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COMPACTION IN LIMESTONES: A REAPPRAISAL

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Abstract

An important outcome of recent research on carbonate diagenesis has been the increasing realization that compaction may be as important as cementation in the lithification of carbonate sediments. Absence of compaction-deformation features should no longer be relied upon as unique evidence against compaction in limestones. The paucity of such features arises from the fact that the load-bearing capacity of allochems (grains) is hardly exceeded in the normal course of lithification of carbonate sediments. Their load-bearing capacity seems to have been exceeded only in the presence of rigid bodies (nodules) or surfaces (hardgrounds or emersion surfaces). Wherever this prerequisite is met, compaction-deformation may be set into motion in the adjacent sediments, irrespective of their environments of deposition.

So far as the mud-supported carbonate sediments are concerned, they may undergo autolithification entirely through solution-reprecipitation without the benefit of subaerial exposure. In the case of grain-supported carbonates too, there are indications *that an appreciable amount of their cement may be derived from elsewhere in adjacent* rocks undergoing deeper burial diagenesis within the same basin. While burial diagenesis seems normal, subaerial diagenesis is exceptional for carbonate sediments, *other than those of the nearshore areas.*

INTRODUCTION

Past few years have witnessed a renewed interest in compaction of carbonate sediments. Since the appearance of Pray's (1960) abstract, the idea of noncompaction of carbonate sediments had been the guiding principle in carbonate sedimentology till recently. It had been the general belief that initial porosity of carbonate sediments is eliminated by cementation rather than compaction, for there is little evidence of compaction in limestone (Friedman, 1975).

Today, in retrospect, it seems that the simple fallacy in the non-compaction argument was somehow overlooked. In the absence of evidence of large-scale seafloor cementation and with imbued prejudice against compaction, most workers had to accede to the primacy of subaerial diagenesis (Friedman, 1975). Curious though it may sound, in order to cement a body of carbonate sediment there was no alternative other than the destruction of limestone equal to half its volume (Bathurst, 1975); and to do so, we required the sea-floor to bounce up and down time and again (Friedman, 1975). Recent experimental and observational data indicate that compaction may be as important as cementation in the lithification of carbonate sediments.

The purpose of the present critique is to attempt an appraisal of the question of compaction in the light of recent research. It does not purport to be a historical review nor does it intend to offer solutions to all existing problems. Compaction, as used here, relates to a reduction in bulk volume of a sediment which is expressed as a decrease in porosity brought about by tighter packing of sediment particles. It may involve grain displacement, grain deformation or solution of grains at point contacts. It ceases when pores are closed or when the load borne by pore fluids decreases to zero and overburden stress is directed entirely to a grain framework which does not yield.

PERSPECTIVE ON COMPACTION

Compaction due to overburden during the initial stages of burial simply leads to dewatering, *i.e.*, volume reduction without any internal change other than closer packing of grains and orientation of elongate grains. External change caused by dewatering mostly does not cause any deformation of primary stratification (except for local water injection structures) or of the internal constituents.

Following dewatering, the grains start to respond either through intergranular pressure solution or by mechanical deformation, depending on the relationship of the critical stress for solubility to the yield stress of the grains in the ambient stress environment. If the critical stress for solubility is exceeded first, the grain will undergo pressure solution; otherwise, it would deform mechanically. When a grain undergoes pressure solution, it shows no evidence of deformation but rather shows indications that parts of it have been dissolved without disturbing the rest. The possibility of grain deformation and intergranular pressure solution ceases with the achievement of a rigid framework by cementation. Pressure solution may, however, continue and be manifested as stylolites cutting across the whole rock, in contrast to grain to grain sutured contacts in unlithified or semilithified sediments.

The relationship between the degree of dissolution at grain-to-grain contacts and the depth of burial is by no means straightforward. Even though pressure solution would appear to be enhanced by increasing depth of burial, the ambient environment as determined by mineralogy, texture, pore water composition, temperature gradient, residence time at depth and presence of clay etc., strongly influence intergranular pressure solution. Hence, given favourable condition, pressure solution may operate even at extremely shallow depths, and it is almost impossible to figure out the depth at which grain-to-grain dissolution may occur (Purdy, 1968). So far as these complicating factors are concerned, the situation under deep burial conditions seems more predictable than that under shallow burial, and as DSDP data suggest, pervasive intergranular pressure solution in pelagic carbonates possibly begins to operate at depths around 200 m (Schlanger and Douglas, 1974), a figure close to that given by Neugbauer (1974) on the basis of theoretical analysis. Apparently, under deep burial condition, overburden pressure completely masks the effect of other subordinate factors.

A grain undergoing deformaton under overburden stress may do so either by brittle or plastic (as for ooids and pellets) failure. Grain reorientation normally accompanies plastic deformation. Grains that were not originally parallel to bedding may become so aligned without the benefit of plastic deformation. It may, however, be difficult to visually distinguish a depositional fabric from a compactional one, where grains are elongate (see Pl. II A, Wilson, 1975, for example). Orientation resulting from deformational flattening of spherical bodies such as ooids is, however, easily discernible. Longer axes of deformed ellipsoidal grains in such instances tend to lie parallel to bedding plane—a plane of no strain. Change in packing accompanying compaction is hard to determine except where original packing is locked in precompaction concretions or nodules. It seems, however, quite probable that a mud-supported fabric may be compacted to a grain-supported one (Dunham, 1962) through differential pressure solution of the supporting mud.

COMPACTION AROUND NODULES AND CONCRETIONS

Differential deformation against already rigid limestone nodules and chert nodules (and clasts) has been noted by Biggs (1957); Jeans (1973); Orme (1974);

Jenkyns (1974); Garrison and Kennedy (1977) and more recently by Chanda *et al.* (1977); Hopkins (1977); and Meyers (1977). All these examples are characterized by either all or one or two of the following features:

1. differential deformation and condensation of grains in the matrix around the nodules, and lack of such deformation and condensation of fossils and other grains within the nodules,
2. deflection of bedding around the nodules, and
3. piling of stylolites or solution seams and enrichment of insoluble residue around the nodules.

