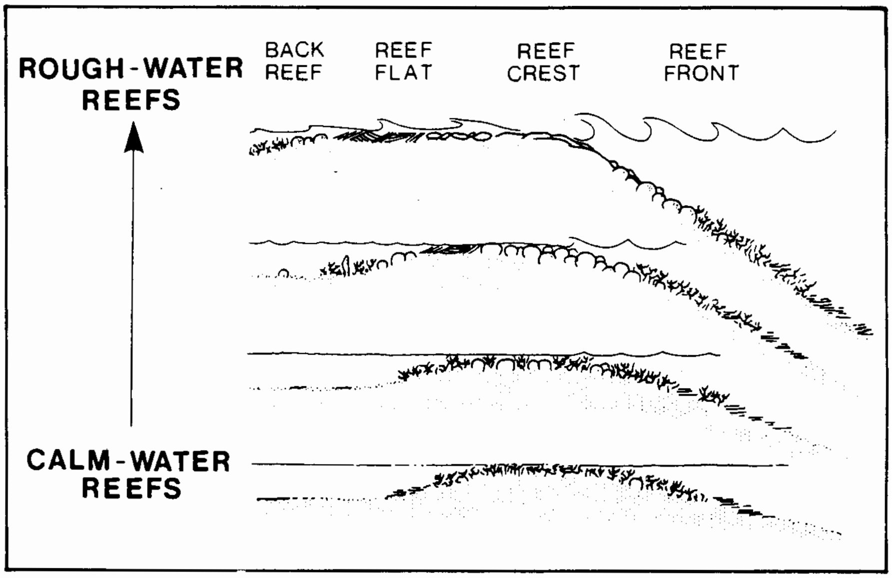


Figure 10
Generalized diagram of the different zonations expected from reefs growing under conditions ranging from calm water to rough water; see Figure 8 for growth forms.

fling or binding organisms.

Stage 2. Lime mudstone or bafflestone core - the thickest part of the mound, consisting of delicate to dendroid forms with upright growth habits in a lime mudstone matrix. The limestone is frequently brecciated, suggesting partial early lithification, dewatering and slumping, and contains stromatactis. Each geologic age has its own special fauna that forms this stage: Lower Cambrian - archaeocyathans; Middle to Lower Ordovician - sponges and algae; Middle Ordovician, Late Ordovician, Silurian, Early Carboniferous (Mississippian) - bryozoans; Late Carboniferous (Pennsylvanian) and Early Permian - platy algae; Late Permian to Middle Triassic - sponges and algae; Late Triassic - large fasciculate dendroid corals; Late Jurassic - lithistid sponges; Cretaceous - rudist bivalves.

Stage 3. Mound cap - a thin layer of encrusting or lamellar forms, occasional domal or hemispherical forms, or winnowed lime sands.

The massive, commonly well-bedded carbonates that flank the reef mounds comprise extensive accumulations of debris and chunks of archaeocyathans, pelmatozoans, fenestrate bryozoans, small rudists, dendroid corals, stromatoporoids, branching red algae or tabular foraminifers along with wholly to partly lithified lime mudstone. Volumetrically these flank beds may be greater than the core itself and almost bury it (Fig. 13).

Carbonate Mud Mounds

One end member of the reef mound category that deserves special attention is a group of structures called either Waulsortian mounds (from the name of a village in Belgium) or more commonly, carbonate mud mounds. These puzzling structures may be just as large and have sides just as steep as reef mounds, but they possess *no* large skeletons. They are made up only of crinoid fragments, sponge spicules or scattered bryozoans that together make up no more than 1/4 of the rock, the rest being lime mud (for a recent review see Pratt, 1982). These mud mounds seem to

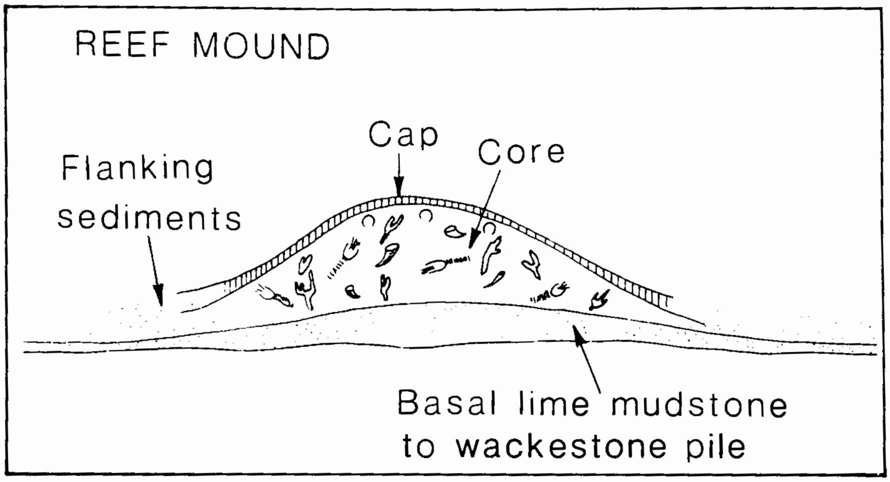


Figure 11
Cross-section through a hypothetical reef mound illustrating the geometry of the different facies.

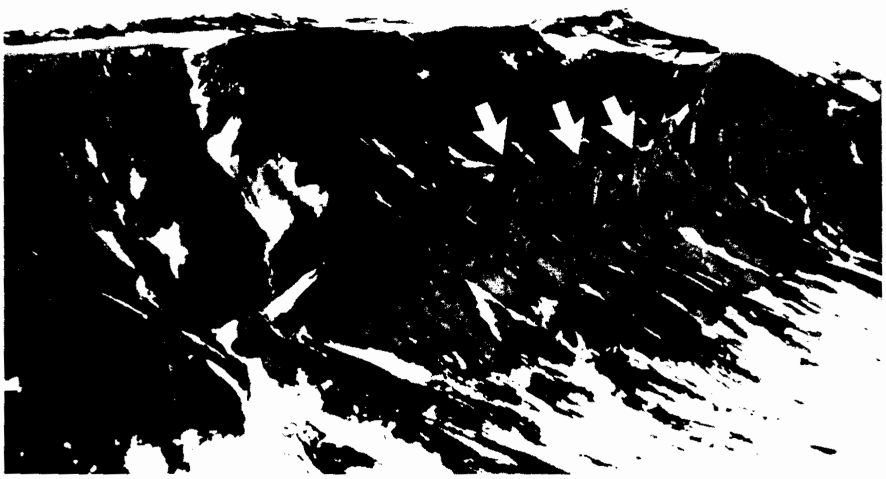


Figure 12
Massive reef limestone (right) of the Nansen Formation (Permo-Pennsylvanian) extending downward and basinward into dark, argillaceous limestones of the Hare Fiord Formation (left). Arrows point out small reef mounds developed on the seaward slopes of the reef front, western side of Blind Fiord, Ellesmere Island, N.W.T.

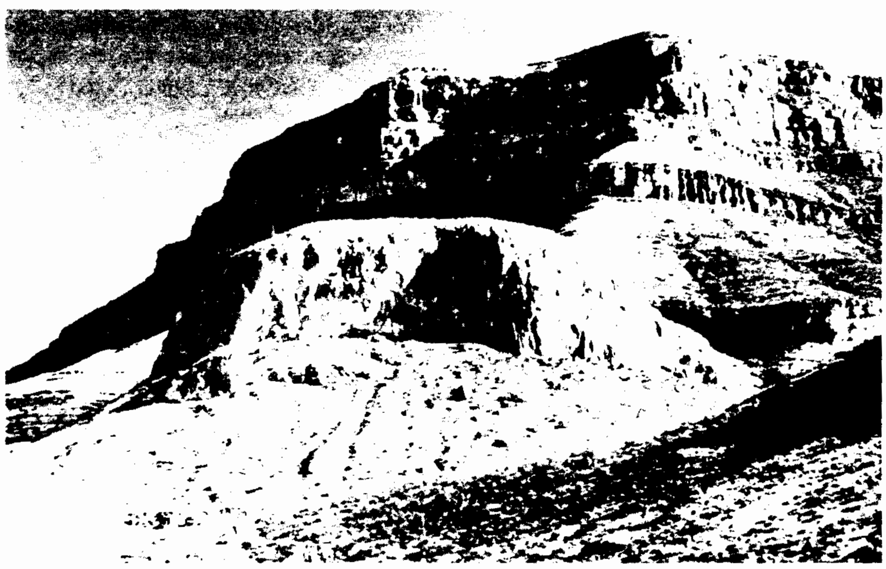


Figure 13 ▶
Muleshoe bioherm, a 60 m high reef mound of Late Missippian age exposed along the western escarpment of the Scaremento Mountins, New Mexico.

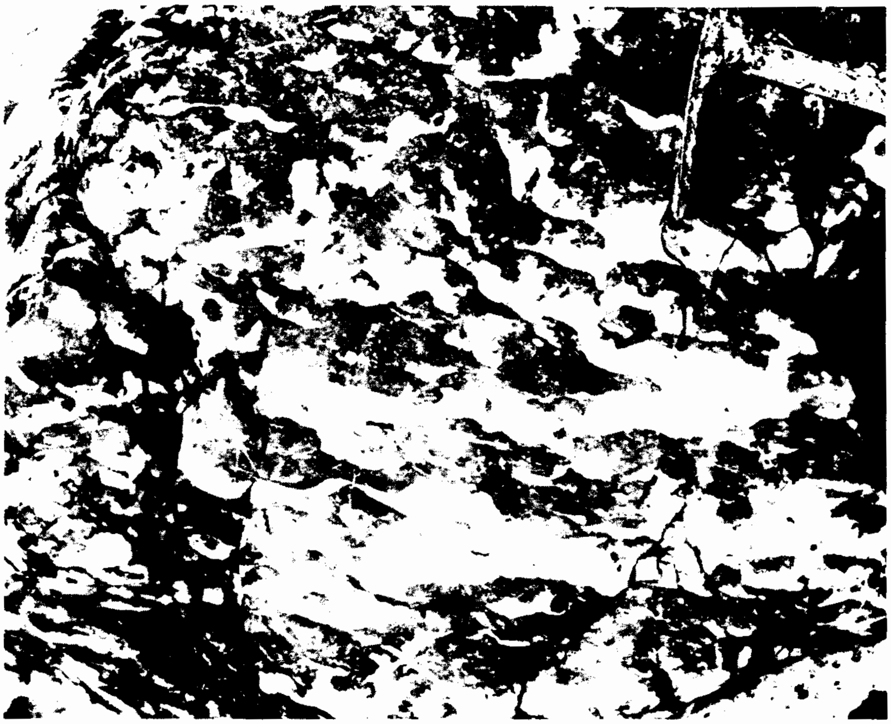


Figure 14
Stromatactis, *sub-horizontal layers of calcite spar illustrating irregular digitate tops and smooth bottoms, from Gros Morbe, reef-mound facies, Silurian West Point reef complex, Gaspé, Quebec.*

occur almost exclusively in deep water on carbonate slopes, and are not as common as on-shelf buildups.

