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Oolitic/pisolitic dolostones from the Late Precambrian of south Spitsbergen: their sedimentary environment and diagenesis

ABSTRACT: The oolitic/pisolitic dolostones of the Late Precambrian Dunøyane Member (Höferpynten Dolomite Formation) in south Spitsbergen reveal a set of features attributable to a physico-chemical accretion of the ooids/pisoids, the enormous size of which (attaining 9.7 mm in diameter for simple ooids, and up to 15 mm for complex ooids) is assumed to have resulted from a profuse carbonate precipitation, possibly of primary dolomite. Under such conditions, the accretionary bodies were prevented more effectively from abrasion, and were kept in suspension longer than in present-day calcium-carbonate oolitic environments. The environmental model of the Dunøyane oolites is reconstructed as comprising a pattern of facies from the shallow subtidal through the supratidal zones, the latter being evidenced by the formation of half-moon ooids that owed their structure to a dissolution of some envelopes composed of soluble salts, presumably calcium sulphates (anhydrite or gypsum). The final sedimentation occurred in a subtidal zone, to which all the recognized inter- or supratidal deposits underwent redeposition. This facies pattern is compared to the present-day Bahamian oolite facies, and a conclusion as to the composition of its Precambrian equivalents is presented. The study of diagenetic deformations in the Dunøyane ooids shows them as being mostly confined to pressure solution (pitted, cracked, snouted, and distorted ooids). The commonly known distorted ooids appear to result from subsequent deformations of the pressure-solution contacts in ooids if the latter were pinch-and-swelled and contorted under conditions typical of sedimentary boudinage.

INTRODUCTION

The peculiar oolitic/pisolitic dolostones from the Late Precambrian Höferpynten Dolomite Formation (Dunøyane Member) of southern Spitsbergen were recognized by K. Birkenmajer in 1957 and 1958, during the

Polish Spitsbergen Expeditions; supplementary samples were taken in 1966 and 1970 during the expeditions sponsored by Norsk Polarinstitutt, Oslo (Birkenmajer, 1972). The samples were collected sporadically throughout the whole Dunøyane Member at the three main sites (Fig. 1), namely Wurmbrandegga, Fannytoppen and Dunøyane; they illustrate the general appearance and variation of the lithologies but they cannot however be used for the recognition of vertical succession of the facies.

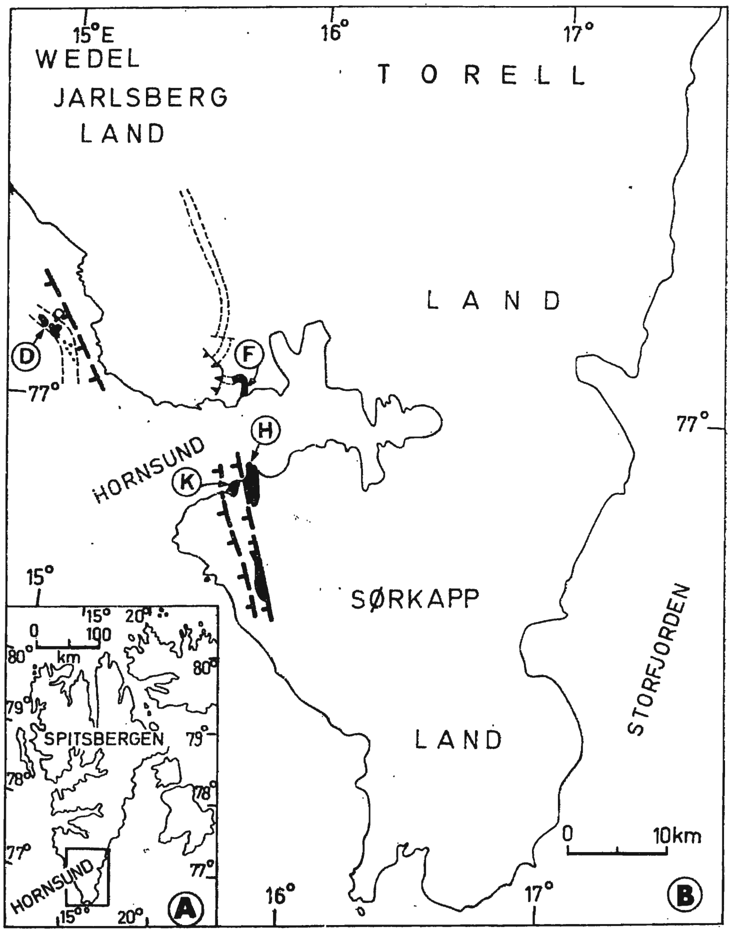

Fig. 1. Location of the Höferpynten Dolomite Formation in Svalbard (*A*) and in south Spitsbergen (*B*)
D — Dunoyane, *F* — Fannytoppen, *H* — Höferpyten-Wurmbrandegga, *K* — Kviveodden

The use of the term peculiar for the oolitic/pisolitic dolostones of the Dunøyane Member is justified both in respect to the size of their particular accretionary bodies (having no relation to the biosedimentary onkolites), and to the presence of many structural varieties, hitherto unknown in such abundance in one oolitic lithology from one lithotope.

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GEOLOGICAL SETTING

The Höferpynten Dolomite Formation is the middle, carbonate unit of the Late Precambrian Sofiebogen Group (Hecla Hoek Succession). Owing to the character of its sediments, the Formation has almost been untouched by metamorphic changes, unlike the underlying Slyngfjellet Conglomerate Formation and the overlying Gåshamna Phyllite Formation which suffered strong recrystallization and internal deformation caused by orogenic stress.

Major unconformities separate the Sofiebogen Group from both the underlying Deilegga Group (Precambrian) and the overlying Sofiekammen Group (Cambrian); another, but weaker unconformity is also present at the base of the Höferpynten Dolomite Formation (Birkenmajer, 1972, 1975). The rocks are strongly folded, with tectonically reversed succession at Höferpynten-Wurmbrandegga and Fannytoppen, but not at Dunøyane (Fig. 2).

The thickness of the Höferpynten Dolomite Formation is highly variable, increasing from the east to west, from 118 m at Fannytoppen to 1150 m at Dunøyane. Four members are distinguishable within the Formation: the lowest Fannytoppen, the Andvika, the Wurmbrandegga, and the highest Dunøyane Member (Fig. 3).

In the Fannytoppen Member, either red, grey, or yellowish limestones predominate over dolostones and slates, the rocks being thickest at Fannytoppen (80 m), thinning toward the south at Höferpynten-Wurmbrandegga (20 m) and wedging out toward the west (Dunøyane). Grey, black and yellowish laminated dolostones of the succeeding Andvika Member often contain black chert nodules. This is a local development known only from Höferpynten (300 m) and Kviveodden (200 m). Grey and yellowish massive or feebly bedded dolostones devoid of cherts have been distinguished as the Wurmbrandegga Member which is thickest on the west (Dunøyane:

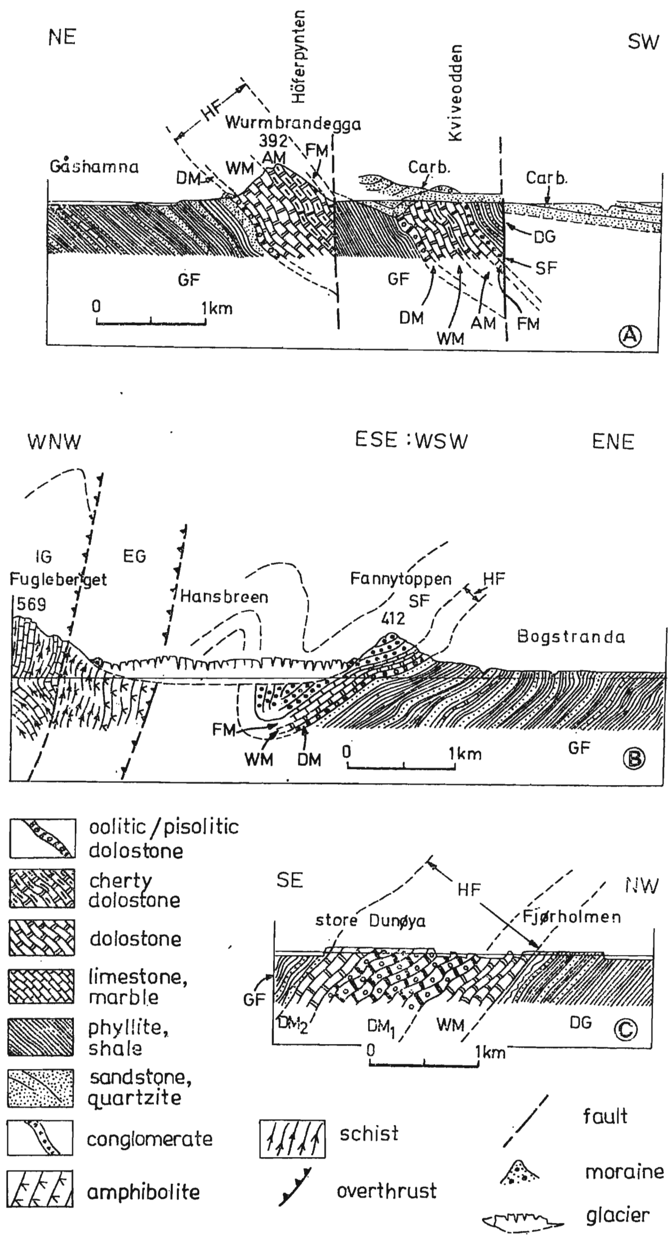


Fig. 2. Geological cross-sections through the Höferpynten Dolomite Formation and associated deposits at Hornsund

Precambrian: *IG* — Isbjornhamna Group, *EG* — Eimfjellet Group, *DG* — Deilegga Group, *SF* — Slyngfjellet Conglomerate Formation, *HF* — Höferpynten Dolomite Formation (*FM* — Fannytoppen Member; *AM* — Andvika Member; *WM* — Wurmbrandegga Member; *DM*, DM_1, DM_2 — Dunoyane Member), *GF* — Gashamna Phyllite Formation; *Carb.* — Lower Carboniferous

400 m) and south-east (Höferpynten-Wurmbrandegga: 350 m) but reduces to a fraction of these values at Fannytoppen (14 m). Intercalations of sedimentary breccias and stromatolitic structures occur in the upper part of the Member at Höferpynten-Wurmbrandegga.