Logan and Semeniuk (1976) as well as Wanless (1979), however, contend that such nodular structures in limestones may be products of post-lithification pressure solution along microstylolites. Significantly, the early cemented nodules which serve as foci of stress concentration in the prelithification stage perforce continue to do so even after lithification. When limestones are subjected to overburden or tectonic stresses, primary structural, textural, compositional and fabric discontinuities simply get overprinted (See Wanless, 1979). Logan and Semeniuk (1976), however, maintain that pressure solution involved in the generation of their 'stylo-nodular' form has little regard for minor variations in the primary fabric. The crux of the problem in the present context, however, is the distinction of pre- from post-lithification nodules. Overpacking and deformation of grains, and lack of these within the nodules seem to provide best clue to such distinction. Since grains no longer remain free to move toward each other in lithified sediments, differential packing and deformation are unlikely to occur around post-lithification nodules. Following lithification, overburden load is borne by the rock as a whole rather than by its constituents, as in unlithified sediments. Rocks subjected to overburden load yield mainly through pressure solution along stylolites and hardly betray any sign of mechanical deformation. Furthermore, post-lithification nodules are likely to cut across bedding, a feature unlikely to be associated with pre-lithification nodules; bedding rather bends around such nodules. So far as precompaction chert nodules are concerned, there is little scope for such confusion.

In this context, the nature of the deformation fabric of ooids around chert nodules in the tectonically undisturbed Precambrian shallow water Bhander Limestone, India, seems particularly instructive. Nodules of chertified oomicrite in the upper part of Bhander Limestone occurs surrounded by aureoles, elongated parallel to bedding, of plastically deformed ooids (Fig. 1). Both the intensity of deformation and the packing density of ooids in the limestone decrease away from the chert nodules until a point is reached where there is neither detectable deformation nor condensation. Significantly, unlike tectonically deformed ooids, these ooids, are inhomogeneously deformed with reentrants and apophyses, and considerably vary in shape and orientation among themselves. Lack of pervasive deformation in the limestone has been interpreted by us (Chanda, *et al.*, 1977) as suggesting that, although compaction-deformation began early, cementation commenced beyond the aureoles almost simultaneously preventing the process from affecting the sediment there. Cementation, instead of evolving in its own way, has apparently followed a path dictated by compaction. Overburden stress which deformed the ooids within aureoles must have at the same time triggered autolithification (by pressure solution and reprecipitation) beyond the aureoles, so that oolitic mud was lithified before sufficient pressure could be built up to deform the ooids. Effective

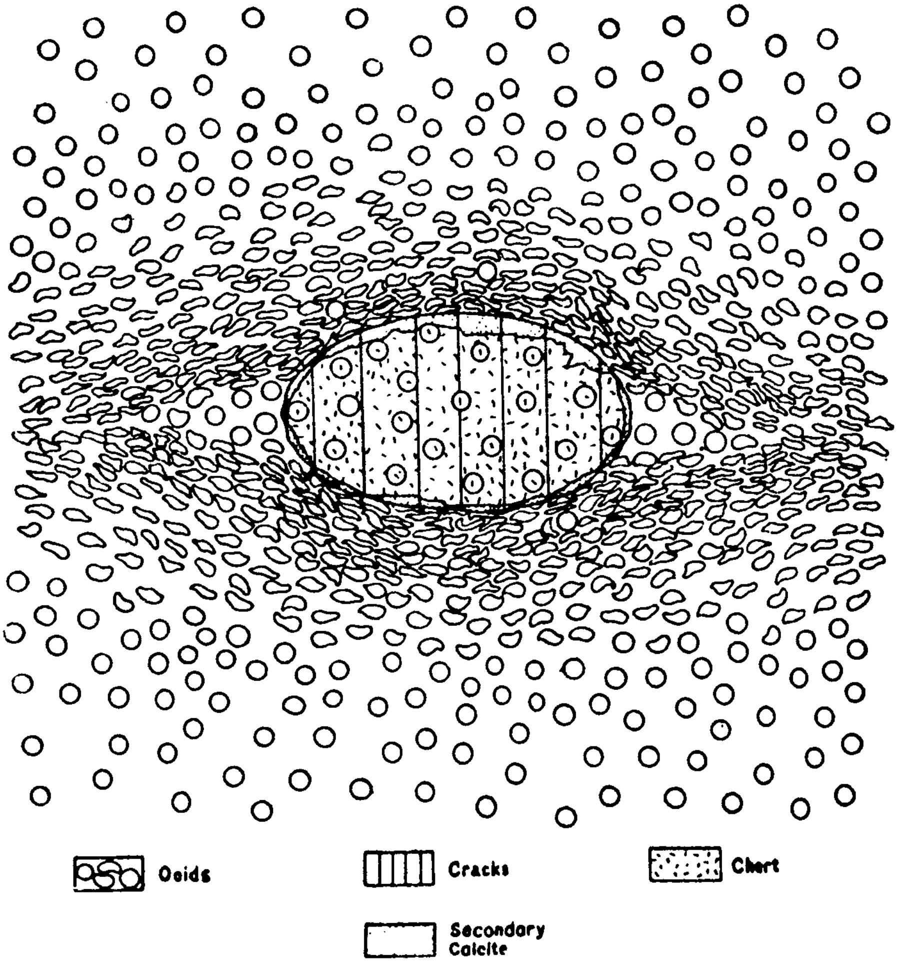


Figure 1. Schematic diagram (normal to bedding) illustrating differential and condensation of ooids in the matrix around the nodules and lack of deformation and condensation within the nodules. Note bilateral symmetry of the aureole parallel to bedding plane, pressure shadow zone immediately against the nodules in the direction of maximum extension, the decrease in the intensity of deformation and condensation of ooids away from the nodule, concentration of secondary calcite around the nodule, mutually conformable ooids in places, and the abrupt occurrence of undeformed ooids within the deformation aureole. Not to scale.

[After Chanda *et al.*, 1977. By permission of the Geological Society of America]

stress apparently exceeded the yield stress of the ooids around the nodules but away from them failed to do so though it was sufficient to maintain a level above the critical stress for solubility of mud particles so that strain was relieved by pressure solution of tiny, supersoluble carbonate mud rather than pervasive deformation of ooids. Thus cementation appears to have been an essentially intrastratal (closed system) and continuous process, occurring neither earlier nor later than compaction but acting in concert with compaction. This example further illustrates that in normal circumstances, i.e., in the absence of nodules, muddy carbonate sediments may undergo autolithification through dissolution-precipitation, for yield stress of most grains (fossils and such other elements) far exceeds the critical stress for solubility of mud-sized particles, and as such may not require introduction of allochthonous cement for lithification.

In showing evidence of compaction-deformation around the nodules, these examples clearly fall off the main trend of lithification of limestones, which normally undergo consolidation before overburden becomes great enough to cause compaction-deformation of grains. Why is there this apparent aberration?