Almost all workers have commented on the striking similarity of these structures regardless of their age, which suggets a common origin for all of them. There are, however, no obvious modern analogues. Structures that come closest are seagrass-stabilized, shallow-water mudbanks or deep-water lithoherms, but both also have considerable differences. The localization of the mud is probably some combination of *in situ* production, baffling, and binding by a variety of organisms, among which algae and cyanobacteria are the most important. The seafloor topography is clearly aided by rapid seafloor lithification.

Central to their genesis is the presence of *"Stromatictis"* (Fig. 14) which consists of masses of calcite-filled cavities which have a digitate roof and a flat floor commonly formed by geopetal sediment. This spar occurs in swarms and has a reticulate distribution (see Bathurst, 1982 for an extended discussion). It is clear that cavity development is penecontemporaneous with deposition, but its origin is still a mystery. Current hypotheses include: 1) dewatering, 2) compaction, slumping, and down slope creep of consolidated but unlithified sediment, 3) cementation and/or binding of sediment by algae to form crusts followed by erosion, and 4) decay and collapse of organic tissue (especially sponges) after partial lithification.

A final unsettling fact is that these mud mounds are almost entirely a Paleozoic phenomenon. They range in age from middle Cambrian to late Jurassic but confirmed stromatactis is found only in Paleozoic buildups. There is no clear answer as to why they do not occur in late Mesozoic and Cenozoic carbonates.

Stromatolite Reefs

During the Precambrian and earliest Paleozoic, prior to the appearance of herbivorous metazoans, stromatolites formed impressive build-ups (Fig. 15). These stromatolite complexes clearly had relief above the sea floor and in terms of morphology were surprizingly similar to later skeletal reefs. Mostly developed in shelf margin settings some exhibit excellent lateral zonation (Hoffman, 1974) while others clearly developed in a series of well-defined stages in which stromatolites had different growth forms (Cecile and Campbell, 1978). Although there has been much study of Precambrian stromatolites, much less attention has been paid to their role in the formation of reefs or reef mounds.

STAGES OF REEF GROWTH

While the composition of the reef core and the different facies can be determined from both modern and ancient examples, information as to the stages of reef growth can only come from the rock record.

It has long been recognized that there is an ecological succession in many Paleozoic reefs (Lowenstam, 1950), i.e., the replacement of one community of reef-building organisms by another as the reef grew. A synthesis by Walker and Alberstadt (1975) of reefs ranging in age from Early Ordocician to Late Cretaceous suggests that a similar community succession is present in reefs throughout the Paleozoic and Mesozoic. Application of this concept to Cenozoic reefs (Frost, 1977), which are dominated by scleractinian corals (the reef-builders in today's oceans), allows us to equate ancient reef community succession with observations on modern reef communities with some measure of confidence.

The reason for this ecologic succession is at present a topic of much debate. Some workers feel that the con-

Figure 15
A large stromatolite mound of Proterozoic age Kuuik Formation, Kilohigok Basin, Northwest Territories. Photograph by M. Cecile.

trol is extrinsic, reflecting a progressive replacement of deep-water communities by shallower water ones as the reef grows to sea level and into more turbulent water; however, there is often abundant evidence that the first two stages are developed in shallow water. Other workers feel that the control is intrinsic, reflecting a natural succession as the organisms gradually alter the substratum and change the energy flow pathways as the community develops; nevertheless, there is abundant evidence of increasing water turbulence as the structure grows.

THE MODEL

The reef model is an integration of information from the modern and the fossil record. The major stumbling block that makes the generation of a model difficult, the ever-changing character of the reef-building fauna and flora through time, is resolvable if the concept of community succession is used as a basis. In the model four separate stages of reef growth are recognized, and these stages, along with the types of limestone, relative diversity of organisms and growth-form of reef-builders in each, are summarized in Figure 16.

1) *Pioneer (Stabilization) Stage.* This first stage is most commonly represented by a series of shoals or other accumulations of skeletal lime sand composed of pelmatozoan or echinoderm debris in the Paleozoic and Mesozoic, and plates of calcareous green algae in the Cenozoic. The surfaces of these sediment piles are colonized by algae (calcareous green), plants (sea grasses) and/or animals (pelmatozoans) that send down roots or holdfasts to bind and stabilize the substrate. Once stabilized, scattered branching algae, bryozoans, corals, soft sponges and other metazoans begin to grow between the stabilizers.

2) *Colonization Stage.* This second stage is relatively thin when compared to the reef structure as a whole, and reflects the initial colonization by reef-building metazoans. The rock is generally characterized by few species, sometimes massive or lamellar forms but more commonly monospecific coppices or clumps of branching forms (Fig. 8). In Cenozoic reefs the one characteristic common to all corals in this stage of reef growth is that they are able to get rid of sediment and clean their polyps, and so are able to grow in areas of high sedimentation. The branching growth form creates many smaller subenvironments or niches in which numerous other attached and encrusting organisms can live, forming the first stage of the reef ecosystem.

STAGES OF REEF GROWTH

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

Figure 16
A sketch of the four divisions of the reef-core facies with a tabulation of the most common types of limestone, relative species diversity and shape of reef-builders found in each stage.

Figure 17
The diversification stage of an Upper Devonian reef, comprising domal stromatoporoids, and domal to branching tabulate corals, Blue Fiord Formation, south side of Eids Fiord, Ellesmere Island, N.W.T.

3) *Diversification Stage* This third stage, usually represented by the bulk of the reef mass, is the point at which most pronounced upward-building towards sea level occurs and easily definable, lateral facies develop. The number of major reef-building taxa is usually more than doubled, and the greatest variety in growth habit is encountered (Fig. 17). With this increase in form and diversity of framework and binding taxa comes increased nestling space, i.e., surfaces, cavities, nooks and crannies, leading to an increase in diversity of debris-producing organisms..

Once a reef reaches the diversification stage, and sometimes even earlier in the colonization stage, the structure is frequently high enough above the sur-

Figure 18
A small patch reef built by bryozoans, corals, and stromatoporoids. Long Point Formation (Middle Ordovician), Port-au-Port Peninsula, Newfoundland.

Figure 19
Reef and reef-flank deposits (R) ca. 100 m thick (Peechee Formation) of Upper Devonian age in the Flathead Range, southern Rocky Mountains, Alberta. Photograph courtesy B. Pratt.

rounding sea floor to affect water circulation and thus to alter sedimentation patterns. At this point not only are surrounding sedimentary environments altered but the reef itself develops a zonation, because its margins now reach from shallow to deep water. This zonation is, as outlined in an earlier section, dependent upon wave energy.

4) *Domination (Climax) Stage.* The change to this stage of reef growth is commonly abrupt. The most common lithology is a limestone dominated by only a few taxa with only one growth habit, generally encrusted to laminated. Most reefs show the effects of surf at this stage, in the form of beds of rudstone.

While these four stages are the norm, some reefs only display the upper two or three stages. Careful investigation, however, usually reveals that the reefs began growth on a hardground or lithified substrate.

The Complete Reef Structure

Reefs in the rock record vary widely in size. A complete Ordovician (Fig. 18) or Cretaceous reef, formed by a variety of organisms and displaying several stages of growth, may be only as large as a single coral head in a Devonian, Jurassic, or Pleistocene reef, kilometres long and hundreds of metres thick (Fig. 19).

While some large reefs display all 4 stages of development, most are internally stratified (Fig. 20) to form a series of superimposed or stacked reefs. Individual episodes of reef growth are commonly separated by periods of exposure, reflected in the rock by intensive diagenesis, calcrete horizons, karst, shales (paleosols) or by periods of non-deposition as indicated by hardgrounds, borings, or manganese and phosphate-impregnated bedding planes. When reef growth begins again after a hiatus, because the surfaces are both hard and elevated, it starts at the diversification stage, and so only two stages are present in the next package. Some workers tend to regard these unconformity-bounded layers as the different stages of reef growth which they are not. The different stages of reef growth are found as separate units between these layers.

THE MODEL AS A FRAMEWORK OR GUIDE FOR OBSERVATIONS

Reefs

The reef facies model is predicated on the assumption that a full spectrum of reef-building organisms are present. We see a full spectrum in the tropical oceans today, but this was not the case for much of the Phanerozoic. The critical element that is often missing, and without which the four stages of development in the reef core cannot occur, is the presence of skeletal metazoans that secrete large robust, branching, hemispherical or tabular skeletons. Without them the reef cannot exist in the zone of constant turbulence, usually wave induced, because smaller and more delicate forms would be broken and swept away (unless submarine cementation is very rapid, pervasive, and near-surface). This zone of turbulence is the optimum area for growth and diversity because sediment is constantly removed, water is clear, and nutrients are constantly swept past the sessile organisms. Such large skeletal metazoa were present only at certain times during the Phanerozoic (Fig. 21), and each period has its own specialized group of frame-builders: 1) Middle and Upper Ordovician - bryozoa, stromatoporoids, tabulate corals; 2) Silurian and Devonian - stromatoporoids, atabulate corals; 3) Late Triassic - corals, stromatoporoids; 4) Jurassic - corals, stromatoporoids; 5) Upper Cretaceous - rudist bivalves; and 6) Oligocene, Miocene, Plio-Pleistocene - scleractinian corals. A more detailed review of these reefs during specific periods is given by James (1983).

Although reefs are found in platform margin, on-shelf, and basinal settings during these periods, reef mounds also develop, but only in environments which were inimical to the growth of larger metazoans. At many times during the Phanerozoic, however, the large frame-building metazoans were absent, and during these periods reef mounds were the only buildups in the facies spectrum (Fig. 21).