The Dunøyane Member consists of grey dolostone with frequent oolitic/pisolitic intercalations and, near the top, of stromatolitic structures. The Members is thickest on the west (Dunøyane: 750 m), thinning out toward the south-east (Höferpynten: 40 m) and east (Fannytoppen: 24 m). The dolostone is poorly bedded (layers from 0.1 to 3 m thick) but usually well stratified, as is apparent in the ooid/pisoid distribution. Horizontal stratification as well as small-scale trough-like cross stratification is best developed in fine-grained oolites with ooid diameter less than 0.2 mm; such

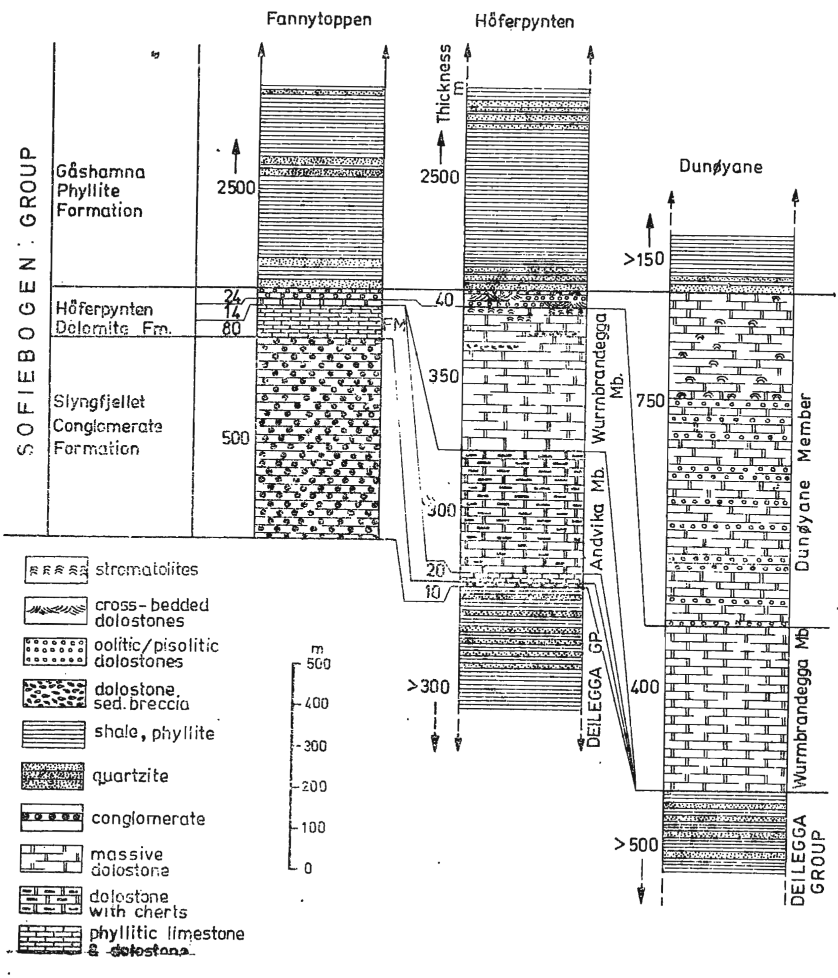


Fig. 3. Lithostratigraphic columns of the Höferpynten Dolomite Formation at Hornsund; *FM* — Fannytoppen Member

being rarely encountered in coarser-grained varieties, including the pisolites. Some dolostone layers show also slump lamination, others graded bedding, both expressed in the ooid distribution. Sometimes, clasts of medium-grade oolites occur in fine or coarse-grade oolitic matrix. The coarsest pisolites show pisoid diameter up to 15 mm at Fannytoppen and Höferpynten-Wurmbrandegga where also half-moon ooids have been found. At Dunøyane, oolitic rocks are usually finer-grained and pisolites are infrequent or absent.

GENERAL FEATURES OF OOLITIC/PISOLITIC DOLOSTONES

The general megascopic appearance of the investigated dolostones from the Dunøyane Member is coarse-granular. All the samples are compact, except for a few in which irregular pores or vugs occur, only partially filled with drusy dolomite. The colour is dark grey, almost black, but with a more or less pronounced dark bluish tint. Granular components are usually darker than the matrix, and therefore they are easily distinguishable; their boundaries are distinct and sharp. The distribution of granular components in the samples is uniform. Cross and graded bedding and slumping have been observed in the field at Höferpynten (Birkenmajer, 1972), but not in the investigated hand-specimens.

Under the microscope, most of the samples reveal oolitic or pisolitic texture which is recognizable either in individual ooids or pisoids, or in compound bodies of various nature. In a few samples, a textureless micritic deposit is present, either preserved in situ (*cf.* Fig. 12), or as intraclasts derived from some interbeddings within the oolitic/pisolitic sequence (Fig. 4).

The matrix of all oolitic/pisolitic dolostones is mostly sparry, often drusy, and only in few places are the relics of primary pelletal or micritic structure observable.

Both the oolites/pisolites and their matrix are composed of dolomite which in some places, mostly in druses, is covered by iron oxides. Traces of other carbonate minerals have neither been observed in thin sections nor detected by chemical staining. A local replacement of ooids by quartz has been reported from Höferpynten by Birkenmajer (1972).

In the description below, following previous recommendations (Rodgers, 1954; *cf.* also Newell & al., 1960; Flügel & Kirchmayer, 1962; Radwański, 1968; Teichert, 1970), the distinction is kept in use of the terms *oolites* or *pisolites* for the rocks, and *ooids* or *pisoids* for individual bodies; the adjective being *oolitic* or *pisolitic* respectively. Within an ooid or a pisoid, *a core* is distinguished and its coating composed of *envelopes*.

The size distinction between ooids and pisoids at the recommended boundary of 2.0 mm (Rodgers, 1954; Pettijohn, 1957; Carozzi, 1960; Flügel & Kirchmayer, 1962) is generally accepted. Since both the ooids and pisoids reveal the same pattern of internal features, such a distinction is however used only when necessary. In the major part of the text, in order not to repeat both these names in each phrase, the general terms Dunøyane oolites and Dunøyane ooids are applied.

OOLITIC ROCKS AND THEIR VARIETIES

In the Dunøyane oolites, several types of components that yield oolitic texture are recognizable. The main types are: oolitic intraclasts, oolitic lumps (grapestones), complex ooids and simple ooids. Of these components, the first and the last ones usually occur separately as predominant com-

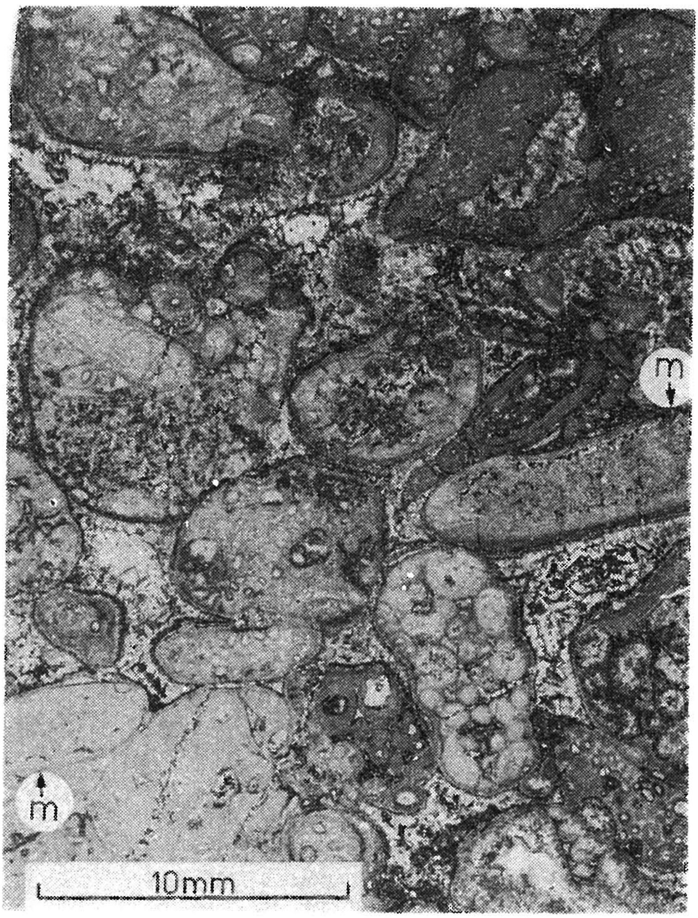


Fig. 4. Deposit composed of intraclasts, most of them being oolitic, a few — textureless, micritic (*m*); one of the latter (right side of the photo) is partially fine-laminated. In the oolitic intraclasts, the complex ooids evidence a two-phase reworking. Some intraclasts with indistinct envelopes are the initial complex ooids. The interspaces are filled with drusy dolomite stained by iron oxides. Locality Fannytoppen

ponents of some rock varieties (Figs 4, 7—8 and 10—11); the remaining ones are usually associated with simple ooids (Figs 5—6).

The oolitic intraclasts, attaining 10—12 mm (maximum 16 mm) in size, are of more or less irregular shape (Figs 4—6), but well abraded and rounded forms also occur (Fig. 5*A*, at right). They are mostly composed of ooids varying in size, and in their texture are of the same kind as that of undisturbed deposits (*e.g.*, such as the matrix presented in Fig. 6*B*). The shape of the abraded ooids indicates their derivation from a fully lithified deposit (*cf*. Fig. 5*A* and 12).

The oolitic lumps, or grapestones, attain 10.3 mm in size, and are usually composed of a group of ooids, all of which protrude more or less distinctly out of the surface. In a few places, the protruding ooids are abraded at their projecting parts (Fig. 5*A*, left side of the grapestone at centre). These forms are interpreted as derived from poorly lithified oolites, the fragments of which have been washed out of the matrix and locally abraded when hard ooids had been projecting from the bunch-of-grape-like clasts.