The characteristic absence of compaction-deformation of allochems (fossils, ooids, pellets and intraclasts) in most limestones implies that transformation of carbonate sediments into a load-resistant framework is accomplished largely before the overburden stresses become high enough to deform the allochems. It would, therefore, follow that not only an early but also an abnormally rapid build-up of compressive stress is an essential prerequisite for compaction-deformation of allochems. As far as the geological record is concerned, such excess stresses are rarely developed in the normal range of lithostatic pressure likely in compaction, unless stress is somehow amplified to outpace lithification. This is an unusual condition and hence, rarely attained in the normal course of lithification of carbonate sediments. Theoretical considerations (Ghosh and Sengupta, 1973) as well as the above cited examples suggest, however, effective stress may be locally amplified in the presence of sizable precompaction rigid bodies, so that the yield stresses of allochems are exceeded around them at a much lower overburden pressure and, therefore, earlier than normal. The presence of a precompaction rigid body in an unlithified sediment reflects exactly the effects one would expect on insertion of an inclusion of different material in a plate undergoing homogeneous strain (Ghosh and Sengupta, 1973). Owing to the presence of a rigid inclusion, the homogeneous strain of the plate will be disturbed around the inclusion. The additional local stress and strain will rapidly die out away from the inclusion.

Beales (1965) also illustrated (his figs. 4-6 to 4-8) compaction effects around large lumps and megafossils in pellet limestone. Differential packing of fossils and pellets inside and outside larger shells have been illustrated by Horowitz and Potter (1971; Pl. 99, Fig. 1) and Wilson (1975, Pl. VB). Examples of such differential compaction primarily reflect differential packing within small areas in exceptionally sheltered situations (within shells or beneath grain arches) remaining in the lowest packing order. Differential packing in such situations may be enhanced to some extent by the juxtaposition of matrix around larger shells.

COMPACTION ALONG HARDGROUND AND EMERSION SURFACES

The difference between nodules and pre-burial rigid surfaces, *i.e.*, hardgrounds and emersion surfaces is, simply a matter of areal extent, that is, discrete spheres or ellipsoids *vs* laterally continuous plates. Indeed, in many limestones, nodules are often numerous enough to coalesce into more or less continuous layers. Such pre-burial rigid surfaces, similar to precompaction nodules, presumably behave as rigid plates against which unlithified carbonate sediments are pressed and a zone of deformation parallel to hardground or emersion surface should predictably result. Although not extensively documented, there are instances of layer to layer variation in the deformation and/or packing of fossils in limestone sequences, or both (Wolfe, 1968; Wachs and Hein, 1974). Similar association of deformed structures with hardground surfaces is also noted in the Upper Thammama Limestone (Cretaceous) of the Trucial Coast, Arabia (Wilson, 1975). A correlation between hardening and compactional crushing of fossils is also suspected to exist in the chalk of England by Scholle (1974).

In the light of the preceding discussion, the following comments by Purser (1978, p. 86) seem pertinent. 'The Dogger calcarenites of the Paris Basin are composed of numerous cyclic sequences most of which terminate in a hardground . . . . The oolitic and bioclastic grains within these paleocrusts are uncompacted. Similar sediments lacking early cement situated directly above or below the zone of synsedimentary lithification, on the contrary, are generally highly overpacked and stylolites are frequent'. This example is again a clear vindication of the idea developed earlier, i.e., amplification of stress along paleo-preburial rigid surfaces.

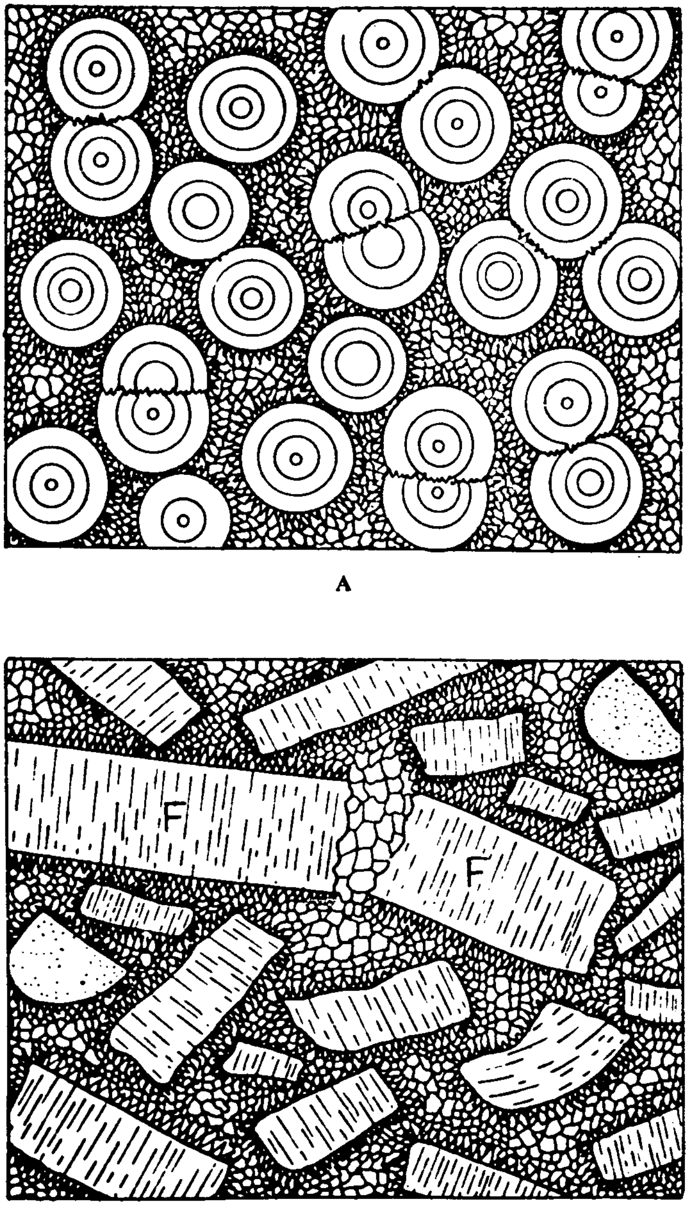
We feel tempted to quote further Purser (1978), for what he surmised with respect to the Jurassic Smackover Formation of Louisiana which is just what we foresaw in terms of our idea. Purser (1978, p. 92) extrapolates the results of his studies in Paris Basin as follows: '. . . . . . It would appear that the top of the Smackover Formation has suffered an early diagenetic lithification . . . . . . near the culmination of a carbonate sand shoal susceptible to local emergence and thus to early lithification of its crestal parts'.