When viewed against the backdrop of the general reef facies model, I think of reef mounds as half-reefs or incomplete reefs because they represent only stages one, two, and occasionally four of the model. These structures did not develop the other upper stages either

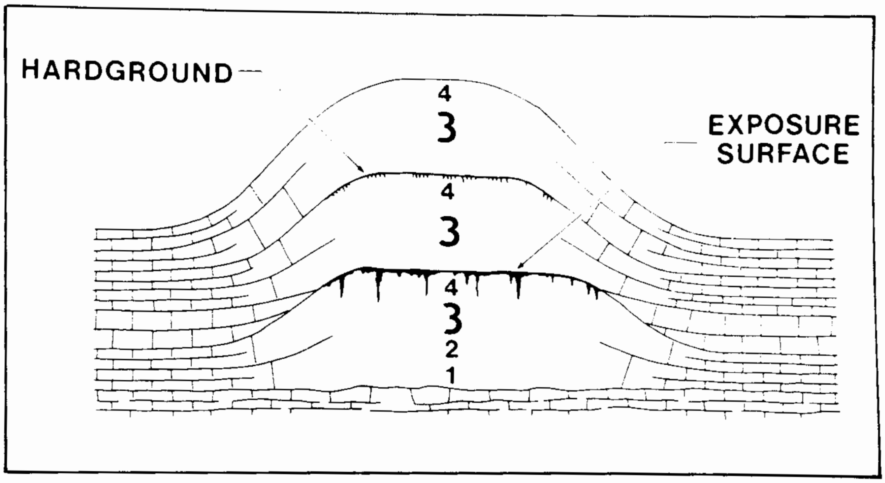


Figure 20
Sketch illustrating a large reef composed of a series of smaller stacked reefs, each with several growth stages (see Fig. 16) and each separated by exposure surfaces or hardgrounds. The size of the numbers represents relative thickness of the stages.

because the environment was not conducive to the growth of large skeletal metazoans or because these larger metazoans simply did not exist at the time when the structure formed.

THE MODEL AS A BASIS FOR INTERPRETATION

The model dictates that reefs, especially large ones, have a series of growth stages. When interpreting an individual structure, which will differ from the model in a variety of ways, special attention must be paid to this internal stratigraphy. If some stages are missing, are thicker or thinner than should be expected, or if the sequence is inverted, then these differences should be a clear signal that extrinsic factors such as sea level fluctuations or changes in subsidence rates occurred during reef growth. In this way the reef is a much more sensitive indicator of such changes than surrounding subtidal facies.

On a larger scale a major question that often arises is whether the reef grew all at once and stood high above contemporaneous basinal strata only to be subsequently buried, or whether it grew in increments, with intervening periods of subaerial exposure and possible continuing off-reef deposition and so never rose very much above the surrounding sea floor. These two hypotheses not only interpret the reef differently but dictate how surrounding basinal deposition took place and the nature of early diagenesis, and can only be separated by careful study of reef stratigraphy.

Finally reefs and associated platform facies are commonly important hydrocarbon reservoirs. In exploitation of these resources it is critical to determine whether the reservoir is a more or less uniform structure with good vertical as well as horizontal permeability or whether it is broken into a series of layers with poor vertical connection because of permeability barriers such as hardgrounds or paleosols (shales).

THE MODEL AS A PREDICTOR

If we know the age of a sequence of carbonate rocks and we have some idea of the paleotectonic setting then we can predict, from limited data, the types of reefs we might expect to be present in a shelf or platform setting.

Times When a Complete Spectrum of Reef Builders was Present

The edge of the shelf or platform is occupied by a marginal reef. The reef is well zoned if the front is steep and wave action intense because of the gradation from rough surface waters down the reef front into progressively more tranquil conditions. Zonation is weak, however, if the reef front slopes gradually seaward and the seas are relatively quiet. Patch reefs can occur anywhere on the platform and each is zoned with respect to depth. Reef mounds are found on the linear shallow parts of the shelf in areas of normal salinity but turbid water. Reef mounds also occur at

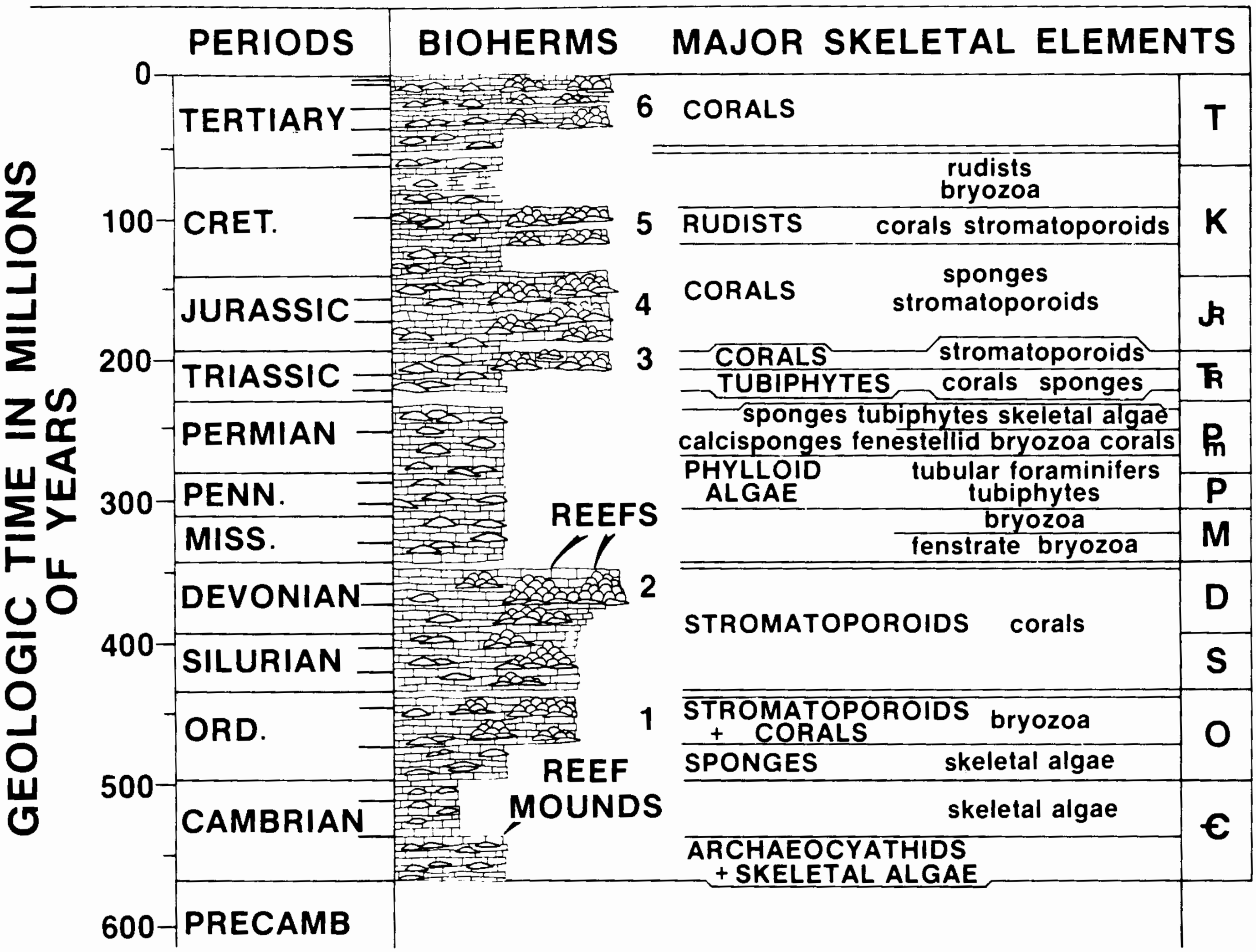


Figure 21
An idealized stratigraphic column representing the Phanerozoic and illustrating times when there appear to be no reefs or bioherms (gaps), times when there were only reef mounds and times when there were both reefs and reef mounds. The numbers indicate different associations of reef-building taxa discussed in the text. (From James, 1983).

depth, in front of the barrier reef down on the reef front or fore-reef.

Patch reefs or reef mounds commonly form a very widespread lithofacies compared to the barrier reef. The stratigraphic thickness of these reefs is dependent upon the rate of subsidence: if subsidence rate is low, reefs are thin; if subsidence rate is high, reefs may be spectacular in their thickness.

Times When Only Delicate, Branching and Encrusting Metazoa Prevail

The margin of the shelf or platform is normally a complex of oolitic or skeletal (generally crinoidal) sand shoals and islands. The only reef structures are reef mounds which occur below the zone of active waves down on the seaward slopes of the shelf or platform and if conditions are relatively tranquil behind the barrier, on the shelf itself. Mounds may display a zonation, with ocean-facing sides in shallow water armoured with accumulations of fragmented and winnowed skeletal debris.

Not only can we predict adjacent facies with some confidence but given only small outcrops or pieces of core we can, within limits, predict the style of the reef in question. For example, a core composed of lime mudstone to floatstone with scattered bryozoans and stromatactis from a Mississippian sequence, when viewed from the point of view of the model clearly points to a reef mound, which since reefs proper did not form at this time, may be 10's to 100's of metres thick. A similar rock in a Silurian sequence, while predicting a similar structure if surrounded by slope or inner shelf strata, would in most other cases predict that the sample is from the basal part of a large reef. Alternatively a small Devonian outcrop, composed of a bafflestone of fasciculate stromatoporoids grading up into a framestone of large tabulate corals and many different stromatoporoids would predict that a massive reef should lie above.

The clear limitations of the model are, however, when using small samples. Reefs are generally large and heterogeneous structures so that a drill core for example may pass through areas of sediment between large skeletons, a very likely possibility since 30% to 40% of modern reefs are sediment or pore space, and so give a false picture of the true deposit.

SUMMARY

The purpose of this article has been to marry the sedimentological and paleontological approaches to the study of reefs into a single facies model, useful to both disciplines. This model is an integration of data from two very different sources: from the modern sea floor, predominantly in the horizontal dimension; and from the rock record, predominantly in the vertical dimension as recorded in mountain exposures, quarries and drill core.

To alter and refine this model more information is needed from two areas. We must learn more about the succession of organisms and sediments that underlies the living surface of modern reefs, by drilling into these reefs. We must learn more about reefs from those parts of the stratigraphic record where reefs are known to occur, but have been little studied – the Precambrian, the Lower Paleozoic, and Cenozoic.