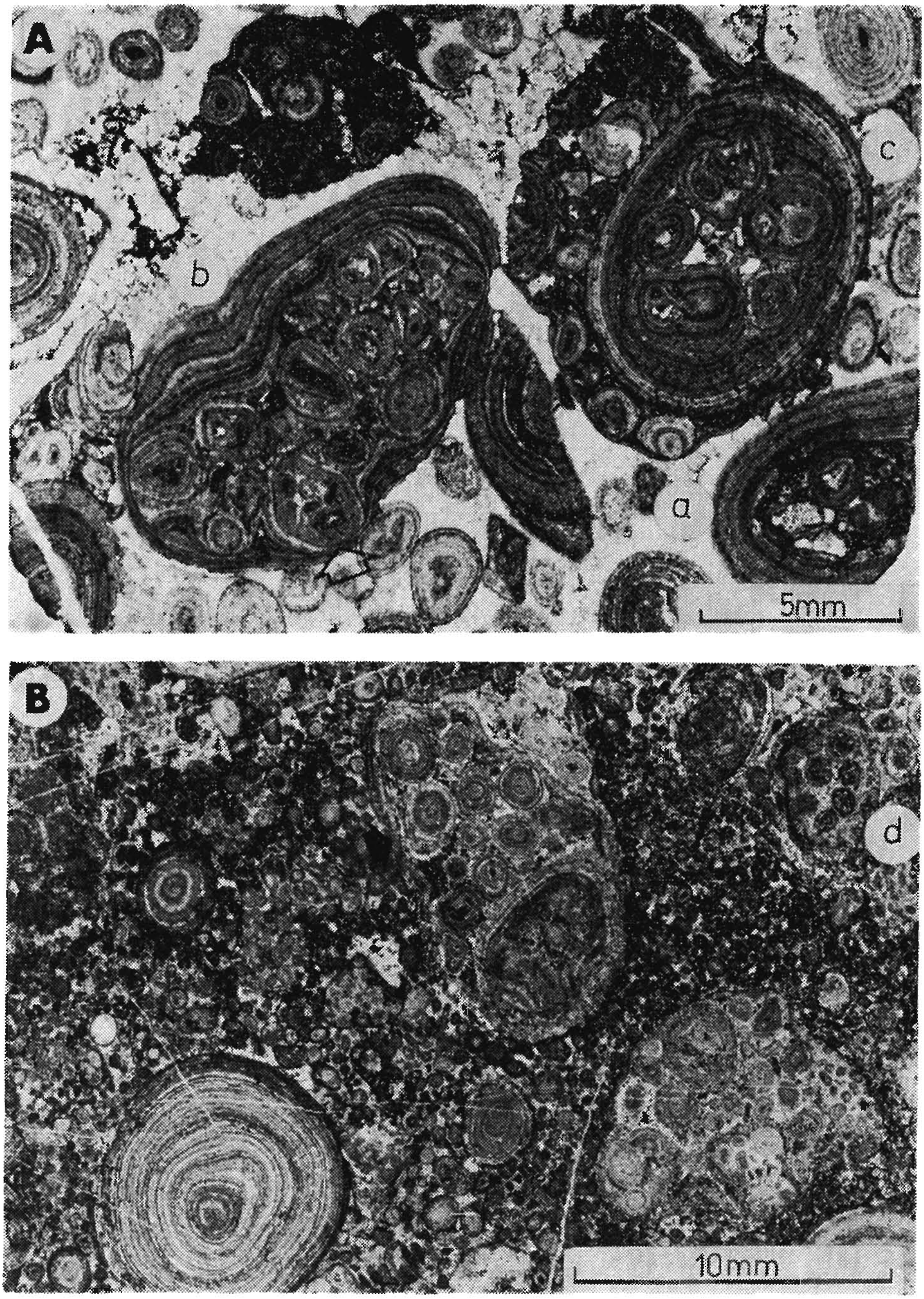
The complex ooids are usually the smaller intraclasts, often grapestones which have been coated with common envelopes. Complex ooids with a few ooids in the core bear a cover of common envelopes that is comparatively thick (Fig. 13); these forms are similar to some of the complex ooids presented by Carozzi (1964, Pl. 3A, B) from a lacustrine limestone of the Upper Triassic (Keuper) of Virginia. The forms that contain larger grapestones in the core are wrapped by a comparatively thinner cover of common envelopes (Figs 5—6); besides analogies to some of the forms presented by Carozzi (1964, Pl. 3H), these are rather comparable to the "ooid pokes", as named in German *(Ooidbeutel)* by Kalkowsky (1908) and Usdowski (1962), which are similar forms, attaining even about 2 cm in diameter that occur in the Lower Buntsandstone of Germany (larger forms, mentioned by Kalkowsky, and not easy to interprete, probably represent local encrustations of the deposit). In the Dunøyane Member, the complex ooids attain 8–10 mm (maximum 15 mm) in diameter, and in a given sample they reach or slightly exceed the size of the largest pisoids (*cf*. Figs 5–6). In some samples, two- or three-phased complex ooids are present, and they may become an important component of the deposit (Fig. 6*A*).

The simple ooids and pisoids which commonly occur as the predominant components of some rock varieties, range in size from the smallest ooids (0.1 mm in thin section) to the largest pisoids with a maximum recorded diameter of 9.7 mm for an elongated, slightly pyroid form (Fig. 5*B*), and 8.3 mm for a near ideally spherical one (Fig. 6*B*). The oolites and pisolites containing them vary in texture, both regarding their sorting and packing. The most common are the pisolites composed of poorly sorted and densely packed pisoids, usually ranging between 4 and 6 mm in diameter (Figs 7–8), and those with pisoids well sorted but loosely or very loosely packed (Figs 10–11). Transitional types with both these features, as well as finer-grained varieties are also present in the samples. In all these types of oolitic deposit, noteworthy is the frequency of ooids either broken or displaying special textural varieties which are to be discussed below.

OOID VARIETIES

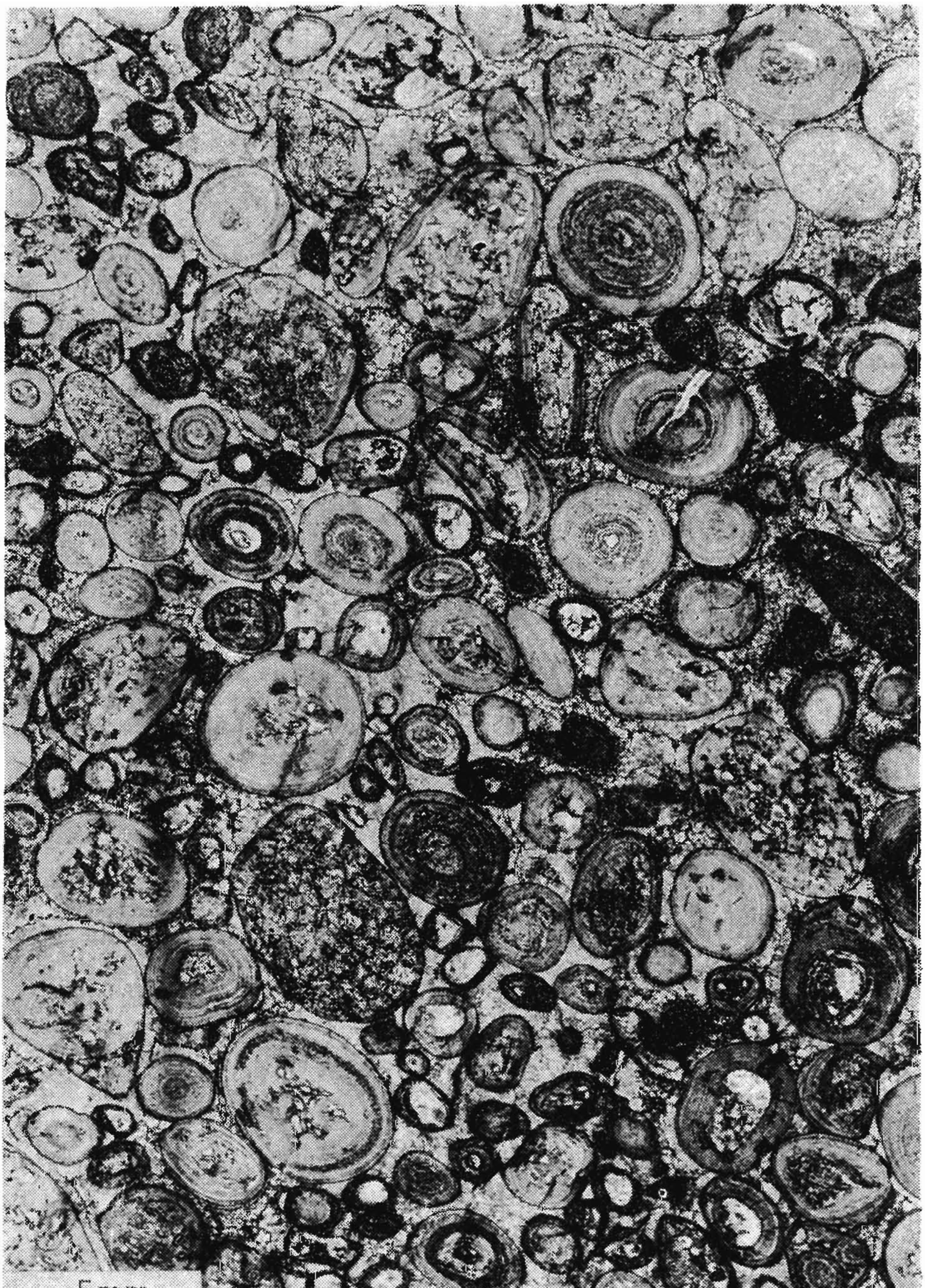
The ooid varieties recognized in the Dunøyane oolites owe their features to the special sedimentary or diagenetic conditions which resulted either in their peculiar shapes that reflected deviations from normal car-

Two examples (*A*, *B*) of loosely packed oolites with various oolitic components: intraclasts (*i*), lumps or grapestones (*g*), complex (*c*) and simple ooids. All the components are of larger dimensions; simple ooids are few, but also of larger diameter (especially in *B*). Note abundant sparry, mostly drusy dolomite cement; in druses, it is covered by iron oxides; in the cement relic patches of pelletal structure (*p*) are visible. Locality Wurmbrandegga