Bishop (1968) in his study of the upper Smackover Limestone, Louisiana, notes closer packing of grains and intergranular microstylolitic contacts which he attributes to prelithification compaction. Coogan also noted extreme packing densities ranging from 82.5% to 95.9% in the same oolitic limestone. Such an increase in packing density, according to Coogan (1970, p. 923) 'reflects the movement of oolith grains closer together, presumably as the result primarily of increased overburden pressure'. Besides overpacking, other compactional features noted by him include spalled-off coatings of ooliths, inter-oolith pressure solution, fracturing of grains and mutually fitted boundaries of closely packed ooliths (see his figs. 10, D, E, F and G). Thus there is strong ground to suspect that the upper Smackover Formation in all probability includes an emersion surface (see Wilson, 1975) that was lithified prior to the deposition of the oolitic limestone in question, and thus provided a preburial rigid surface that hastened compaction deformation, condensation, and grain-to-grain pressure solution of ooids in the adjacent carbonate sediments. In fact, working on this idea, we have been able to locate such an emersion surface below a zone of deformed ooids in the lower part of the Bhander Limestone, India (Sarkar, *et al.*, 1980).

COMPACTION IN GRAIN-SUPPORTED LIMESTONES

As indicated earlier, prelithification compaction may cause either grain deformation or intergranular pressure solution. Grain-to-grain pressure solution operates when grains are free to move towards each other, under the influence of nonhydrostatic pressure on a grain-to-grain contact. Examples of such prelithification pressure solution are numerous (Henbest, 1968; Whitcombe, 1970; Selley, 1976). In the examples cited, pressure solution contacts between grains such as ooids and crinoidal ossicles do not extend into the surrounding cement (Fig. 2A). Such intergranular pressure solution was also reported by Powers (1962), Wilson 1975, Pl. XXI, B) and Cussey and Friedman (1977, Fig. 6). Fruth *et al.*, (1966) also illustrated similar features between ooids and fossils (see their Figs. 5e, 5f and 6a).

Besides these examples of prelithification pressure solution, some grainstones are also characterized by prelithification grain deformation too. A striking feature shown by ooliths and pisoliths in the Upper Pennsylvanian Plattsburg Limestone



A

B

Figure 2. Schematic drawings of compaction fabrics (not to scale).

A. Oosparite showing prelithification pressure solution of ooids. Note pressure solution contacts are confined within the ooids and not extending into the groundmass.

B. Biosparite showing precompaction (nonferroan) and post-compaction (ferroan) calcite cement. Note: while the nonferroan calcite is confined to the primary surface of the fossil fragment (F), the ferroan calcite overlies the former as well as the broken faces of the fossil fragments (F).

of southeastern Kansas in plastic deformation caused by compaction which preceded cementation (Kettenbrink and Manger, 1971). Some grains, of course, show fractured laminae, interpenetration and crushing. Chains of deformed pisoliths, connected by arcuate apophyses, are not uncommon. In some deformed ooids illustrated by Fruth *et al.* (1966, Fig. 5d), drusy calcite has grown in the cavities formed by differential buckling of ooid laminae, indicating that cementation followed buckling.

Strong compactional effects in Precambrian Dunoyane oolites of Spitbergen have been reported in detail by Radwanski and Birkenmajer (1977). The ooids may be pitted, cracked, snouted and distorted. The cracks are supposed to have been formed by action of mechanical force associated with pressure solution. The commonly known 'distorted ooids', according to these authors, arise from deformation of pressure-welded contacts of ooids, where the latter were pinch-and-swelled and contorted under conditions typical of sedimentary boudinage. Though they did not specify, it is implicit in their description that compaction antedated cementation, for pressure solution contacts are restricted to the ooids.

Ruptured and distorted envelopes of pellets were noted by Gustadt (1968) in the Beck Spring Dolomite (Precambrian), California. These features together with peeling and penetration of one ooid into another were attributed to compaction, apparently in prelithification stage. Beales (1965) also illustrated squashed pellets from pelleted limestone.

Whereas in these examples, compaction predated cementation, there are other examples where compaction may even outlast precipitation of early incipient cement. Grain fractures in many grainstones, characterized by more than one generation of cement, have been observed to postdate precipitation of a first generation calcite (usually nonferroan) cement crust but to precede precipitation of second generation cement (Fig. 2B). The first generation cement, in such instances, remains confined to the primary surfaces of the grains and their moldic pores, whereas secondary fractured surfaces, bare of first-generation cement crust, are directly overlain by second-generation calcite cement (Bathurst, 1975, Fg. 312). The first generation cement is commonly attributed to dissolution of aragonite. The second generation cement, on the other hand, is supposed to have been derived from pressure solution in adjacent limestone (Oldershaw and Scoffin, 1967).

The cementation history of the Corallian Beds in Southern England, as reconstructed by Talbot (1971) is most instructive in this regard. Sediments that got exposed to fresh water underwent skeletal aragonitic dissolution and were cemented by nonferroan calcite. The bulk of the sediments remained, however, insulated from fresh water diagenesis and suffered burial cementation. At shallow depths, burial cementation was manifested by precipitation of non-ferroan calcite. With further increase of overburden pressure, grain breakage and intergranular pressure solution occurred in still uncemented sediments. These were ultimately cemented by granular ferroan calcite. Some calcites, apparently the early ones, were generated by dissolution of skeletal aragonite and some from intergranular pressure solution. However, according to Talbot (1971), other sources must also have been involved.

The oolitic carbonate units of the Jurassic Twin Creek Limestone in northwestern Wyoming display textural relationships that clearly indicate dissolution of aragonite ooid nuclei prior to compaction and subsequent crushing of cortical sheaths (Wilkinson and Landing, 1978). That calcitization precedes compaction is also supported by radial fracture of presumed secondary radial fibrous fabric of ooids (Bathurst, 1975, Fig. 245). The interpretation is, however, open to question, for radial-fabric in many ancient ooids may be primary and not related to post-depositional diagenesis (Sandberg, 1975).