The trend in the past has been to compare specific fossil reefs with modern reefs. The comparative approach has just begun on fossil reefs, to compare and contrast the sedimentology and paleoecology of reefs formed by different groups of organisms at different times in geologic history.

ACKNOWLEDGEMENTS

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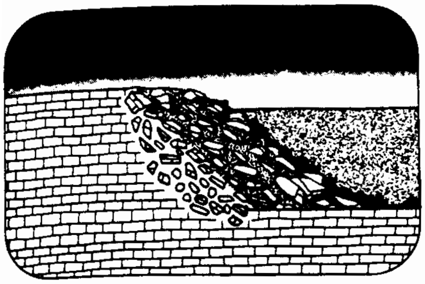

Carbonate Slopes

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INTRODUCTION

To any geologist who has seen them in the field, sediments that comprise the slope facies of carbonate complexes are often the most staggering and long remembered of all. The sheer size of the enormous limestone blocks chaotically intercalated with delicately laminated lime mudstones tests our understanding of sediment genesis and deposition more than for almost any other deposits.

While the sediments themselves are intriguing, they are also useful as the only remaining clues as to the nature and composition of a new dolomitized or tectonically obliterated platform margin. Furthermore, the very presence of huge blocks is an excellent indicator of a nearby carbonate platform or reef complex, and this principle has been successfully used to locate reefs in the subsurface. The lime sands and conglomerates of these deposits, where intercalated with organic-rich basinal sediments, can be reservoirs for oil and natural gas.

We cannot interpret these deposits with the same level of confidence as shallow water carbonate sediments because: 1) modern slope deposits are not easily accessible for field study (the limited use of small research submersibles and seismology is slowly changing this); 2) ancient slope deposits commonly occur in orogenic belts, where facies and tectonic relationships are so complicated that these deposits are often mistaken for tectonic breccias or mélanges; 3) slope sediments in the subsurface generally have not been serious exploration targets as long as adjacent platform margins remained the primary objective; 4) slope sediments are formed in a series of environments that transect major pressure, temperature and oxygen-level boundaries in the ocean and the precise effects of these physiochemical parameters on the sediments are poorly known.

As a result our present facies models are based on the rock record, with some additions from recent sediments. In addition, our understanding of downslope sediment emplacement is based in part upon processes and models determined for siliciclastic deposits (Dott, 1963; Middleton and Hampton, 1973). It has been easy to apply this comparative approach (Cook *et al.*, 1972; Cook and Mullins, 1983) because many similarities do exist, but there are also fundamental differences.

In the paper on "Turbidites and Associated Coarse Clastic Deposits" (this volume), Walker outlined the attributes of a turbidite model and then integrated all associated siliciclastic lithofacies that encompass the slope-to-basin transition into an overall, larger scale, submarine fan model. Similarly our approach in this article will be to outline the major aspects of the slope facies in carbonate sedimentary sequences, first by examining the major sediment types and their modes of emplacement, and second by relating these to general facies models, which are very much dependent upon the nature of the adjacent margin and the depositional setting.

Since the first edition of this summary appeared there has been an increased interest in synthesizing the various aspects of fore-reef and slope carbonates and excellent, well-illustrated summaries have been written by Read (1982), James and Mountjoy (1983), Enos and Moore (1983), and Cook and Mullins (1983).

CARBONATE SLOPE SEDIMENTATION

The slope facies is a transitional one between the rapid and active production of calcium carbonate in shallow water and the slow gentle rain of fine-grained pelagic sediments in the basin. The platform-to-basin transition may in places be abrupt, in the form of a steep cliff, but more commonly is a gently inclined slope decreasing in grade with depth and merging imperceptibly into basinal deposits at some distance, which may be 100s of km from the actual margin. Because the environment as a whole is an incline, short periods of gravity-induced catastrophic sedimentation alternate with long periods of relatively quiet pelagic sedimentation, or to paraphrase Ager (1973, p. 100), "long periods of boredom alternating with short periods of terror".

Pelagic Carbonates

Pelagic carbonates are those sediments deposited in the open sea and derived from the skeletons of planktonic micro-organisms which inhabit the overlying water column. Such deposits include ooze and its lithified equivalent, chalk, and consist primarily of the skeletons of various nannofossil groups, especially coccoliths, the tests of planktonic and sometimes benthic foraminifers. Macrofossils such as pteropods, pelecypods, echinoderms and, in older units, ammonites, nautiloids, tentaculitids and styliolinids are present as accessory components. An excellent summary of such deposits can be found in Hsu and Jenkyns (1974), Scholle (1977), and Scholle *et al.* (1983).

True pelagic carbonates are apparently not known from the early Paleozoic and are first recognized from rocks of Upper Silurian age (Tucker, 1974). Planktonic foraminifers and coccoliths appear to have evolved in the Jurassic. During post-Jurassic time pelagic carbonate has increased to the point that in the last 100 Ma it comprises about 67% of worldwide carbonate deposition, and more than 50% of the present sea floor is covered with this type of carbonate sediment (Hay *et al.*, 1976).

Most chalks accumulate a a rate of between one and 30 cm per year. The sedimentary structures and colours depend upon the degree of circulation and oxygenation. Dark colours and preserved laminations reflect stagnation; lighter colours, more burrows and fewer preserved sedimentary structures reflect stronger bottom circulation.

Because of relatively narrow shelves and low sea level stands, Holocene pelagic sediments are restricted largely to deep ocean basins. However, during times of eustatic high stands of sea level, pelagic carbonates can and did accumulate on continental shelves, such as the North American mid-continent seaway in Cretaceous times.

The water depth of Recent pelagic carbonate deposition ranges from less than 100 m to greater than 4500 m. The limiting factors for such accumulations are the relative rates of sedimentation of carbonate versus non-carbonate components, physical erosion by submarine currents and chemical dissolution. Chemical dissolution is particularly important in carbonate slope facies because the environment passes, with depth, through several important increasing pressure and decreasing temperature boundaries (James and Choquette, 1983). Aragonite components, such as pteropods and benthic foraminifers, may be selectively removed by dissolution in water as shallow as 500 m (the aragonite compensation depth) while calcite components are completely dissolved at the carbonate compensation depth, between 4,000 and 5,000 m in today's oceans. Much less is known about the removal or recrystallization of Mg-calcite. This progressive removal by dissolution results in a residual sediment composed largely of siliceous skeletons, red hemipelagic clays and wind-blown silt.

In some ares of the modern ocean the production of siliceous plankton (silicioflagellates, diatoms and radiolaria) exceeds that of calcareous nanno- and microplankton. During the Paleozoic, when pelagic carbonate was reduced or absent, siliceous sediment was much more widespread in deep-water areas.

Hemipelagic Slope Sediments

Sediments that make up the fine-grained pelagic component of most slope deposits come not only from the water column but from the adjacent platform as well (Wilson, 1969). While the contribution at any one time from the water column is more or less constant, that portion derived from the platform is episodic. Most often storms stir up the wide, shallow, mud-floored areas of the shelf and the milk-white water streams out across the shelf margin to settle in deep water. A less voluminous but more regular transfer process exists at such near-vertical shelf-to-deep-oceanic-basin transitions as St. Croix, Virgin Islands, where warm sediment-rich shelf waters "float" over the cooler basinal waters by tidal exchange. These fine-grained, shallow-water derived slope sediments have been called "peri-platform ooze" by Schlager and James (1978) because they occur as an apron around the platform and because they are significantly different in their mineralogy and composition from the wholly pelagic sediments of the open sea.

In the Precambrian and Paleozoic most pelagic slope carbonates may well have been almost wholly peri-platform ooze.

The resultant hemipelagic slope deposits are monotonous, uniform dark grey, fine-grained lime mudstones, generally thin-bedded with flat planar contacts and internal micro-laminations (Fig. 1). Mudstone beds are often separated by partings into very thin beds of similar mudstone or beds of shale, forming characteristic "rhythmites" or "ribbon limestones". The original depositional textures and fabrics are often modified by downslope creep leading to sedimentary boudinage, while differential compaction and/or cementation frequently transforms the evenly-bedded sediments into a nodular limestone. The irregular nodules may, in some cases, be so packed together to form a jig-saw puzzle resembling an *in situ* breccia.

From the Mesozoic to Recent, pelagic components become progressively more important. The criteria for distinguishing the relative contribution of peri-platform vs pelagic ooze is speculative because most fine silt-mud size shallow water indicators are susceptible to early diagenesis (Enos and Moore, 1983).

Peri—Platform Talus

Directly seaward of the shallow water reefs or lime-sand shoals that form a platform margin, there is commonly a debris apron of limestone blocks (Fig. 2), skeletons of reef building metazoa, sediment bounded by submarine cement or encrusting organisms, lime sand and muds. These accumulations are the result of rock-fall and sand-streams from shallow water and, as illustrated in Figure 3, are very common along the seaward margins of modern reef complexes (James and Ginsburg, 1979; Mullins and Neumann, 1979; Schlager and Chermak, 1979; Land and Moore, 1977). The blocks themselves may be multi-generation in composition because the reefs, sand shoals and other deposits at the platform margin are characteristically susceptible to early lithification, either by submarine cementation, or if there are slight fluctuations in sea level, by complex subaerial diagenesis. In addition, parts of the talus wedge are commonly cemented on the sea floor (James and Ginsburg, 1979; Land and Moore, 1977). The lithified portions of these limestones become hard and brittle, and so

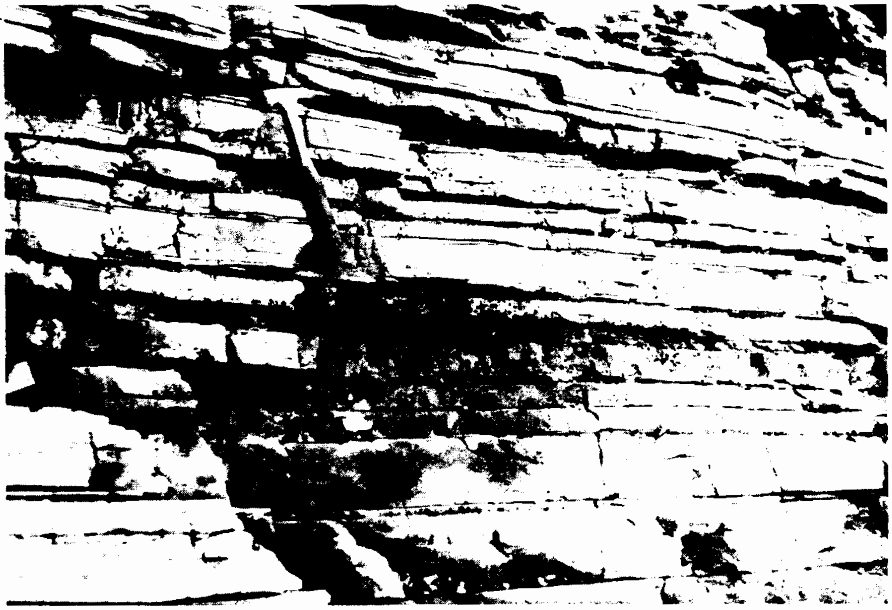

Figure 1
Peri-platform ooze; evenly-bedded, grey lime mudstone with thin interbeds of argillaceous lime mudstone, Cooks Brook Formation (Middle Cambrian), Humber Arm, Western Newfoundland.

are particularly susceptible to fracturing and fragmentation.