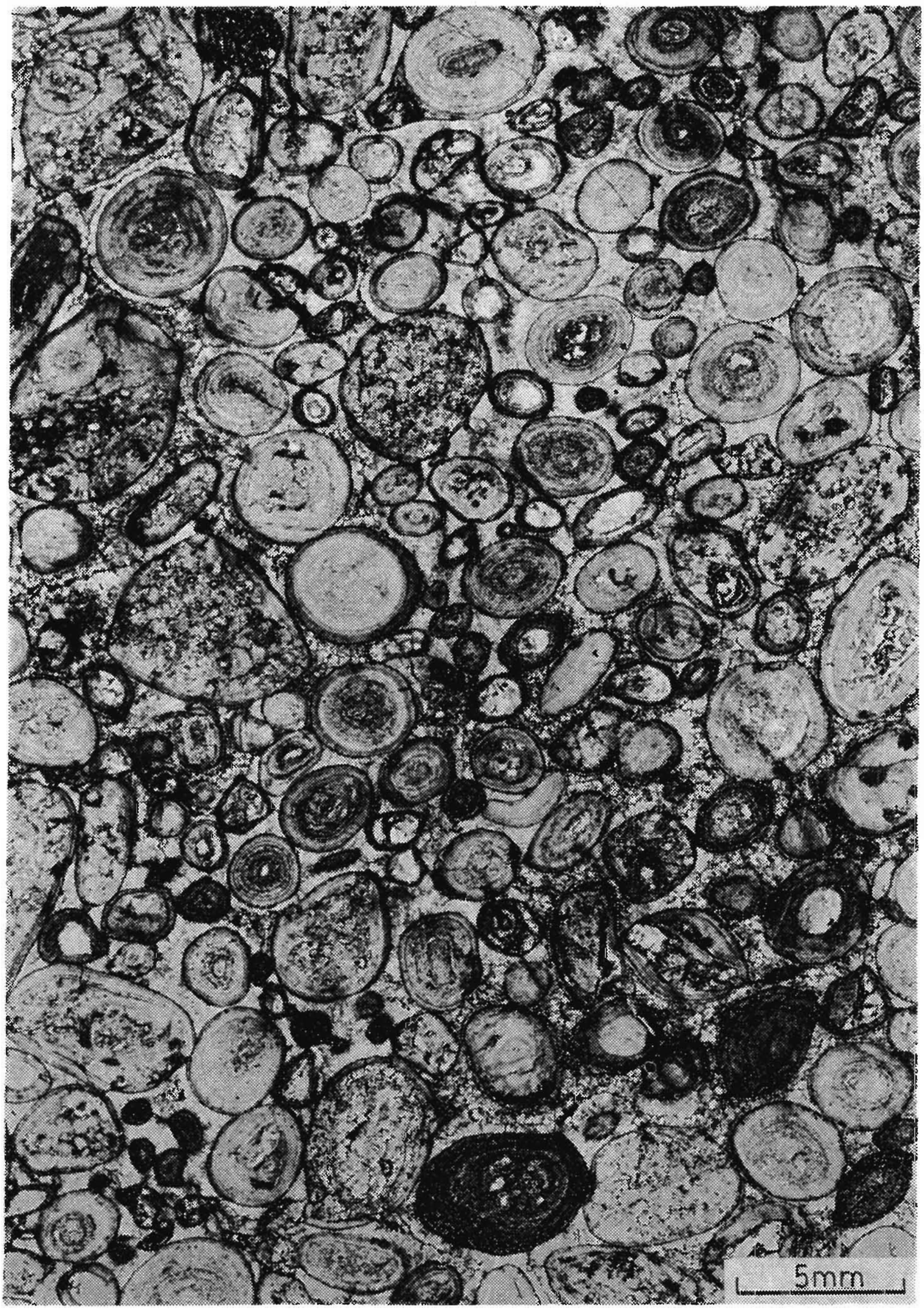


A — Complex and simple ooids in loosely packed oolite; of the complex ooids (*a, b, c*), one is normal (*a*), one two-phased (*b*, the first phase arrowed), and one three-phased (*c*, the first phase arrowed, the third — locally destroyed by a stylolitic seam). Locality Wurmbrandegga

B — Densely packed oolite with various oolitic components: intraclasts (bottom right); grapestones, one of which is covered with fragmentary oolitic envelope (arrowed) featuring it as a form transitional to a complex ooid; another complex ooid (*d*) and a few larger ooids are encountered in the matrix containing small ooids, pellets and poor sparry cement. Locality Fannytoppen



Densely packed and poorly sorted oolite with sparry, partly drusy cement. The ooids are associated with intraclasts of pelletal or oolitic structure. A few ooids are complex; in some ooids an indistinct half-moon effect is visible (oriented vertically in the photo). Details of the ooid structure are presented in Figs 9 and 14—16. Locality Fannytoppen



Another example of densely packed and poorly sorted oolite with sparry cement; details presented in Figs 9 and 14—16. Locality Fannytoppen bonate accretion, or developed during various post-depositional disturbances. All the discussed varieties refer to single ooids, although some are formed by mutual contacts of a pair or a group of several ooids. The same features as observable in simple ooids are also revealed by those making up the core of complex ooids (*cf.* Fig. 13). As separate categories the following varieties are considered:

(1) Abraded and broken ooids (*cf.* Fig. 9).

(2) Half-moon ooids, as called by Carozzi (1963), the internal structure of which (*cf.* Figs 10—13) result from a collapse of some internal envelopes that settled into a heap on the ooid bottom, the void being filled with usually light-tinted secondary dolomite (a "*half-moon aspect*" of Wherry, 1915; or "*bipartite ooids*" of Choquette, 1955). As the collapse was due to gravity, this geopetal bipartition is a remarkable indicator of bottom/top relation in the layers (*cf.* Shrock, 1948, pp. 283—285; Choquette, 1955; Carozzi, 1963; for Spitsbergen oolites — Horsfield, 1973).

(3) Pitted ooids (*cf.* Fig. 14 and 15b).

(4) Cracked ooids (*cf.* Fig. 14c and 15).

(5) Distorted ooids, as called by Carozzi (1961a), and underdeveloped forms termed here snouted ooids (*cf.* Fig. 16). The distorted ooids, which usually occur in groups of two or more being jointed in a shape recalling the typographic symbols *S* or *§* (Carozzi, 1961a; Misik, 1968; Bachmann, 1973), are also known under different names (*spastoliths* — Pettijohn, 1957, *oolites confluantes* — Misik, 1968; *verformte Ooide* — Bachmann, 1973).

To illustrate varieties here recognized such ooids are selected (Figs 9 and 14—16) which bear only one kind of variation, although forms with a few of the discussed features are also common (*e.g.*, Fig. 15*d*).

ABRADED AND BROKEN OOIDS

The presence of abraded and broken ooids is a common feature in densely packed and poorly sorted oolites (*cf.* Figs 7—8). In other rock varieties they are found rather sporadically, e. g. in loosely packed oolites (*cf.* Fig. 10). The shape both of the abraded and of the broken ooids shows that these were supposedly produced by the same agents. Abraded ooids are usually broken with some respect to their concentric texture: the abrasion progresses to various extents in different portions of the ooid body, and longer sectors tracing internal envelopes are visible. It appears as if the ooid had been crushed from various sides until a more resistant envelope was reached at least in some portions of the abraded sphere. A more external part was at that time crushed mostly along radial sectors resulting in step-like boundaries of the abraded ooids (Fig. 9*d*). The same effect is also present in broken ooids (Fig. 9*c*) and in the split-off parts of more damaged ooids (Fig. 9*a*), and it may therefore be regarded as evidencing a more or less advanced radial orientation of carbonate minerals, which is however invisible in undisturbed ooids.

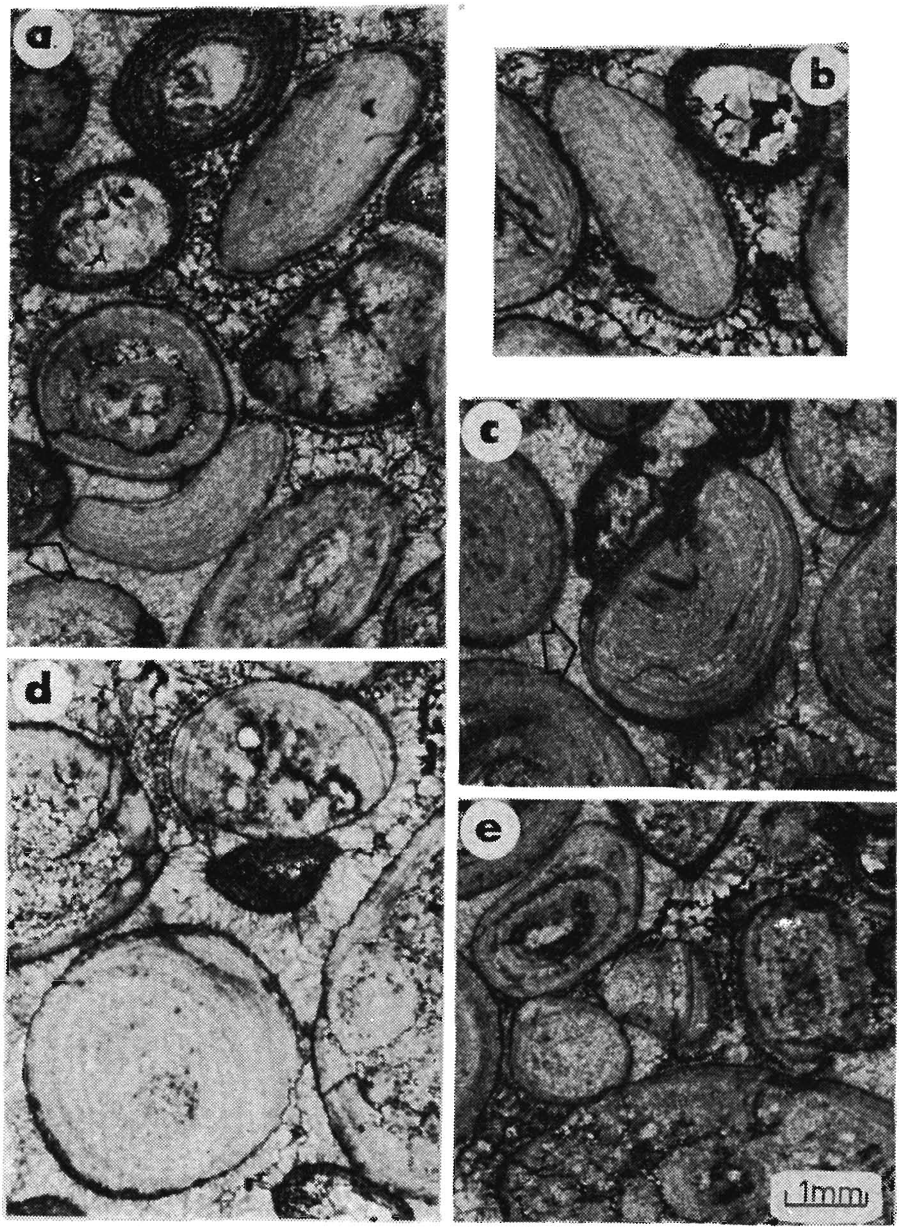
The broken ooids are represented either by half-split bodies (Fig. 9*c*) or split-off portions of the envelopes, presumably the most external ones

(Fig. 9*a-b*). Of the latter, some pieces look freshly broken (Fig. 9*a*, near left bottom, arrowed), and some are subsequently abraded (Fig. 9*a-b*). Additionally, irregular chips of broken ooids are also encountered (Fig. 9*e*).