During early diagenesis of oolitic calcarenites in the Ste. Genevieve Limestone (Mississippian), Missouri, aragonite in fossils and ooids underwent extensive dissolution to form moldic porosity within the insoluble micritic envelopes (Knewtson and Hubert, 1969). Precipitation of calcite cement crusts around the grains prevented collapse of most micritic envelopes. Intergranular and moldic pores were subsequently filled by mosaic calcite cement. In contrast to these observations, Kendall (1975) has documented post-compactional calcitization of molluscan aragonite in a Jurassic limestone from Saskatchewan, Canada.

There are indeed certain examples of extreme compaction, where first-generation cement shards, together with distorted ooliths ond pisoliths, were cemented by ferroan dolomite (Conely, 1977) or where grain-to-grain pressure solution operated even after the emplacement of first-generation cement, so that grains though overly packed rarely touched each other (Coogan, 1970; Bathurst, 1975, Figs. 322 and 323). In the St. Brice-Villeneuve wells of the Chailly oil field (Dogger, Middle Jurassic) of the Paris Basin, France, initial porosity has been obliterated in two ways: by pressure solution which created a tightly cemented oolitic limestone lacking porosity, and by cement, generated elsewhere by pressure solution (Cussey and Friedman, 1977).

Recently Meyers (1980) has shown that reduction of porosity in the Mississippian echinoderm-bryozoan packstones and grainstones in southwestern New Mexico took place through mechanical and chemical compaction. Mechanical compaction comprising of rearrangement of grains, and plastic deformation and breakage of grains took place before and during cementation. Chemical compaction, on the other hand, is manifested by pressure solution between grains, between grains and syntaxial cement crystals, and within bryozoan grains. Chemical compaction occurred during cementation and was thus a potential source of $CaCO_3$. Here too, compaction outlasted early syntaxial cementation.

COMPACTION IN MUD-SUPPORTED LIMESTONE

As noted earlier, it was actually on the basis of observation on Mississippian calcilutites, that the idea of noncompaction was implanted in carbonate sedimentology by Pray (1960). The problems of lithification of carbonate muds was later explored by Bathurst (1970). His exploration, took off from where Pray (1960) left *i.e.*, minor compaction of calcilutites.

Surprisingly, however, just a year earlier Brown (1969) recorded only rare fracturing of pelecypod valves and whole shells in a carbonate mud matrix even after subjecting them to a compaction pressure of 15,000 psi. His observations seem to have received little attention until the recent experimental work of Shinn *et al.*, with modern muds. Compression of an undisturbed carbonate sediment (wackestone) core under a pressure of 556 kg/cm² by them produced a 'rock' with sedimentary structures similar to typical ancient fine-grained limestones (Shinn *et al.*, 1977). Fossils contained in the cores, notwithstanding compaction, however, survived almost intact. It does not appear surprising now that there were only few fractured skeletal fragments together with a significant amount of grains, retaining their original shape in the carbonate muds compacted to 1000 bars by Fruth, *et al.*, (1966, fig. 2e and 6c). Small wonder, perhaps, that on the basis of their observation Shinn *et al.*, (1977, p. 23) ventured to comment that no problem perhaps exists with regard to the question of elimination of porosity of fine-grained carbonate sediment and that relative absence of fossil breakage has caused us to reason along unproductive paths.

The question of compaction in mud-supported carbonate sediments assumes a dimension entirely different from that of grain-supported ones, for in this bicomponent system *i.e.*, grains and mud (matrix), response to overburden stress of the two components differs greatly. Intergranular mud, because of its fine size and, hence, greater pressure solubility, would apparently yield under stress more easily as well as earlier than the associated grains (allochems), which have a higher degree of organization than simple muds. Presumably, there exists a critical upper limit of overburden stress which must be exceeded before grains are mechanically deformed; this limit normally far exceeds the critical stress of solubility of tiny, supersoluble mud particles. Compaction, therefore, aids lithification of carbonate muds through not only occlusion of pores but also by generation of cement through particle-to-particle pressure solution. Autolithification, therefore, perforce is normally accomplished much before overburden stress becomes high enough to deform the grains. Autolithification is apparently achieved by mechanical compaction and by pressure welding and epitaxial cementation, these features are reflected in the development of ameboid and mosaic fabric in micrites (Fischer *et al.*, 1967).

Chalks, however, stand an exception to this model. They may withstand an overburden of 1500 m without undergoing autolithification. Resistance to autolithification apparently arises from suppression of pressure solution. This is caused by stable mineralogy of chalks (low-magnesian calcite) coupled with oversaturation of pore fluid with respect to low-magnesian calcite; oversaturation being maintained by concentration of magnesium above 0.01 M or Mg/Ca ratio from 1-2 (Neugebauer, 1974).

In recent years, the results of the JOIDES drilling programme have opened a new avenue of research into the progressive diagenesis of deep-sea carbonate sediments, a domain hitherto unexplored.

On the basis of the ocean-bottom sediments of Leg I, Beall and Fischer (1969, p. 589) reported that the alteration of nannoplankton oozes with increasing burial appear to be mainly one of compaction to harder and harder chalks and under sufficient pressure, solution welding occurs and a hard fine-grained limestone results. This initial observation has been subsequently reinforced in detail through the study of all the pelagic carbonate sections so far drilled by Glomar Challenger (Schlanger *et al.*, 1973; Schlanger and Douglas, 1974; Packham and Van der Lingen, 1973; Matter, 1974; Van der Lingen and Packham, 1975; Matter *et al.*, 1975). It has emerged that diagenesis of pelagic carbonate sediments is dependent on both the depth and the duration of burial and is generally reflected by increase in sonic velocity and density. Reversals in progressive diagenesis are, however, not uncommon, and these are related to primary sedimentary factors.

Schlanger and Douglas (1974) in their study of pelagic ooze-chalk-limestone transition in DSDP cores of Magellan Rice in the central North Pacific made the following observations (p. 121):

> There is a trend over a long stratigraphic interval, towards decreasing porosity and increasing lithification in carbonate sections; the ooze-to-chalk-to-limestone transition appears to have two stages in the reduction of porosity within a long section.
>
> 1. An early dewatering stage in which half of the total porosity reduction takes place in the upper 200 m (shallow burial realm) where porosity is

reduced from 80% to 60% in most sections. The dominant mechanism in this realm is gravitational compaction; cementation is a subordinate process.

2. A slower dewatering stage (deep burial realm) where porosity is reduced from approximately 65% to approximately 40% at a depth of 1000 m. The dominant process in this realm is cementation; gravitational compaction is the subordinate process.