Large passes through a reef also act as conduits, funnelling back-reef sediments into this zone so that, along strike, areas of chaotic breccia may alternate with fans of lime sand. The later sediment is also commonly cemented, forming numerous hardgrounds.

Examination of sediment dispersal seaward of the platform in areas with low to intermediate slopes (up to 30°) indicates that this talus does not travel any significant distance away from the margin by day-to-day processes.

Carbonate Breccias and Conglomerates

These deposits have been called debris flows (deposits), submarine mass flows, mass breccia flows, breccia and megabreccia beds, debris sheets, rudite sheets, debris avalanches, or olistostromes (in the non-tectonic sense). They are certainly the most impressive parts of the slope sequence. They originate in two very different areas, high up on the slope in shallow water or from lower down the slope profile.

Figure 2
Peri-platform talus: a block of shallow-water reef limestone (approximately 30m high) enclosed in thin-bedded, dark grey, peri-platform lime mudstones. Block occurs approximately 250m down slope from the toe of a near-vertical, 200m high platform margin. Note vertical orientation of bedding within the block, Cathedral Formation (Middle Cambrian), north face Mt. Stephen, British Columbia.

A) Breccias Derived from Shallow Water. These breccias are generally exposed in discontinuous to laterally extensive sheets, channels with lenticular cross sections or irregular masses. They stand out as resistant masses of light-coloured carbonate against a background of dark-coloured, well-bedded limestone and shale (Figs. 4 and 5). They are characterized by blocks of all sizes and shapes, but often equidimensional and somewhat rounded. Some of the blocks are so enormous that they have been mistaken for bioherms (see Mountjoy *et al.*, 1972). One exceptional clast in the Cow Head Group (Cambro-Ordovician) at Lower Head, Newfoundland is 0.2 km x 50 m in size, with surrounding blocks often 30 x 15 m in size, (Kindle and Whittington, 1958). The breccias commonly have a matrix of lime mud, lime sand or argillaceous lime mud.

The deposits are bedded, with a planar to undulating basal contact accentuated by differential compaction and a planar to irregular and often hummocky upper contact. The nature of the bedding contacts often cannot be determined accurately because the bedding planes are stylolitic, and so any original bedding-plane features are often destroyed. Bedding thicknesses range up to tens of metres. Davies (1977) made the interesting observation that the common occurrence of crinoids, bryozoa and ammonites at the upper surface of Permo-Pennsylvanian debris deposits on Ellesmere Island may represent an indigenous fauna inhabiting the "reef-like" upper surface of the deposit.

The polymict nature of the clasts reflects the complexity of the source area, namely the platform margin, which consists of partly lithified reefs and/or limesand shoals, downslope (yet still shallow) reef mounds, or peri-platform talus. Among the talus blocks, expected types include well-sorted and well-bedded lime sands which can be differentially submarine cemented (Fig. 6), individual colonies of reef builders, multigeneration reef rock, limestones with subaerial karst features, tidal flat lithologies, and even cemented talus that has been refractured to give breccia clasts within breccia.

The fabrics of "analagous" coarse siliciclastic deposits have been discussed by Walker (1976). In lime breccias they range from mainly chaotic to imbricated to horizontal to vertical, and rarely include graded or even reverse graded bedding. They range from clast-supported to most commonly matrix-supported with extremely poor sorting. The matrix ranges from shale to argil-

Figure 3
Peri-platform talus; looking across the steeply dipping fore-reef slope at a depth of 130m seaward of the Belize barrier reef complex. The slope is composed of blocks of limestone, plates of coral and lime sand composed of the plates of the green alga Halimeda; *the small block at the center (arrow) is about one metre high.*

Figure 4
Lime breccias; light grey, shallow-water reef-derived limestone breccias occurring in a 'channel' - a (approximately 8m thick); sheets - b (approximately 2m thick); and irregular masses - c (up to 12m thick), all enclosed in thin-bedded, dark grey, peri-platform lime mudstones, Cathedral Formation (Middle Cambrian), southface Mt. Field, British Columbia.

Figure 5
A sequence illustrating two different types of carbonate slope deposits; debris flows with large limestone clasts (right) and thin-bedded, graded calcarenites (the thin, grey limestone beds), interbedded with black fissile shale. This overturned sequence (top at lower left) of Middle Ordovician age occurs at Cape Cormorant, Port-au-Port Peninsula, Western Newfoundland.

laceous lime mud to lime mudstone with occasional lime sand. As Hopkins (1977) points out, however, what is often taken to be lime mud in outcrop turns out to be peloid lime sand in thin section, so that sand-sized matrix may be more common than supposed.

The exact mechanisms by which these sediments are transported are not yet clear. Submarine debris flows (Hampton, 1972) are sediment gravity flows in which granular solids such as boulders, pebbles and sand are more or less "floated" during transport by the yield strength of the matrix which is composed of interstitial fluid and fine sediment. Buoyancy of the fluid matrix also contributes to the support. Because not all such deposits have a clay mineral matrix, the transport mechanism is thought to be a combination of debris flow and grain flow (Middleton and Hampton, 1973). A major problem in this regard is that almost all experimental work to date has been done on clay-water mixtures; none of the experiments has been carried out on sediments with a clay-lime or lime mud matrix.

B) Breccias Derived from the Slope. The evenly-bedded calcilutites or lime muds of the slope facies are often prone to downslope creep. Individual beds can be seen to neck or wedge out, or whole intervals will move downslope within a series of slump folds (Fig. 7). Dislocation and movement of large masses of slope material downslope leads to the formation of breccias or submarine glide masses composed of numerous tabular clasts of slope limestone that have been bent or fractured, that are poorly-sorted and that exhibit random to subparallel orientations, often resembling shallow-water "flat pebble conglomerates" (Fig. 8). Enormous blocks of bedded slope sediments, often internally folded, are caught up in the breccias.

The source of these breccias is thought to be the large "intraformational truncation surfaces" (Fig. 9) or "cut-and-fill structures" (Wilson, 1969) which are sharp concave-up discontinuity surfaces that truncate underlying beds and are overlain by a downslope thickening wedge of sediment with an angular relationship on the truncated beds. In these deposits, reduction of shear stress occurs by displacement of

Figure 6
Thinly-bedded, nodular foreslope sequence comprising cemented nodules in compacted calcarenite (N) and a laterally continuous bed of cemented calcarenite (S). These calcarenites form the predominant foreslope facies below the Miette and Ancient Wall buildups (Upper Devonian), Alberta. Photo courtesy of J.C. Hopkins.

Figure 7
Extensive syn-sedimentary distortion of bedding developed by creep in thin to very thin-bedded, upper basinal slope, peri-platform lime mudstones, Eldon Formation (Middle Cambrian), Wapta Mountain, British Columbia.

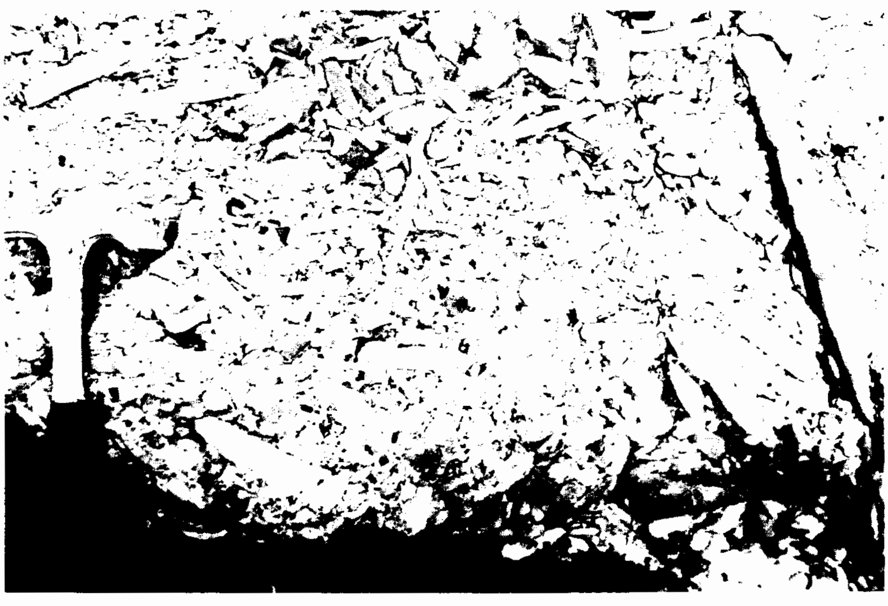

Figure 8
Slope-derived breccia; clasts of partly lithified peri-platform ooze (see Fig. 1) that have been eroded and transported as a clast-supported breccia, Cooks Brook Formation (Middle Cambrian), Humber Arm, Western Newfoundland.

coherent masses along discrete shear planes and not usually by deformation within the mass as occurs in slumps.

The tabular clasts of slope material, although derived by separation along bedding planes, clearly indicate that the slope sediments were partly consolidated very early, probably by submarine cementation. Cementation may have been similar to that in shallow-water with lithified and unlithified layers reflecting times of slow and rapid sedimentation respectively. If cementation and neomorphism took place below the thermocline, dissolution of aragonite and possible precipitation of calcite may have caused the same effect in layers of different original composition (Schlager and James, 1978). Alternatively, if the lime mudstone is interlaminated with shale, cementation of the carbonate may have taken place while the shale remained soft.