All the presented features show that the ooids were rigid bodies but more or less brittle and fragile when the fracturing was operating in the sedimentary basin. It took place shortly before final deposition as there are only a few examples of regeneration in broken ooids (Fig. 9*c*).

As appears from the referenced literature, broken and regenerated ooids are commonly met with in hypersaline lacustrine or lagoonal environments, which may be exemplified by the shorezone of the Great Salt Lake (Carozzi, 1961b, 1962; *cf.* also D'Argenio & al., 1975, Fig. 3) and by lagoonal Keuper conditions with sulphate precipitation in the Upper Triassic of the Holy Cross Mts, Central Poland (Łabęcki & Radwański, 1967). In both these environments, the more or less pronouncedly developed so-called cerebroid ooids (Graf & Lamar, 1950; Carozzi, 1962) are frequently present, the origin of which was explained by Carozzi (1962) as resulting from non--uniform growth and convertion of aragonite into calcite which preferentially took place in some sectors of successive envelopes; the mineralogical nature of this process has however been recently objected to by Kahle (1974) who finds it as being recrystallization of aragonite, and by Sandberg (1975) who recognizes it as originally differentiated accretion of aragonite. It was also suggested (Łabęcki & Radwański, 1967), that the cerebroid structure is responsible for a smaller resistance of the ooids against the mechanical reworking and breaking. Although the cerebroid structure is not well shown in all examples of lagoonal ooids, it seems that this type of ooids is attributable to non-marine environments. The occurrences of broken ooids, usually associated with regenerated and cerebroid ones, support this assumption, as demonstrated by the following examples.

Usdowski (1962, 1963) presented such a set of oolitic structures, following the classical description by Kalkowsky (1908), from the Lower Buntsandstone of Germany, the environmental conditions of which are regarded as very shallow marine, supposedly of the tidal-flat type with temporary hypersaline lagoonal conditions indicated by anhydrite precipitation. Krebs (1968, 1974) reported irregular, superficial and broken ooids from a very local oosparite facies in the Devonian of the Rhenish Schiefergebirge which developed in the intertidal zone of a backreef lagoon associated with coral "reef" bioherms. Both these authors drew attention to the large size of the ooids: Usdowski (1962) recorded average diameter in some samples as attaining 2.24 and 2.61 mm, and in one sample 3.22 mm, maximum even up to 6.3 mm, whereas Krebs (1968, 1974) records up to 2.5 mm. Furthermore, Simone (1971, 1974) described a facies with similar ooids, including regenerated ones, from the Lower Cretaceous of the Campanian Apennines and other regions of Italy and Yugoslavia (*cf.* also D'Argenio, 1967; D'Argenio & al., 1975), which she interpreted as inter- or supratidal. It is important that all these authors presented very similar conclusions as to the sedimentary environment of the ooid-bearing sequences, and these conclusions were drawn from the analysis of associated deposits and not from the ooids themselves; they also do not overlap each other with their references. Finally, a facies with larger (up to 2.0 mm), very often irregular or superficial, but broken and regenerated ooids was reported by Szulczewski (1971) in a situation similar to that presented by Krebs, viz. in lagoonal deposits that formed at the topmost part of a coral "reef" bioherm when it was rising up near to the water level in Upper Devonian time in the Holy Cross Mts, Central Poland. Szulczewski assumed the existence of a former lagoonal environment, with supposed hypersaline conditions, when analysing the development of the whole sequence; the discussed



Abraded and broken ooids; locality Fannytoppen

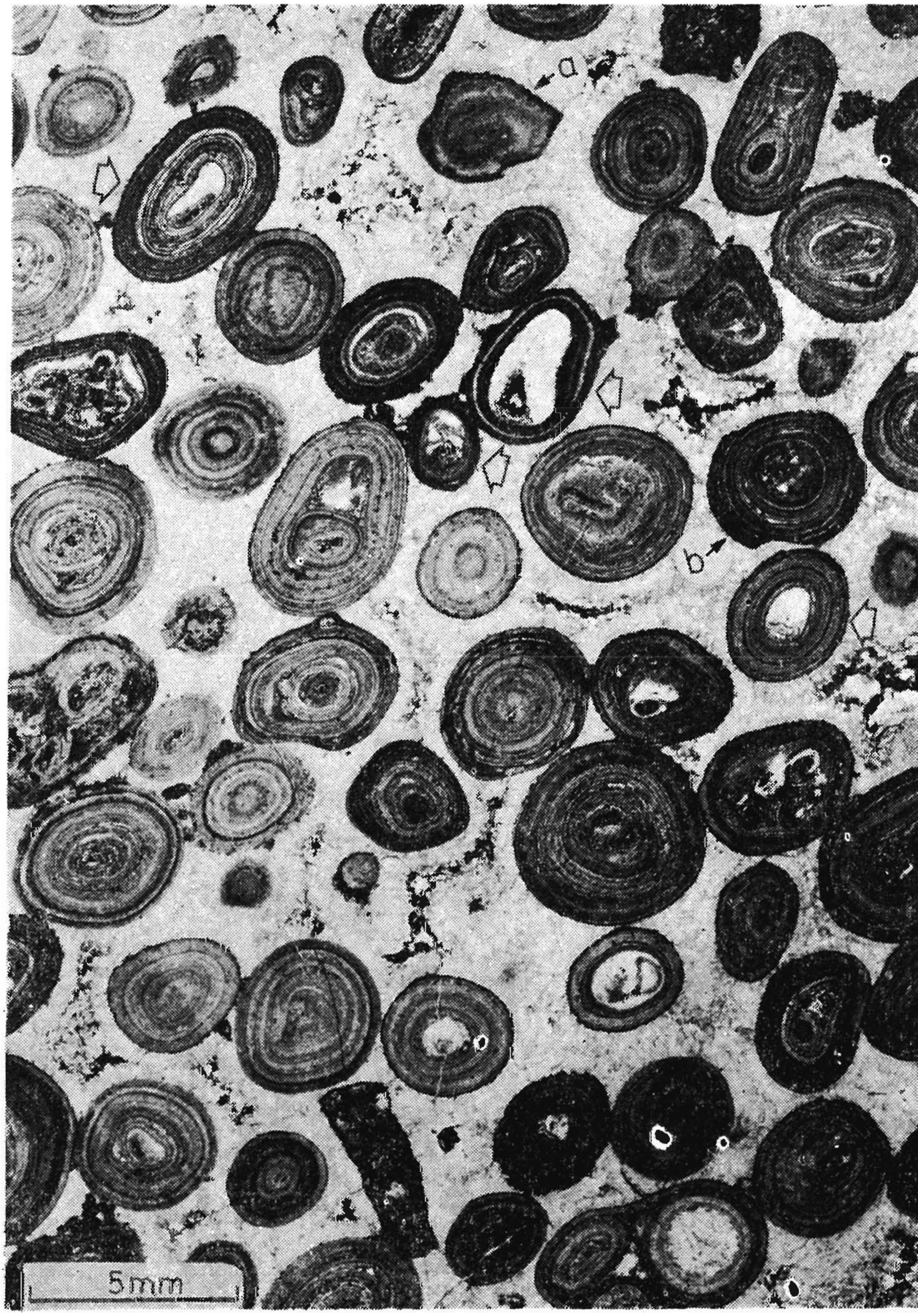
a — Two fragments of larger ooids, peeled-off along the envelope boundaries: one fragment abraded (top right), and the second one (bottom left) fresh with a sharp edge reflecting its primary radial texture (arrowed). Note a crack in the ooid above the latter, outlining a sector of similar shape

b — Abraded fragment of a broken ooid

c — Broken and indistinctly regenerated ooid; regeneration (arrowed) consists of a darker envelope traceable along the whole ooid surface

d — Two abraded ooids; in the upper one a sector of the last envelope is alsc peeled-off

e — Small chip (at centre) formed by radial breaking of an ooid



Loosely packed and well sorted oolite, some ooids of which are complex, and some reveal the half-moon effect (arrowed); abraded ooids (*a*, *b*) are also indicated. Locality Wurmbrandegga oolitic facies itself was, however, interpreted with a reference to the papers by Łabęcki & Radwański, and by Krebs.

As follows from the presented review, it is reasonable to regard the broken and regenerated ooids, and likewise the irregular (with scalloped appearance *sensu* Freeman, 1962) and cerebroid ones (these are not present in the Dunøyane Member), as indicative of the conditions ranging from very shallow marine, if not only intertidal, to supratidal, mostly lagoonal hypersaline. Such an environmental range was inferred by Łabęcki & Radwański (1967) for a few other occurrences of broken and regenerated ooids, of which one presented by Roda (1965) from the Miocene of southern Italy is the most remarkable. It may be also mentioned that a facies with broken and regenerated ooids, first recorded by Carozzi (1961b), from a longer rhythmic sequence of the Mississippian in Central Alberta, Canada, was interpreted as lagoonal by Walpole & Carozzi (1961; their microfacies 5, subtype *5a*). In the sites where the broken and regenerated ooids appear as admixture in other, mostly calcarenic deposits, they are interpreted as having been swept away from the place of growth and redeposited in deeper marine facies (Carozzi & Roche, 1968; Simone, 1971); the same is certainly true for some cerebroid ooids (*cf.* Lacey & Carozzi, 1967).