The primary diagenetic mechanism was supposed to function through the solution of less stable, very small coccolith elements and walls of foraminifera and re-precipitation of calcite upon large crystals such as make up discoasters and large coccoliths. The diagenetic model given by Schlanger *et al.*, (1973) and Schlanger and Douglas (1974) is 'calcite conservative' i.e., the process operates without any introduction of carbonate from outside sources. Matter's (1974) investigation of deep-sea sediments, recovered from the Arabian sea also led him to conclude that lithification of nannoozes takes place largely by dissolution of supersoluble particles and reprecipitation of syntaxial, low-magnesian calcite overgrowth on discoasters, coccolith and micarb grains (autolithification). The large amounts of carbonate necessary to cement are not introduced from an extraneous source, but rather derived from the surrounding material. Matter *et al.*, (1975) noted further that progressive lithification is accompanied by downhole loss of strontium and δO^{18}. Significantly, compaction in pelagic carbonates is not so much by compaction-deformation of microfossil tests as by chemical compaction of particles (Van der Lingen and Packham, 1975, p. 472). This accounts for lack of compaction-deformation features in limestones despite considerable compaction that might take place during lithification of carbonate sediments.

Recently, Scholle (1977) made an exhaustive analysis of burial diagenesis in relation to chalk. According to him (Scholle, 1977, p. 995). 'The abundant evidence of cement addition to chalks, coupled with lack of evidence of extensive compaction, implies cementation is a continuous process. It starts with early (sometimes even sea-floor) precipitation of framework of cement at grain contacts (e.g. Bathurst, 1970), forming a rigid, load-resistant scaffolding into which is inserted later diagenetic cement at a rate sufficient to prevent compaction under increasing overburden'. The evidence marshalled by him in favour of burial diagenesis is as follows:

(a) consistent less of porosity with increase of depth of burial,

(b) though not marked, broken and crushed fossils tend to increase in chalk which has a deep burial history,

(c) common presence of solution seams, stylolites, grain embayments and other forms of evidences of chemical compaction,

(d) progressive addition of intrastratal cement (overgrowth cement) during burial, and

(e) increasingly negative oxygen isotopic values with increased overburden and cementation.

Despite this model of autolithification of shallow and deep-water carbonate muds, there are exceptions when their lithified products show detectable signs of grains deformation. It is plain that, in these exceptional situations, grain deformation overtakes autolithification of mud.

In the on-land chalk sequence, as noted earlier, most evidence of excessive compaction seem to be associated with hardgrounds (Wolfe, 1968) or with nodules (Garrison and Kenndey, 1977). The existence of the relationship between compaction-deformation of grains and precompaction rigid bodies, as visualized earlier, is nowhere better stated than in the following lines of Garrison and Kennedy (1977, p. 127). 'The most highly deformed grains (Fig. 16c) are those that lie at the boundaries between chalks of differing compositions or physical consistencies; the intensity of this deformation seems to have resulted from stresses imposed by differential flowage during compaction'.

Besides chalks, there are other examples of compaction-deformation of grains in muddy carbonates (Scholle, 1971, Wachs and Hein, 1974). In the case of Franciscan limestone described by Wachs and Hein (1974), laminae containing relatively well-preserved coccoliths alternate with laminae of more fragmented coccoliths. Wachs and Hein (1974) suggest that the less-fragmented layers were subjected to earlier lithifiication than the adjacent laminae containing crushed coccoliths. They further added that the most likely source of calcite cement is the micrite itself. The setting of the Franciscan limestone mimics in miniature the multiple hardgrounds, stacked over each other, of chalk sequences. It seems that effective stress was amplified along the interfaces of the lithified and unlithified lamine in the same way as it was in the case of chalk hardgrounds, and so the grains in the unlithified laminae were deformed before overburden stress by itself could become high enough to cause grain deformation.

Evidence for extensive pre-cementation compaction is widespread in Upper Cretaceous Monte Antola flysch, northern Appenines, Italy (Scholle, 1971). The remarkably small amount of total cement and very close packing of most of the grains indicate extensive compaction. Associated with these there occur grain penetration, pressure solution, breakage of fossils, and flattened burrows. According to Scholle (1971), it was due to the relatively stable mineralogy and the absence of early fresh-water flushing that compaction could outpace cementation in Monte Antola sediments. The two factors, presumed to be responsible for delayed lithification, also hold good for other analogous deep-water sediments—yet in them mechanical compaction rarely appears to be important enough to account for a significant porosity loss during the deeper burial diagenesis (Scholle, 1977, p. 993). Does the unusual compaction effects have something to do with the rigid ophiolitic basement over which they were laid, combined with an exceptionally high rate of turbidite sedimentation? Quite possibly, compaction-deformation has occurred owing to generation of excessive stress by the subjacent structurally resistant ophiolite.

MODES OF COMPACTION

Sediment response to overburden pressure is controlled evidently also by the nature of sediments, not simply by the dynamic factors related to burial. The effects of these inherent factors have been discussed in detail by Coogan and Manus (1975) and need no elaboration.

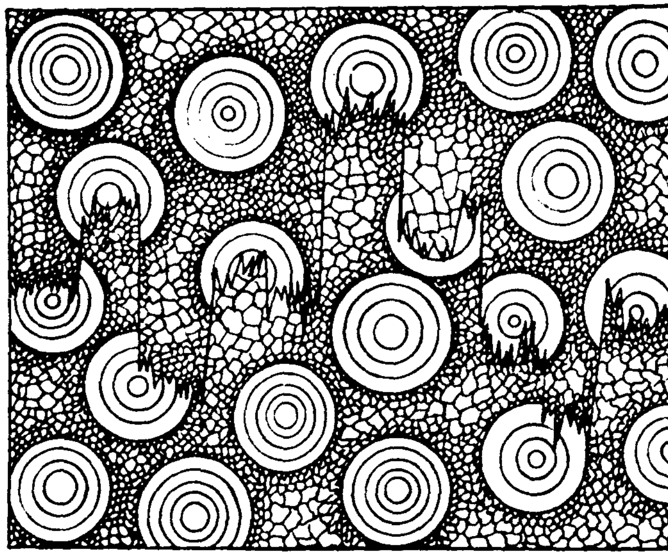
So far as the results of experimental studies are concerned, the grain/mud ratio exerts strong control on the nature of response to compaction (Fruth, *et al.*, 1968; Bhattacharyya and Friedman, 1979). In the mud-supported sediments, muds tend to cushion the effects of compaction by sharing the stress among innumerable (mud) particles, whereas in grain-supported sediments, effective stress is greater for strain is applied to fewer grain contacts. As such, grain-supported sediments are apt to show greater visible effects of campaction than mud-supported sediments.