Deposits of the two end members, one originating high on the slope and the other down on the lower slope are sometimes found intermixed in extensive breccia masses (Fig. 10). Such deposits are similar to what Schlager and Schlager (1973) term marl-flaser breccia, characterized by a chaotic fabric of plastically deformed, dark grey, argillaceous lime mudstone lithoclasts separating irregular lenses of subangular limestone and other lithoclasts, with the deformed marls forming the flaser fabric. There are thought to be shallow-water derived breccia flows that incorporated lime mudstone clasts from the floor of the slope environment as they moved basinward and they may grade downslope into turbidites.

Graded Calcarenites

A large proportion of any slope sequence commonly consists of size-graded beds of clastic textured limestone, mainly of sand size, interpreted to be the carbonate equivalent of siliciclastic turbidites (Fig. 5). They are envisaged to be deposited from turbidity currents that formed by the sudden surge-type release of dense fluid rather than from a steady state flow such as described by Harms (1974). These deposits have also been called allodapic limestones (Meischner, 1964). Such sediments are well-bedded and characteristically have sharp planar bases that can be coplanar with, or locally scour and truncate underlying

Figure 9
Large intraformational truncation surface in agrillaceous and cherty limestones of the Hare Fiord Formation (Permo-Pennsylvanian), north side of Svartfjeld Peninsula, Ellesmere Island. Note smooth, curved concave-up (listric) geometry of the truncation surface and the lack of macro-scale deformation of beds below or above truncation surface. Shadow at lower left centre is of heliocopter: width of view 150m. Photo courtesy of G.R. Davies.

Figure 11
A bed of light grey calcarenite composed of a graded lower portion, planar laminated middle unit, and the upper portion having climbing ripples (A,B and C Bouma subdivisions respectively), capping a lime breccia, Sekwi Formation (Lower Cambrian), Cariboo Pass, Mackenzie Mountains. Photo courtesy of F.F. Krause.

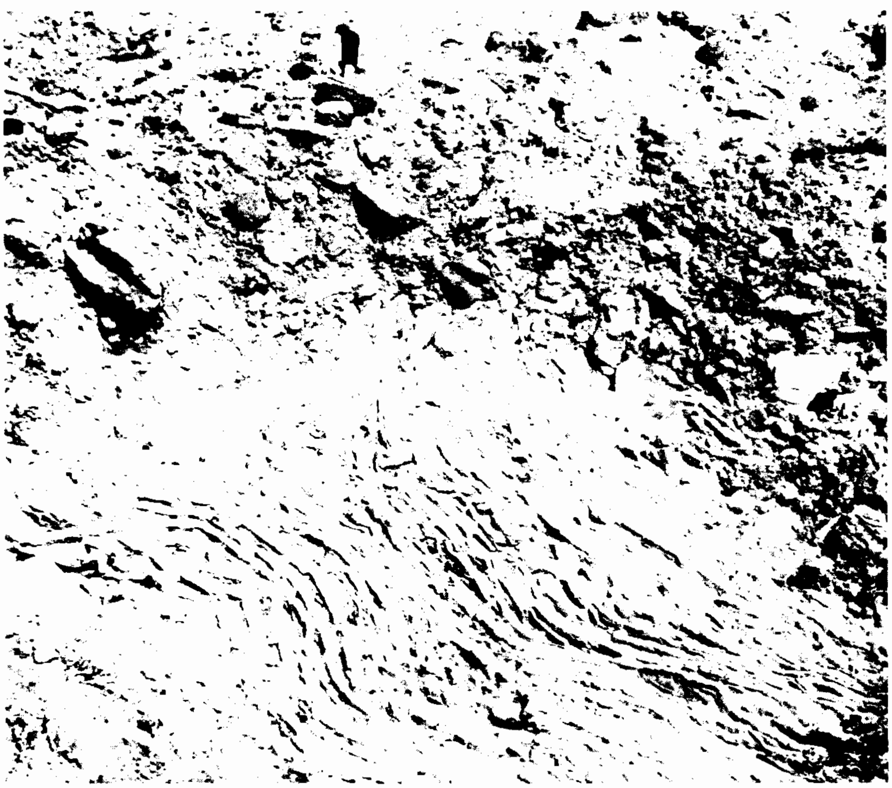

Figure 10
A large deformed clast of well-bedded peri-platform ooze that was eroded, transported and redeposited as part of a debris flow, Cow Head Group (Middle Ordovician), Cow Head, Western Newfoundland.

slope beds. Sole marks and load structures are usually absent although in some cases they may be obliterated because of stylolitization and solution along bedding contacts. Calcareous turbidites can exhibit all five of the typical ABCDE division of the Bouma sequence but most commonly it is the A, and sometimes the B and C divisions that characterize the deposits (Fig. 11). The particles in the basal parts of division A are often cobble size and larger and the more common grain types are lithoclasts, skeletal debris and ooids, the petrology of which indicates a shallow water origin in contrast to the surrounding pelagic deposits.

The most obvious sources for these units are the unstable accumulations of lime sand and gravel that build up near the platform margin and are occasionally set into motion. It is also possible that they are the distal parts of carbonate debris flows representing the uppermost more dilute turbulent portions of the debris flow (Krause and Oldershaw, 1979). Davies (1977) has suggested a third origin, that skeletal material produced on the slope particularly by pelmatozoans may be easily remobilized.

Post-Paleozoic graded calcarenites derived from sediments further down the slope profile can be virtually indistinguishable compositionally from pelagic limestone. These calcarenites are generally rich in pelagic components such as coccoliths and foraminifers but may also contain lesser amounts of pteropods, sponge spicules, radiolarians, and coarser-grained skeletal debris (especially pelmatozoans). The sediments are size-sorted and may be mixed with clastic terrigenous or volcaniclastic sediment if they have travelled great distances. Although the sedimentary structures such as horizontal laminations, convolutions, occasional channels, flute and groove casts and trace fossils may be present, the A and B divisions of the Bouma sequence are commonly missing and they generally start with the C or D divisions.

Turbidity currents and debris flows appear to be the dominant transport mechanisms for the downslope movement of coarse detritus in modern carbonate slopes (Cook and Mullins, 1983; Enos and Moore, 1983).

Non-Graded Calcarenites

Massive to cross-bedded and ripple-marked calcarenites are an enigmatic type of deposit found in many slope sequences. These deposits are fine- to coarse-grained wackestones to grainstones with occasional large clasts or fossils. Individual beds have sharp bases and vary in geometry from lenticular to irregular masses. The fabric may be random or grains may be aligned parallel to the paleoslope.

The grains in these deposits are variable, ranging in composition from shallow-water derived particles to pelagic grains.

Non-graded calcarenites occurring below a carbonate margin of considerable slope and having an abundant supply of lime sand may have formed from grain flows or more likely, through a modified grain flow mechanism whereby the addition of lime mud matrix and turbulence may have aided dispersive pressures in supporting the grains during transport (Lowe, 1976). It is unlikely that liquified flow, or fluidized flow contribute significantly to the formation of these particular slope calcarenites. Another viable mechanism is reworking of previously deposited slope sediments by bottom currents. Perhaps the massive deposits having an apparent lack of sedimentary structures are nothing more than the product of downslope mass movement of well-sorted lime sands produced at a rapid rate near the platform margin such as occurs on modern leeward open margins during large storms (Hine *et al.*, 1981).

Sedimentary structures in the cross-bedded deposits indicate some sort of bottom currents, often running parallel to the slope (contour currents). Well-sorted, rippled ooid lime sands, sometimes with large scale bed forms, occur in the deeper parts of the slopes around the margins of the Tongue of the Ocean, Bahamas, and are also common on the slopes along the western parts of the Bahama Banks (Mullins and Neumann, 1979) where currents flow along and parallel to the slope at speeds of 60 cm/sec and more (although such velocities are high and not characteristic of today's oceans). These currents may rework bank-derived sands (Hine and Neumann, 1977) or rework pre-existing pelagic slope deposits, leaving only the larger foraminifers and pteropods together with lithoclasts of cemented pelagics to form a deep-water grainstone. They may also winnow the upper parts of turbidites, removing the finer layers and leaving a sequence composed only of shallow-water clasts, and divisions A and B of the Bouma sequence, capped by a cross-bedded lime sand.

Such clean, well-sorted sands are commonly sites of submarine cementation and hardground formation. In such areas precipitation of cement may lead to displacive expansion of grain-to-grain distance, resulting in fracturing and the formation of *in situ* breccias.

THE DEPOSITONAL SETTING

Even though there may be a continuous rain of pelagic sediment on the slope, most slope sediments come from the platform, as re-sedimented gravity flows or peri-platform ooze. Thus, the nature of slope deposition closely reflects events on the platform, or more specifically whether the platform is close to sea level, drowned or exposed (Fig. 12). If the platform surface is close to sea level, then there is a constant source of carbonate sediment being produced and so slope as well as adjacent basin deposits are correspondingly abundant. If sea level drops, the platform is exposed to meteoric diagenesis and the platform sediments are turned to limestone, thus shutting off the supply of sediment to the slope. Starved deep-water sedimentation can also reflect rapid sea level rise. If carbonate production is outpaced, a tranquil deep-water shelf consisting primarily of muddy, skeletal carbonates can develop.

The style of shelf-slope break in the fossil record also reflects the interaction between rates of carbonate production and relative sea-level movement (James and Mountjoy, 1983; Bossellini, 1984), the latter being due to the combined effect of subsidence (tectonic controls) and sea level fluctuations (Schlager and Ginsburg, 1981). If the rate of carbonate accretion is more or less equal to the rate of relative sea level rise, the shelf break will remain more or less in the same position and ultimately the relative relief will increase between the margin and the basin. When relative sea level rise is less than carbonate accretion, the slope facies prograde out over older slope deposits, and will consist of thick accumulations of many re-sedimented lime sands and conglomerates. Opposite circumstances create onlap margins which can be drowned or will move shelfwards in steps. In this situation, little sediment is transported seaward so slope and basin deposits are thin and mostly carbonate muds. Drowned and emergent margins due to large changes in relative sea level result ultimately in starved basinal sedimentation.