In the Dunøyane oolites, the broken and regenerated ooids are mixed with wholly preserved ooids, and their presence may be therefore regarded as attributable to a nearby lagoonal environment and a reworking of the ooids therefrom. It is important to note the presence of the forms (*cf.* Fig. 9*a*, *c*-*e*) reflecting the indistinct radial texture of these ooids which remains undetectable in the ooids that did not suffer the abrasion and breakage.

HALF-MOON OOIDS

The half-moon effect in ooids is well developed only in a few rock varieties, while in others it is weak (*e.g.*, Figs 5 and 7) or absent. If present, it is marked to varying extents, as follows:

(*a*) In loosely packed and well sorted oolites (Fig. 10) it appears in some of the ooids, and it is mostly restricted to their internal parts that collapsed, while the thick shell of median and external envelopes remained undisturbed (ooids arrowed in Fig. 10).

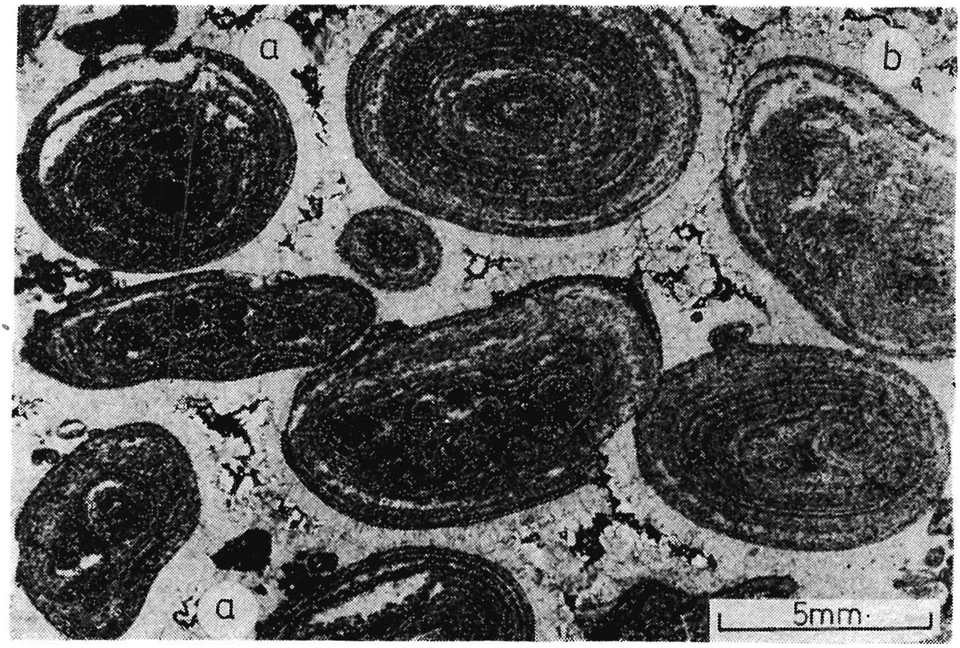
(*b*) In oolites composed of the largest ooids (pisoids — up to 8.5—8.8 mm) that are usually of the same diameter (Fig. 11), the half-moon effect is poorly developed and corresponds to a slightly advanced dissolution in some envelopes. In these ooids the collapse is marked either by irregular crumpling of the envelopes (*b* in Fig. 11), or by their more regular settling accompanied by a rippling of particular envelopes (*a* in Fig. 11). Noticeable in this variety is a strong flattening of ooids that collapsed as a whole in the same direction as their envelopes.

(*c*) In complex ooids (pisoids) deposited in fine oolitic and pelletal background (Fig. 13).

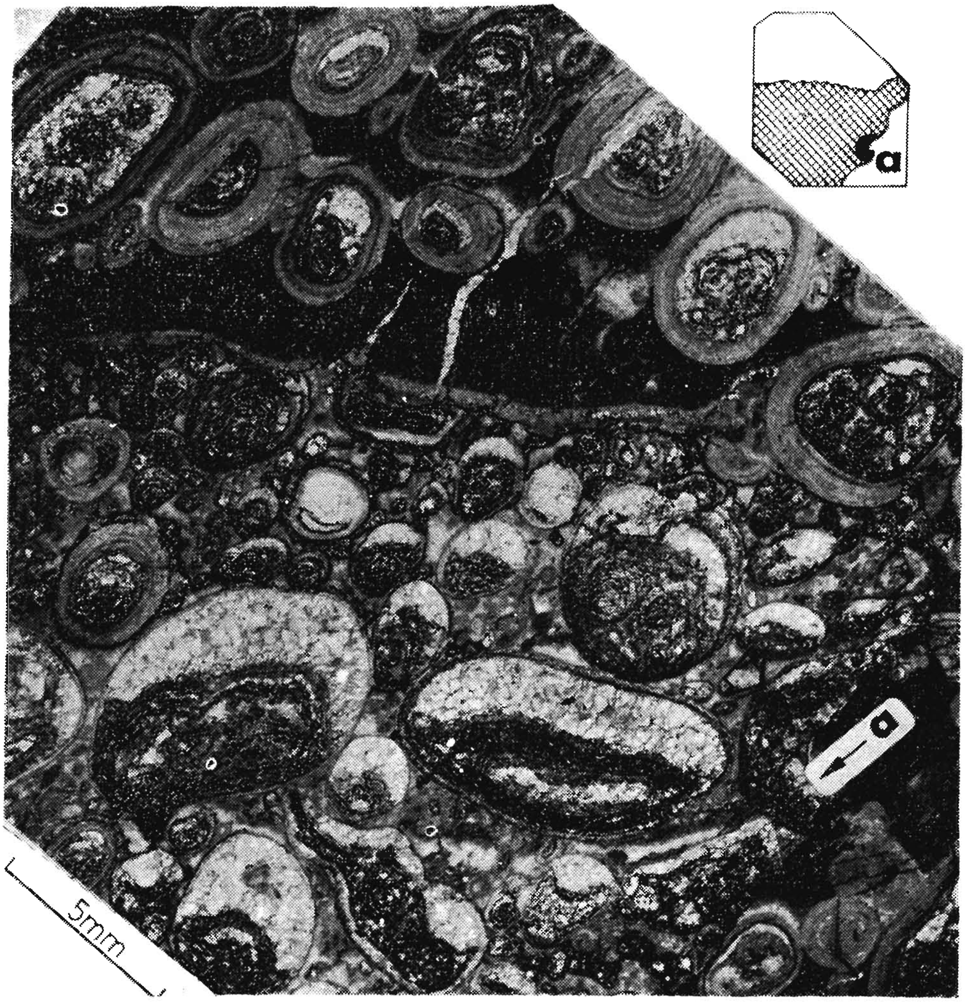
(*d*) In a peculiar and rare rock variety in which all the ooids are of variously developed half-moon type, regardless of their deposition either in a micritic matrix, or their being in large intraclasts (Fig. 12). In these ooids of the matrix, the half-moon effect is mostly marked in internal envelopes, whereas the external ones remain undisturbed (upper part of Fig. 12). In the intraclast (lower part of Fig. 12), the half-moon effect is the most pronounced, and the volume decrease of the collapsing envelopes in ooids is the greatest.

The origin of the half-moon ooids was satisfactorily explained and proved in experiments by Carozzi (1963) who showed that these structures result from the complete dissolution of some envelopes composed of soluble salts, presumably calcium sulphates (anhydrite or gypsum), and that they owe their shape to subsequent collapse of the insoluble parts. The half-moon ooids must therefore have formed in any environment which was at least temporarily hypersaline. In the fossil records, the half-moon ooids were first reported from the United States in various shallow marine or intertidal (containing *e.g.* stromatolites or desiccation breccias) deposits of older Paleozoic age (Upper Cambrian, Lower Ordovician), all of which had however undergone a diagenetic substitution by dolomite and/or silica (Wherry, 1915; Choquette, 1955; Zadnik & Carozzi, 1963; Carozzi, 1963; Carozzi & Davis, 1964). The presence of primary sulphates may therefore only be conjectured and, as the half-moon structure in ooids shows, the sulphates were removed prior to the substitution. In limestones, small half-moon ooids (*c* 0.3—0.4 mm) were presented by Castellarin & Sartori (1973 a, b) from the lowermost Liassic calcareous tidal-flat deposits of the Lombardian Alps, Italy. These deposits bear a set of structures (desiccation and sheet cracks, bubble-like vugs or "birdseyes") that evidence an intertidal or lagoonal environment; in respect to the half-moon ooids, Carozzi's (1963) interpretation is not however referenced, and the ooids are interpreted as subtidal and subsequently leached under inter- or supratidal conditions during temporary emersions. It may be noted that in other intervals of this sequence, also caliche-like deposits with secondary pisoids are present, the same as in the Central Apennines (*cf.* Bernoulli & Wagner, 1971; Castellarin & Sartori, 1973 a); the latter differ however from the half-moon ooids distinctly and they represent different structures, discussion on which will be omitted here.