Besides the bulk properties of the sediment, response to stress is obviously a function of the mechanical properties of the grains—ductile *vs.* nonductile. To begin with, fossil fragments are certainly brittle, compared to ooids and pellets, and as such they are prone to brittle fracture. In rare instances, they too may be plastically deformed and flattened parallel to bedding (Wachs and Hein, 1974, fig. 6B, Scholle, 1974, fig. 5). Response of fossil fragments to stress again depends very much on their skeletal microarchitecture, e.g., bryozoan shells (aggregates of microcrystals) are found to be more prone to compaction (mechanical and chemical) than single crystal echinoderm grains (Meyers, 1980). Ooids, on the other hand, are known to fail both by brittle fracture and plastic flowage in ancient oolites. Confirmed reports of modern ooids deforming plastically, however, seem to be unknown (Conley, 1977), nor has experimental compaction of oolitic sediments ever shown failure of ooids by plastic flowage. Pellets usually tend to be squashed under compaction; brittle failure is almost unknown among them. Because of their ductility, the pellets often have elliptical and fusiform cross sections (Hattin, 1975) and deform easily under relatively light load. Intraclasts, because of early lithification, normally resist compaction, and if they at all fail, they do so by fracturing (Sarkar *et al.*, 1980, Fig. 7).

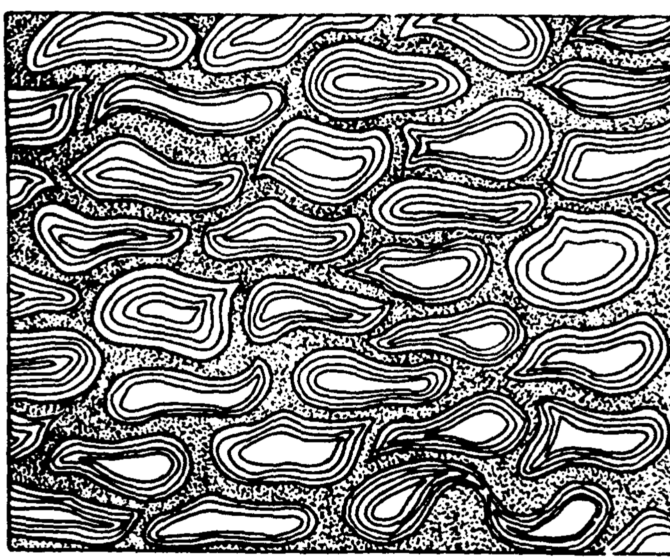
Floating grains themselves, if incompressible and brought into direct contact by the removal of intergranular mud, may ultimately yield by pressure solution along surfaces of contact. On the other hand, ductile grains when closely packed, develop mutually conformable outlines; whether they remain separated depends on the presence of mud or incipient cement (Chanda *et al.*, 1977). Nonductile grains in porous grain-supported carbonate sediments yield either along intergranular microstylolites or fracture (Bathurst, 1975, fig. 319). Other conditions remaining the same, the mode of failure, i.e., rupture versus dissolution, is determined by rate of strain. A high rate tends to favour brittle fracture. It follows that under compaction, ductile grains are initially deformed plastically and on reaching the compressibility limit, yield by pressure solution (Cussey and Friedman, 1977, fig. 4). The reversal of the mode of response from chemical to mechanical failure as in Dunoyane ooids (Radwanski and Birkermajer, 1977), where pitted ooids are distorted, possibly reflects, on the other hand, a change from slow to high rate of stress applications.

As indicated earlier, intergranular pressure solution contacts differ from stylolites basically with respect to the time of origin. While grain-to-grain sutured contacts are a prelithification phenomenon, stylolites transect grains as well as their intergranular micrite or cement (Fig. 3A) and hence clearly develop in a postlithification stage. Solution seams have been recently claimed by Garrison and Kennedy (1977) to be the soft sediment analog of stylolites (see, however, Wanless, 1979). These seams lie predominantly parallel to bedding but lack the fretted form typical of stylolites. These streamers of clay are known to be affected by penecontemporaneous faults (Hancock, 1976) but may cut omission or erosional surfaces. Burrows may be cut by these solution seams, but in many cases solution seams may wrap around burrows. These streamers of clay thus appear to represent products of post-depositional solution and seem to have formed when the sediment was in a semiconsolidated stage.

Inherited factors, namely grain size, shape, sorting, packing and mineralogy, are known to strongly affect the compaction of carbonate sediments (Coogan and Manus, 1975) and consequently compaction-deformation fabrics are not as homogeneous as the fabrics of tectonically deformed rocks. Co-occurrence of highly deformed and undeformed grains within the range of a thin section is commonplace in limestone that has undergone soft-sediment compaction-deformation. Even in plastically deformed ooids in oolitic sediments, where the interplay of inherent fac-



A



B

Figure 3. Schematic drawings of compaction fabrics (not to scale)

A. Oosparite showing postlithification stylolite cutting across ooid as well as cement.

B. Compaction-deformation fabric (normal to bedding) of oomicrite. Note elongation of the plastically deformed ooids, variation in their shapes, and in places mutually conformable boundaries of ooids.

tors is not that pronounced, ooid deformation is found to be uneven with respect to both shape and orientation (Sarkar, 1979). Though there is a strong tendency for the preferred dimensional orientation of deformed ooids parallel to bedding (Fig. 3B), directional variability is not uncommon. This difference of soft-sediment deformation fabrics from tectonically deformed ones presumably arises from the fact that (1) overburden stresses, unlike tectonic stresses, act vertically, and (2) are transmitted heterogeneously from grain to grain in an unconsolidated sediment resulting in differential deformation of grains in contrast to broadly homogeneous distribution of stress in lithified rock (wherein ductility contrast between grain and matrix is relatively subdued). This is because lithification equalizes stresses throughout the rock and, as such, reduces concentration of stress at grain boundaries (Sibley and Blatt, 1975). Furthermore, compaction-deformation of grains is entirely internal and is in no way related to external geometry of the host rock. In tectonically deformed rocks, on the other hand, external as well as internal structures (as of constituent grains) are deformed and the deformation fabrics represent the strain of the host rock (Cloos, 1947). As a rule, overburden stress is vertical and therefore hardly affects the fabric elements that lay parallel or at low angles to bedding. Tectonic stresses, on the other hand, are directed at varying angles parallel to bedding and, as such, strongly reconstitute the depositional fabric.