In the drowned situation, the style of carbonate shelf is close to what Ahr (1973) has termed a "carbonate ramp", where the high energy zone is the shoreline and the carbonate facies pass progressively into deeper water lithologies. Ginsburg and James (1974) have called these "open shelves" and good modern examples are present on the Yucatan Shelf, West Florida Shelf and in the Persian Gulf. If the ramp has a marked increase in slope at the seaward edge, slope facies will have abundant slumps, slope breccias and turbidites. This situation is termed by Read (1982) a "distally steepened ramp". The absence of a slope break results in a lack of significant slump and sediment gravity flow deposits in the deeper water facies and the formation of a "homoclinal ramp".

The nature of slope facies is also

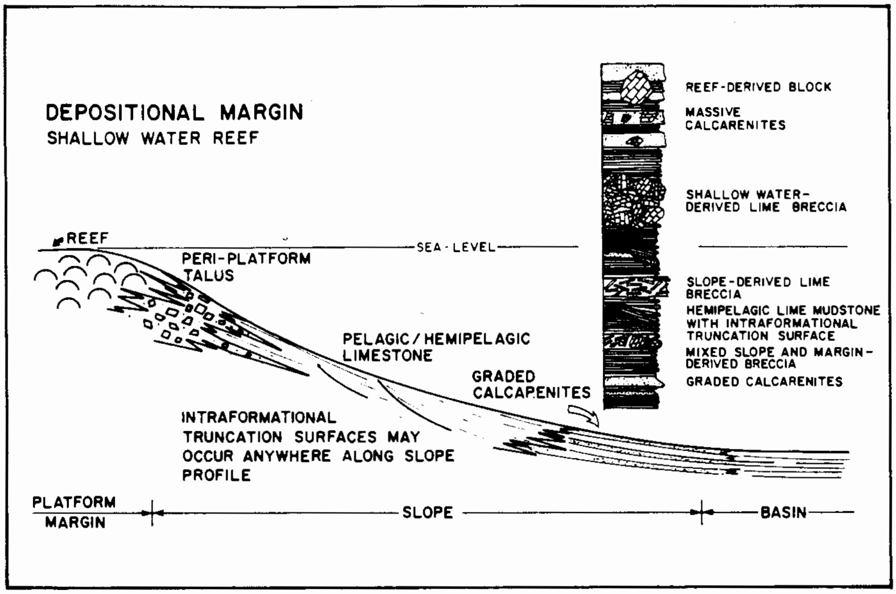

Figure 12
Variations in the style of slope sedimentation as a function of water depth on an adjacent carbonate platform.

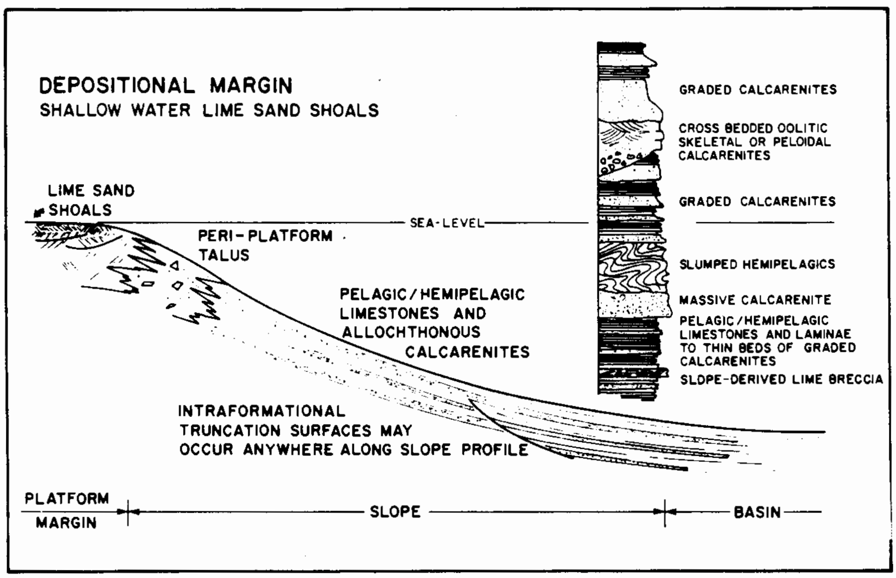

Figure 13
Schematic model for a shallow-water, reef dominated, depositional carbonate margin and illustration of a hypothetical sequence of deposits within the adjacent basin slope.

dependent on the direction and magnitude of off-shelf sediment transport resulting from the net effects of waves, storms and tides. Antecedent topography in the form of islands or subtidal rock ridges create energy barriers which can control the volume of sediment flux on or off the shelf. Oceanic circulation (bottom currents) can modify, through physical transport and erosion, pre-existing slope sediments as well as promote submarine cementation and hardground development on the slopes. For additional details on these parameters and how they affect slope sedimentation along the Northern Bahamas, the reader is referred to Hine *et al.* (1981) and Mullins and Neumann (1979).

In discussing the topic of sea level fluctuation and slope deposition, Schlager and Ginsburg (1981) make the important point that in siliciclastic systems, a lower sea level results in increased erosion and delivery of terrigenous material to the deep; the reverse is true in carbonate systems.

Carbonate slope sedimentation also differs from that adjacent to terrigenous shelves in another important way. Carbonate sand is delivered all along the platform margin. Deep sea fans are absent because no submarine dendritic drainage systems are developed that would funnel sediment into canyons. Instead, the reefs or sand shoals of the platform act as a line source, creating a continuous belt of overlapping turbidites and gravity-flow deposits at the toe-of-slope (Schlager and Chermak, 1979); the resulting deposit is more appropriately termed a debris apron (McIlreath, 1977b; Mullins *et al.*, 1984) which has a geometry that is distinct from a fan.

Carbonate sedimentation providing slope stability results in another distinction from terrigenous shelf-basins. Massive failure of a partly lithified carbonate slope will reduce the possibility of the simple evolution of a single flow (slump → debris flow → turbidity current) as postulated for siliciclastics (Middleton and Hampton, 1973).

FACIES MODELS

Because of the preceding variables we have chosen not to integrate the spectrum of carbonate slope deposits into one model. Instead, we have chosen to model carbonate slope sedimentation adjacent to a platform near sea level. In this situation, the style of slope and basinal sedimentation is dependent upon the relief between platform and basin, and the nature of the shallow portion of the margin. Where the margin itself is a facies transition with a gradual slope profile, then the sequence of slope deposits is very much different from the sequence where the margin is arupt. We have differentiated between these two types of margins and called them depositional and by-pass margins respectively (Fig. 12 to 16)

The nature of the slope sediments in each case also depends on whether the shallow-water margin is formed: 1) by metazoan, calcareous algal or stromatolitic reefs, and occurs either at the edge or slightly downslope below the zone of the most wave movement, or 2) by skeletal algal or oolitic lime-sand shoals. In the case of reefs, the nature of the reef-building biota changes with the

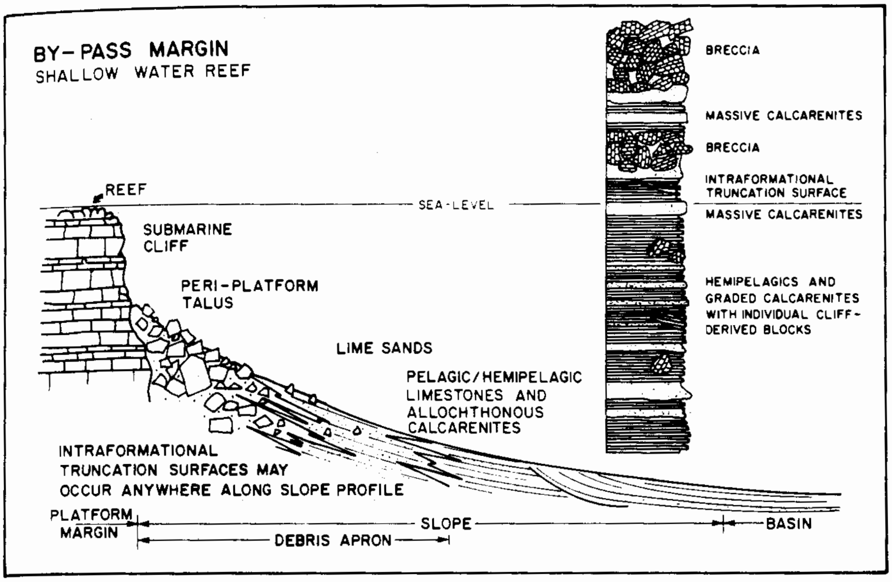

Figure 14
Schematic model for a depositional carbonate margin dominated by shallow-water lime sands and illustration of a hypothetical sequence of adjacent basinal slope deposits.

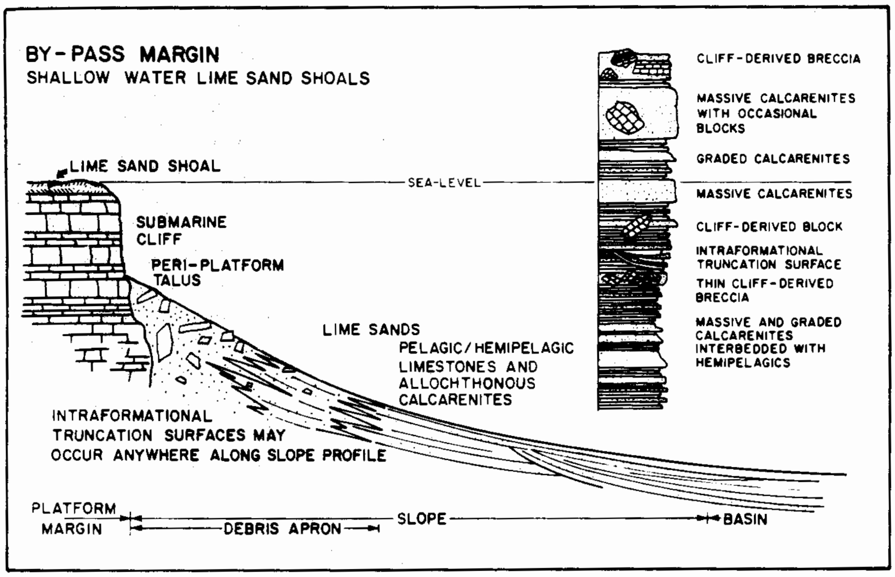

Figure 15
Schematic model for a shallow-water, reef-dominated, by-pass type of carbonate margin in a shallow-basin and illustration of a hypothetical sequence of deposits within the adjacent basin slope. In a deep basin there is an extensive by-pass slope below the peri-platform talus. Debris, including turbidites is funnelled through gullies onto the basin floor.

time and therefore the composition and nature of the resulting debris correspondingly changes.