In the Donøyane oolites there is no indication of any subaerial solution; on the contrary, the structure of the half-moon ooids, the envelopes of which collapsed after deposition in normal oolite or pelletal facies (*cf.* Fig. 13) and kept their continuity when settled, evidences that Carozzi's (1963) interpretation should be supported. The distinct boundaries of each collapsed envelope (Fig. 12) demonstrate these to have been separate when settled, and therefore something other than carbonate had to be removed from between them. It may be suggested that it was calcium sulphate, the same as concluded by Carozzi. A gradual advance of the leaching in some ooids (*cf.* Fig. 11) may however suggest also that some envelopes were not composed of, but only interwoven with calcium sulphates. On the other hand, the ooids which seem to be devoid of collapsed envelo-

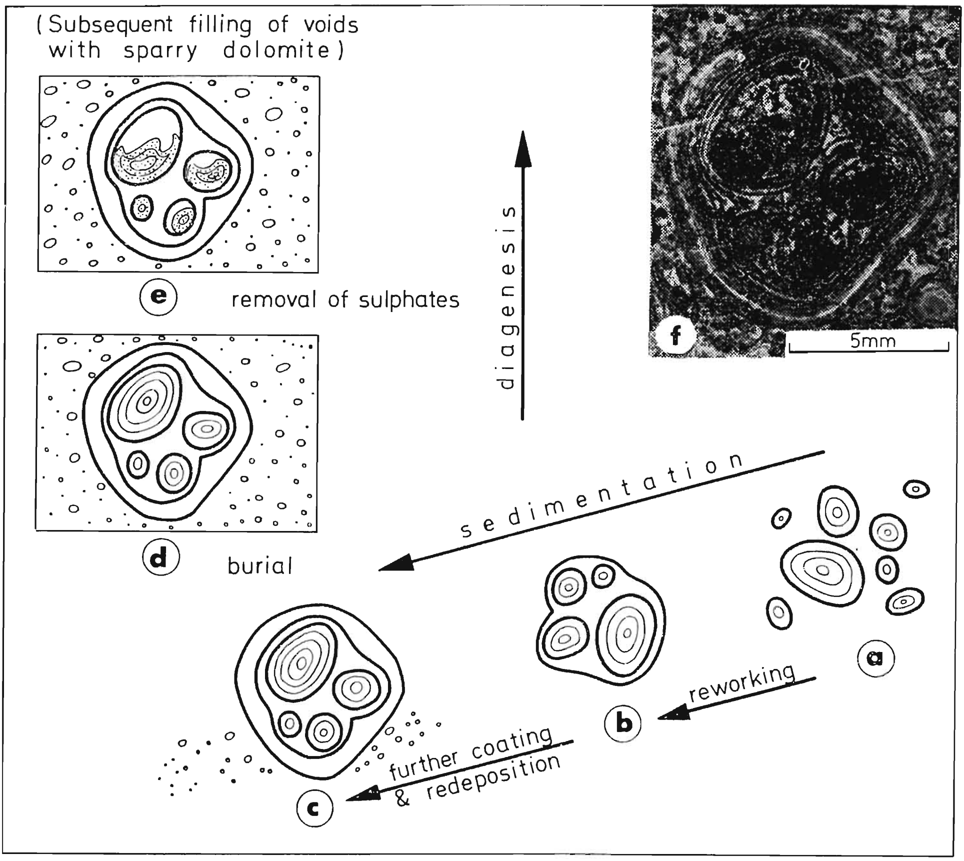


Ooids, some of which show leached-out envelopes; the half-moon effect is therefore progressing (*a*). In other ooids the dissolution of external envelopes is marked to various extent: when more advanced, the crumpling of envelopes is pronounced (*b*). All the ooids are flattened, due to collapse, in the same direction as the internal envelopes settled. Locality Wurmbrandegga



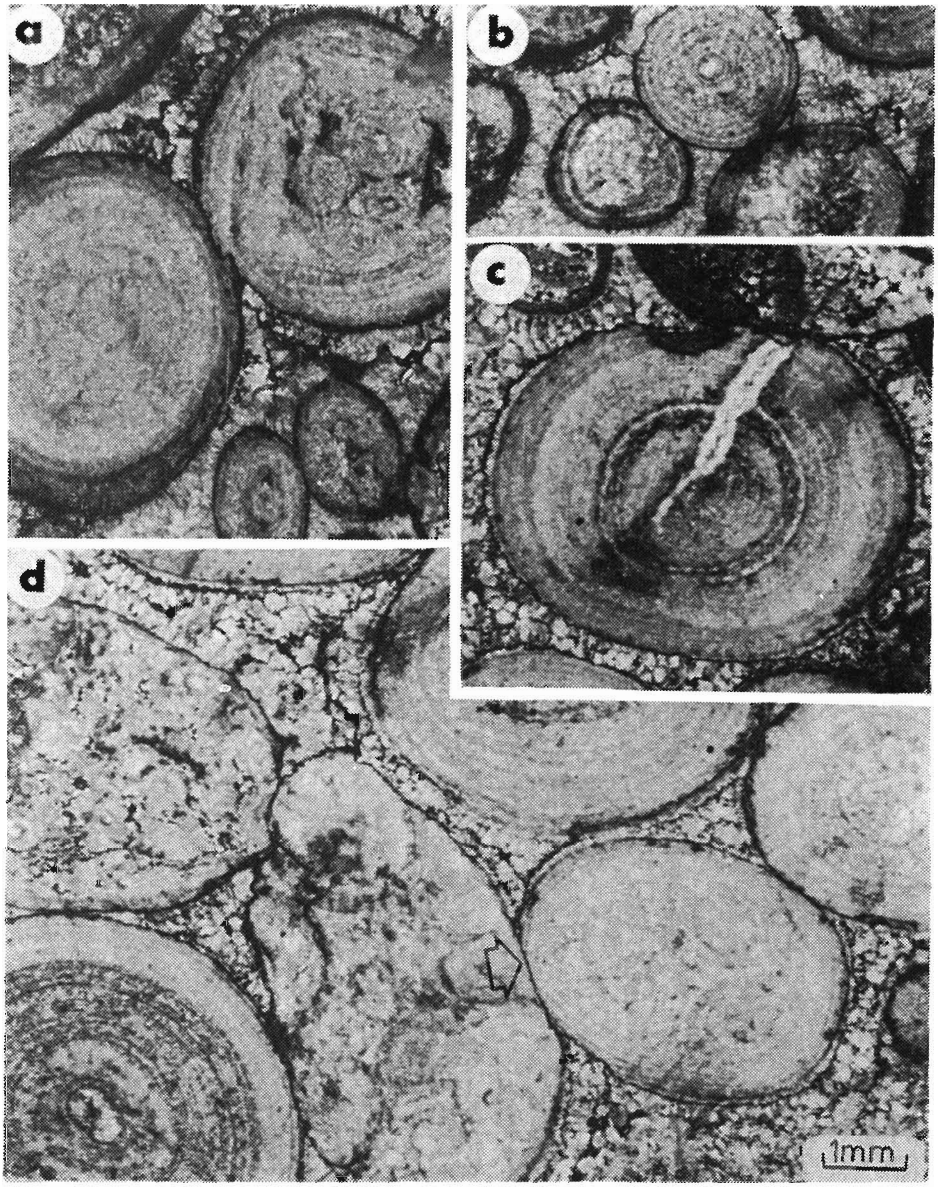
The most developed half-moon effect both in the ooids of a larger intraclast with sparry matrix (lower part), and of the surrounding micritic deposit. All the half--moon ooids are oriented in the same direction (down in the photo) evidencing the collapse to have progressed when the intraclast was brought into the micritic deposit. In this intraclast, some of the half-moon ooids are almost completely dissolved and only a small residuum is left; on the other hand, the greatest ooids are strongly flattened and deformed by compaction

The inset shows the outline of the intraclast and position of the ooid (*a*) which was abraded before a collapse; the latter appeared therefore independently in the both incised parts of this ooid (arrowed). In the background, a few veins filled with dolomite are visible; one of them cuts also an ooid. Locality Wurmbrandegga



Summarized history of syn- and postdepositional events in the sulphate-cored ooids of the Dunøyane Member:

a — formation of sulphate-bearing ooids in the sub-evaporitic facies, *b* — coating of the sulphate-bearing grapestone, *c* — complex ooid deposited in the fine-oolitic and pelletal facies, *d* — burial of the complex ooid, *e* — collapse of the sulphate-bearing envelopes: the half-moon effect originates to such extent as visible in the resulting rock (*f*) from Wurmbrandegga



Pitted ooids; locality Fannytoppen

a — Two ooids penetrating each other by a mutual pit of the dentate type with an accessory serration. Note indistinct smooth pits in other ooids (bottom right)

b — Chain of three pitted ooids, the median one of which was of variable solubility along its surface: as a result, it is pitting one of its neighbours (at left), and is itself pitted by another one (at lower right)

c — Disturbance at the pit contact: an ooid with less soluble surface (at centre) cracked radially when penetrating the more soluble one. Note pure drusy dolomite (white) healing the crack, and differing from the iron-stained dolomite in the matrix

d — Shallow pits between ooids arranged in a chain: most of the pits are smooth, one is dentated (between the oolitic intraclasts at centre). Note an effect of "flattening of the nose" (arrowed) by a smaller ooid (indistinctly complex) when pitted by a larger grain (intraclast) with lower sphericity pes (Fig. 12, at centre), if this is not the sectioning effect, should be regarded as built mostly of sulphates.

Nevertheless, when assuming that the soluble envelopes were built of or interwoven with calcium sulphates, a temporary hypersaline environment with carbonate/sulphate precipitation (*sub-evaporitic* of Carozzi, 1963) must be ascribed to the investigated sequence. Furthermore, formation of mostly or almost completely sulphate ooids, may be similar to some anhydritic ones reported by Voorthuysen (1951) from the Upper Jurassic (Purbeckian) Münder Marls of the Dutch subsurface, cannot be excluded.

In comparison with the half-moon ooids from the Upper Cambrian Allentown Dolomite, New Jersey, described in detail by Carozzi (1963), those from the Dunøyane Member bear analogous features as to the relative time of dissolution of some envelopes and collapse of the others. It may be also noted that the former ones attain large dimensions, 2.5 mm in average, and even 5.0 mm in maximum diameter as recorded by Carozzi (1963).

The presence of complex ooids with similarly oriented half-mooned components in their interiors (Fig. 13*f*) demonstrates that the reworking of sulphate-bearing ooids in the Dunøyane environment took place before their final burial and collapsing of envelopes (*cf.* Fig. 13*a-e*). A similar reworking is certainly the case reported by Payton (1966, Fig. 6), and Knewtson & Hubert (1969, Fig. 14), who pictured simple half-moon ooids with their cores displaced laterally or upward; the redeposition of these ooids had however to happen after their sulphate-bearing envelopes had settled.

PITTED OOIDS

The pitted ooids occur in densely packed and poorly sorted oolites (*cf.* Figs 7—8), and they are developed either as smooth (Fig. 14*b-d*) or dentate pits (Fig. 14*a* and 15*b*). All these pits are rather shallow and the ooids penetrate only superficial parts of their neighbours.