SUMMARY AND CONCLUSION

The idea of noncompaction in limestone, as this critique shows, was founded primarily on a negative criterion—the lack of detectable evidence of prelithification compaction manifested by grain deformation or grain-to-grain dissolution visible at the level of the light microscope. Further, the point that had been missed on this issue is the fact that a grain may be pressed, yet it may not yield till its load-bearing capacity is exceeded. Again in a mud-supported carbonate sediment, strain may be relieved entirely through dissolution of mud without any visible effect on the grains.

In any event, it is also true that compaction-deformation of grains in limestone is an exception rather than the rule in the normal course of lithification of carbonate sediments. There is no characteristic difference between shallow and deep-water limestones in so far as the nature of compaction-deformation of grains are concerned (Scholle, 1971). The few rare instances of compaction-deformation of grains do not seem to reveal any preferential distribution with respect to the depth of deposition of the host rocks. The cause of unusual compaction-deformation of grains is not environment of deposition but is linked with the presence of preburial rigid body or surface, coupled with an appreciable rate of sedimentation. There are ample grounds for suspecting that the rare examples of unusually compacted limestones that fall off the common trend of lithification must be related to some special geologic situation, i.e., presence of precompaction rigid bodies or surfaces, as explained above. Such unusually compacted limestones warrant re-examination in this context and it is hoped that these exceptions may provide insight into the lithification of carbonate sediments.

The role of pressure solution is being increasingly recognized as of considerable significance in the lithification of limestones (Schlanger and Douglas, 1974; Matter, 1974; Bathurst, 1975; Hudson, 1975; Scholle, 1977). Many limestones that show evidence of extensive early pressure solution curiously lack compaction-deformation of grains. This would imply compaction and cementation may not be two independent phenomena but, to paraphrase Bathurst (1975), may exemplify a weaving of two melodies, on the one hand, generation of cement from within by pressure solution and, on the other, cementation by reprecipitation in pores and/or as epitaxial growth on grains. The process apparently resolves the problem of mass balance in the transformation of carbonate sediment to limestone (Friedman, 1975). Initial porosity may be obliterated to a large extent either by pressure solution in compacted limestone or by precipitation of cement generated elsewhere by pressure solution. Calcium bicarbonate in solution so produced may migrate over distances along a declining pressure gradient before precipitation or may provide cement locally within a matter of millimetres from where it originated (Friedman, 1975; Mimran, 1977). It is not uncommon to find gradational lateral as well as vertical changes from porous to dense limestone in normal geologic settings, the latter generally lacks early diagenetic fabrics and show a higher degree of pressure solution and overpacked grains (Purser, 1978).

Following Scholle (1977), it may be suggested that limestones undergoing burial diagenesis may serve as donors of cement to the receptor limestone which had not been buried to any great depth.

Nelson (1978) has drawn an analogous model of lithification for the temperate shelf carbonate sediments in the Cenozoic of New Zealand. Lithification, according to him, took place, however, during shallow burial and involved intergranular solution of calcitic skeletal particles, especially at those levels in the sediments, enriched in terrigenous material. The lithification process has generated a kind of rhythmic alternation of less well-cemented, microstylolitized, impure limestone beds ('cement-donor' beds) and well cemented, more open textured, purer limestone beds ('cement-receptor' beds). Interestingly stabilization of metastable carbonates is supposed to have been aided by the absorption of magnesium in pore waters by montmorillonitic clays and by the complete oxidation of all organic matter in the bottom sediments (Nelson, 1978). Magnesium purging may also take place through formation of dolomite (Folk and Land, 1975).

Subaerial diagenesis owed its prominence in literature to its impressive documentation from studies of nearshore Pleistocene and Recent limestones, as well as to the fact that most limestones so far studied are of shallow-water origin and hence are prone to exposure. Burial diagenesis, on the other hand, operates spontaneously in the deep crustal environment, and is not as rapid as subaerial diagenesis. Admittedly, limestones may be exposed, but as the geologic record demonstrates, they are not exposed in the way or as frequently as we desire. Even if we assume this to have occurred, emergence has to be always so timed that cementation does take place invariably before sufficient build-up of overburden stresses to affect any detectable deformation of grains and only rarely long after. Even for shallow water limestones, burial diagenesis seems quite probable, for deposition of carbonates, though sporadic through geologic time, was generally rapid (Wilson, 1975). This happens to be precisely the condition which favours generation of a pile of sediment interrupted with surfaces of nondeposition. In any case, however, scope of preburial exposure for majority of limestones deposited over the past 100 m.y. seems rather limited, for in contrast to the Palaeozoic limestones, most of them are of deep-water origin (Hay *et al.*, 1976).

Grainstones can be conveniently grouped into three categories on the basis of nature of response to overburden compaction. The most common type lacks any detectable sign of compaction and of the other types, one may show evidence of pre-cement compaction and the other is usually characterized by two generations fo cement separated by a period of compaction and fracture. Precipitation of first-generation of cement in this type of grainstones precedes compaction but follow dissolution of aragonite. The second-generation cement follows compaction and is believed to have been generated from pressure solution in adjacent limestone. That the late cement is generated in a deep burial environment is suggested by its ferroan composition which would apparently require a reducing environment below the water table (Evamy, 1969). Secondly, the late calcite cements have δc^{13} value near those of marine carbonates and therefore does not seem to have been influenced by soil gases (Allen and Matthews, 1977); rather cementation involved simply solution-precipitation of primary carbonate (Hudson, 1975, also see Lloyd, 1977). The equant habit of late calcite cement crystals, according to Folk (1974), suggests that these crystals could be precipitated from magnesium-poor subsurface brine. Magnesium is apparently lost through reaction to form dolomite and is snatched out by clays to form chlorite or montmorillonite (Folk and Land, 1975, p. 66).

However, the characteristic absence of compaction features in most grainstones with more than one generation of cement suggests that early incipient cement crust normally converts the grain-supported sediments into a rigid, load-resistant scaffolding of the pores into which late cement is emplaced to reduce the porosity to the value commonly found in limestones. In some exceptional situations these rigid frameworks may even be disrupted; this of course appears to require extreme accentuation of overburden stress.

Finally, there seems to exist no unique course of lithification of carbonate sediments. They may be lithified in more than one way and in more than one phase. The relative importance of these ways and phases varies from limestone to limestone and each should be assessed on its own merits.

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