It should be noted that none of the models is mutually exclusive and within a buildup or platform margin all four (depositional reef, depositional shoal, by-pass reef, by-pass shoal) may be present an any one time. They may be repeated in time and space; in the case of a buildup, it may even be possible to have all four occurring simultaneously in different places along the buildup margin. It is quite common in ancient platforms to progress from a ramp or depositional margin to a by-pass type and even to an erosional margin.

Depositional Margins
The slopes are generally gentle and decrease basinward to merge with the flat basin floor. Because ramps have even lower gradients, slope sediments are normally much finer and consist of pelagic carbonates and shales with minor slumping and slide development.

A) Shallow-Water Reef. The zone of peri-platform talus is relatively narrow but the full spectrum of allochthonous deposits is present downslope (Fig. 12). Because most of the allochthonous material comes from the reef or talus pile many of the allochthonous deposits generated high on the slope are deposited far down on the slope or in the basin. Consequently that zone seaward of the peri-platform talus is often composed of hemipelagic limestones and is often by-passed by the mass movements. This type of depositional slope occurs most frequently around reef complexes and basinward of platform-margin barrier reef systems along paleotopographic highs, structurally positive elements or hingelines in fairly stable cratonic or miogeosynclinal basins. Examples of this style of slope deposit occur in the Upper Devonian and the Cambrian of Western Canada (McIlreath, 1977a) and the Devonian of the Canning Basin, Australia (Conaghan *et al.*, 1976).

B) Shallow-Water Lime Sands. The slope flanking this style of margin is generally a calcarenite wedge of proximal-to-distal turbidite plain (Fig. 13). These slopes probably represent a depositional equilibrium in that sedimentation controls the slope angle and is active all along the profile. Turbidity currents and grain flows are the predominant transport mechanisms, and debris sheets and breccias rare. Some minor debris sheets composed of cemented lime-sand clasts or other slope-derived lithologies may be present. Hardgrounds and incipient brecciation are common.

Examples of this style of slope deposit includes the Pennsylvanian Dimple Limestone, Texas (Thomson and Thomasson, 1969); Silurian of California and Nevada (Ross, 1965); several of the Devonian Fairholme carbonate complexes of Western Canada, and Devonian encrinite banks in Arctic Canada.

By-Pass Margins
In these situations the magin is on top of a cliff or submarine escarpment so

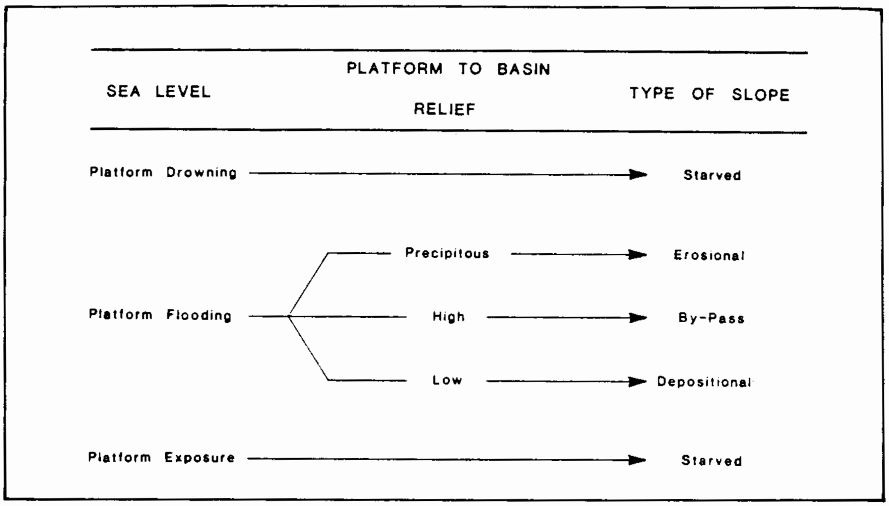

Figure 16
Schematic model for a by-pass type of carbonate margin dominated by shallow-water lime sands in a shallow-basin, and illustration of a hypothetical sequence of adjacent basinal slope deposits.

that sediments are transported directly from shallow to deep-water. They may bypass much of the slope along a wide front, or be funnelled through channels and canyons to accumulate at the toe-of-slope and adjacent basin. In shallower basins the submarine cliff is the actual by-pass slope, below which accumulates a debris apron of coarse peri-platform talus fining basinwards into peri-platform and pelagic oozes with occasional turbidites. Flows are triggered on the upper slope by oversteepening of peri-platform talus, collapse of the reef wall or by slumping of previously deposited sediments, and are seen to erode gullies on modern slopes (Schlager *et al.*, 1976). The cliffs may result from faulting, large fluctuations in sea-level or just rapid upbuilding of the platform as compared to the basinal deposits. This style of margin is particularly common along block-faulted oceanic margins or at the structural hingeline where a basin is subsiding faster than the adjacent platform.

A) Shallow-Water Reef. Since the reef crowns the escarpment, the most characteristic and spectacular style of accumulation is the wedge of peri-platform talus (Fig. 14). This wedge of material may be enormous, especially if the area is subject to tectonics, with the main transport mechanisms being a combination of rock-fall, sandstreams and gravity-induced downslope mass movement. If the cliff is dissected by channels or canyons, the peri-platform talus may interdigitate along strike with carbonate submarine fans similar to those described for siliciclastic deposits (Evans and Kendall, 1977). Slumps creep and sliding are more active in the deposit than on the adjacent slope due to the variations in lithification. The talus wedge grades downslope into a relatively narrow zone of lime sands and then into pelagic calcilutites to form a debris apron.

This is the style of many modern slope deposits in Belize (Ginsburg and James, 1973), Puerto Rico (Conolly and Ewing, 1967), Jamaica (Goreau and Land, 1974), the Bahamas (Mullins and Newmann, 1979; Schlager and Chermak, 1979), and the Pacific Atolls (Emery *et al.*, 1954). Perhaps the most spectacular fossil example is in the Cretaceous of Mexico (Enos, 1977). Other fossil examples include the Cambrian Cathedral Escarpment, Western Canada, and Upper Devonian margins in the Front Ranges of Western Canada.

B) Lime Sand Shoals. If the shallow-water margin facies is lime sand, the peri-platform talus will also consist predominantly of lime sand intercalated with calcilutites (Fig. 15) with fewer limestone blocks than in the previous model (unless there have been substantial movements in sea level exposing the margin to subaerial diagenesis). Away from the escarpment the lime sands grade relatively quickly into slope or basinal pelagic lime muds. There are minor contributions from turbidites.

A fossil example of such a debris apron of calcarenite is the Cambrian Boundary Limestone (McIlreath, 1977b). Modern examples have been found in the northern Bahamian slopes (Mullins and Newmann, 1979).

The Models as a Norm

In these models we have not consciously placed the slope lithologies in any particular sequence because we feel that the sequence on such a broad scale represents more the complex interactions of sea level and tectonics at the shallow rim than any secondary sedimentary process on the slope. As a result unusual features are likely to record not so much the style of sedimentation on the slope, as the style of sedimentation and tectonics at the shallow margin.

The Models as a Framework and Guide for Description

The differences in carbonate deposition through time are, in large part, a function of the appearance and disappearance of different types of carbonate secreting organisms and thus affect the use of these facies models as a guide in two ways: 1) shallow-water benthic organisms build massive reefs, cause relief at the platform margin, and contribute major amounts of sediment to the slope; however, they are present only at specific times in geologic history; 2) the pelagic calcareous zooplankton and phytoplankton are insignificant in the early Paleozoic, minor in the middle and late Paleozoic, and prolific in the Mesozoic and Tertiary.

As a result, the hemipelagic slope deposition is almost entirely peri-platform ooze in the Precambrian and early Paleozoic, and perhaps one-half peri-platform ooze and one-half true pelagic carbonate in the Mesozoic and Tertiary. Interruptions in the fallout of peri-platform ooze in the Paleozoic sometimes resulted in shale interbeds, whereas in the Mesozoic and Tertiary interbeds are thinner but are wholly pelagic carbonate.

The Models as Predictors in New Situations

Based on a few observations, and bearing in mind the age of the deposits as well as their tectonic setting, one can extrapolate and formulate three critical conclusions: 1) examination of the overall sequence indicates the relative position on the slope and possible proximity of the platform; 2) the lithology of the lime-sand beds and relative calcarenite to hemipelagic ratio gives some idea as to the nature of the slope facies; depositional versus by-pass; and 3) the composition of the clasts indicates the nature of the margin, which has often been obliterated or is inaccessible.

The Models as Basis for Interpretation

The interpretation of carbonate sediment gravity flows has, to date, been based primarily on an analogy with siliciclastic deposits which have similar sedimentary characteristics. One of the important differences between carbonate and siliciclastic sediment gravity flows, however, is that a dispersal model for the hypothetical evolution of a single flow of carbonate platform-derived debris into deep water remains speculative. In contrast to the relatively unconsolidated sediments on continental shelves, carbonate sediments in similar environments tend to be stabilized by organisms and/or well-lithified. This results in distinctively different slope deposits being produced by a variety of gravity-driven transport processes rather than different types of deposits evolving from the same flow. Unequivocal examples of an ancient deposit evolving from the same flow are rare. It should be noted, however, that the concept of a singular flow spawning a series of deposits may apply where slide failure occurs in the lower portion of the slope, remobilizing and transporting these mixed deposits even further basinward.

SUMMARY

Carbonate slope sediments have, in the past, often been either ignored or interpreted as tectonic in origin. Their identification as deposits, separate from tectonically formed mélanges has come largely from a detailed analysis of not only the chaotic deposits but the fine-grained interbeds as well. Our understanding is increasing as more deposits are documented and the first timid steps are being taken beyond the reef into deeper water by submersibles. This latter aspect of carbonate sedimentology is still very much in its infancy.

Refinements of the models presented in this paper must come from two directions, experimentation and more observations from modern carbonate slope environments (in addition to those from current studies on Bahamian slopes). The hydrodynamic parameters and processes for gravity-induced mass movements involving only carbonate materials must be carefully documented and contrasted with the results from siliciclastic materials. A combination of detailed observations from submersibles, and high resolution seismic and bottom sampling is needed to make an inventory of the spectrum of sediments and structures that make up carbonate slope environments in the modern ocean.

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