The pitted contacts in ooids have been recognized to be the result of pressure-solution phenomena (Graf & Lamar, 1950; Carozzi, 1960; Radwański, 1965). Their origin is of the same kind as that of pitted pebbles (*cf.* Macar, 1937; Kuenen, 1942; Radwański, 1965), quartz grains, especially in some types of quartzites (*cf.* Heald, 1965; Weyl, 1959; Carozzi, 1960; Radwański, 1965; Skolnick, 1965), as well as in other detrital or biogenic clasts (*cf.* Weyl, 1959; Radwański, 1965; Trurnit, 1968). Pitting processes develop if these deposits are loose, and compaction pressure may therefore be transmitted through the selected contact points of detrital particles. At these points as a result the total pressure is locally increased. If the intergranular solution is saturated with the substance of which these particles are composed, dissolution has to take place at points of increased pressure in the same way as postulated by Riecke's principle. When the detrital particles are embedded in a matrix, the pressure is transmitted through all the dispersed contacts and it is not sufficient to initiate the pressure-solution processes at any contact.

In the pressure-solution process, if two particles of different solubility are in contact, the more soluble one is dissolved first and a pit is formed by the less soluble one which starts to penetrate the more soluble neighbour. In ooids, which are bodies of more or less the same solubility, the differences in solubility of the external surfaces control the development of pits (Radwański, 1965), the same as in quartz grains in quartzites (*cf.* Carozzi, 1960; Radwański, 1965). When the ooid surface at the contact point is of lower solubility throughout the pitting processes, a simple, smooth pit develops. In a given ooid it may happen that its surface is less soluble in some parts than in others when compared with the solubility of the contacting ooids; as a result such an ooid may pit one of its neighbours, and simultaneously be pitted itself by another one (Fig. 14*b*). In other cases, where the ooid surface is of variable solubility at the contact point, or along the progressing pit surface, a mutual pit develops, either simply dentated (Fig. 15*b*), or with accessory serration (Fig. 14*a*). The latter type is very close in its shape, and just the same in mode of formation, as that of stylolites (*cf.* Radwański, 1965; Trurnit, 1968).

In the ooids, the formation of pits and their shape depends only on differences in solubility of the contact points under pressure. The pit shapes are thereby independent both of the relative dimensions of the contacting ooids, and of their sphericity. Detailed studies of pitted pebbles in conglomerates and pitted ooids in calcareous oolites (Kuenen, 1942; Radwański, 1965) illustrated this rule distinctly, and this is contrary to some familiar opinions on the control of sphericity during formation of pits. It may be noted that in loose pebbles, pits of smaller pebbles into larger ones were mostly observed as being more spectacular, but careful inspection of polished slabs of conglomerates shows that the reverse situation occurs to a similar extent (Kuenen, 1942), and features smaller pebbles with a pit having a nearly flat surface. This effect of a "flattening of the nose", as called by Kuenen, by a smaller pebble when it is pitted by a larger one (Kuenen, 1942, Pl. 1, Figs 2—3; *cf.* also Trurnit, 1968, *13* in Fig. 2) is recognizable also in the Dunøyane ooids (Fig. 14*d*).

In the studied Dunøyane deposits, complex ooids and oolitic intraclasts (Fig. 14*d*) are pitted to the same extent as simple ooids, which indicates that the pitting processes developed here in all the granular components without preference for the ooids. Generally, when compared with other oolites having pitted ooids, the pitting in the Dunøyane Member is mostly superficial and restricted to isolated pairs, and rarely to three or more ooids arranged in more or less regular chains (Fig. 14*b,d*). There are no more advanced pits resulting progressively in a dense welding of the rock, and eliminating the interooid space (*cf.* Radwański, 1965, Pl. 18, Figs 3—6; the same phenomenon in organodetrital clasts — Weyl, 1959, Fig. 1b; Walpole & Carozzi, 1961, Pl. 1E; Carozzi & Roche, 1968, Fig. 3H; Reijers, 1972, Pl. 13, Fig. 5; Stricker & Carozzi, 1973, Pl. 1H). It may therefore be concluded (*cf.* Radwański, 1965) that a reasonable amount of fine-grained matrix was present in the Dunøyane oolites when the pressure-solution processes were in action.

CRACKED OOIDS

The cracking is most pronounced in ooids where they are in firm contact with their neighbours, and it may yield various effects. Where the external envelope is rigid, it is cracked in segments, some of which are pressed inwards towards the core, and some are roofed up at acute angles (Fig. 15*a*). In other ooids of the same envelope rigidity, only a pressing-in effect is realized (Fig. 15*b*) which however must correspond to much greater disturbances inside the ooid as a remarkable shortening of the surface takes place here. Both these examples show that the external envelopes of these ooids were much more rigid than their interiors. Of the latter, which was subsequently substituted by drusy dolomite, little may be recognized, but in the first of them the traces of internal envelopes are visible and these are cracked independently (arrowed in Fig. 15*a*). In this case, it is clear that the last but one envelope is not cracked, but only slightly arched; it is therefore concluded that the envelopes of at least some ooids were able to resist cracking, and they reacted in a plastic way.

The examples discussed (Fig. 15*a*—*b*) show the cracking to be a result of mechanical pressing by the neighbouring ooids which were attached to each other with pressure-solution contacts. The shapes of the cracked ooids match and adapt to the free space available, neighbouring ooids looking as if they were plastically kneading a less resistant one. Pressure-solution contacts reveal however that this is a secondary effect of mechanical action during development of the pitting processes in hard ooids.

An effect of a less rigid interior is also recognizable in the ooids having one depressed sector that is usually shaped like a flap that enters deeply into the interior. In the best example (Fig. 15*c*), the role of pressing neighbours is not clearly visible. The very sharp outline of the projecting tips that border the depressed flap suggests however that this is not a sedimentary structure of the kind obtained in experiments by Carozzi (1961a) or reported from the fossil state by Łabęcki & Radwański (1967; especially their Pl. 1, Fig. 5), but is a result of the stress by one of the neighbouring ooids (probably that at the top in the photo, the contact of which is not intersected by the plane of the thin section).

Another structure is obtained where the stress of the two neighbouring ooids results in a partial peeling-off and rippling of the last envelope, and therefore in formation of a void underneath. In the specimen presented (Fig. 15*d*), such an internal crack was formed in an abraded ooid, the remnants of the last envelope of which had been tangentially stressed by two neighbouring ooids whose pressure-solution contact is however also slightly touched by the plane of the thin section. A resulting swell is of the same kind as that recorded in pitting processes (*cf.* Radwański, 1965, Pl. 19, Fig. 1).

In addition to those ooids in which the shapes of cracks are controlled mostly by the concentric texture of envelopes, forms with radially oriented fractures also occur. These are usually simple cracks cutting the ooid to its centre or beyond. In the best example (Fig. 14*c*), the propagation of the crack from a pressure-solution contact is well expressed; along the cracked-off parts of the ooid a small dislocation and twisting took place. The general shape of the crack is identical to that formed by reciprocal impacts in the sedimentary environment (*cf.* Carozzi, 1961b; and from the fossil record — Łabęcki & Radwański, 1967, Pl. 1). In the specimen studied, the propagation of the crack and its displacement is instructive also in the respect to its filling. This is pure drusy dolomite, not spoiled by iron oxides as that of matrix, and therefore distinctly differing in its tint (*cf.* Fig. 14*c*, and the same ooid in upper part of Fig. 7); it is regarded as precipitated from a solution percolating through the ooid along the crack which was caused by the pressure-solution contact.

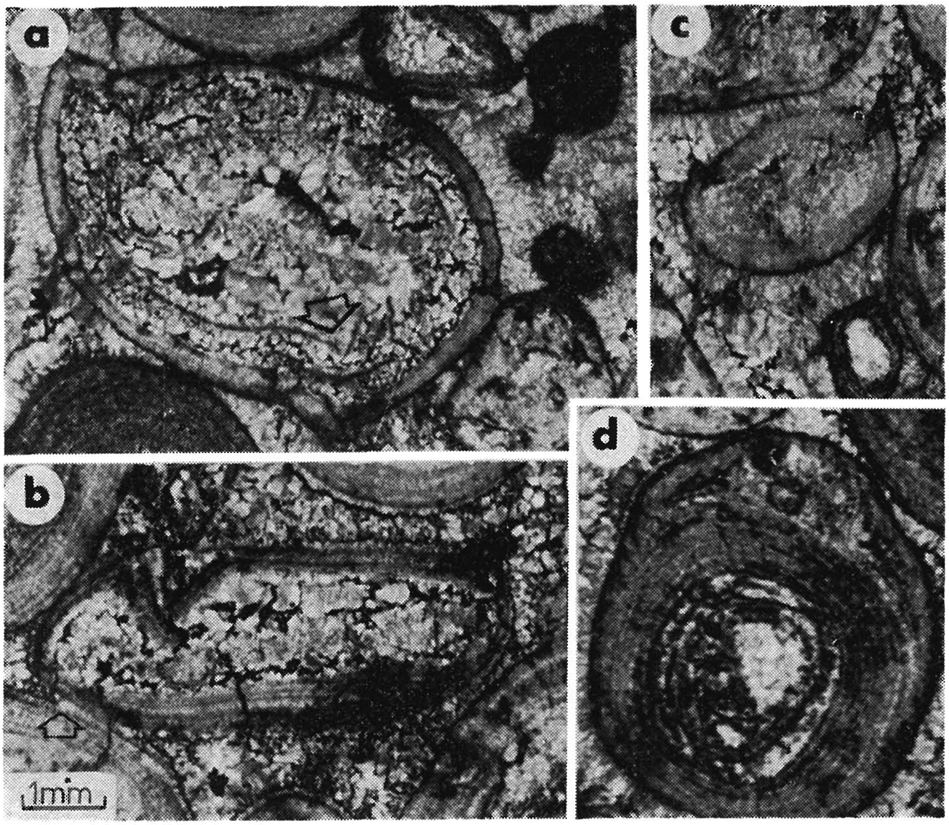
The cracks seen in ooids are assumed to have resulted from the action of mechanical forces associated with pressure-solution processes. In the two examples, viz. simple cracks (*cf.* Fig. 14*c*) and depressed flaps (*cf.* Fig. 15*c*) these are, however, of the same shape as those formed by mechanical agents during sedimentation (*cf.* Carozzi, 1961a, b; Łabęcki & Radwański, 1967).

The compactive stress acting through the contact of one ooid against another was analysed into two components: one operating in the direction of the principal compactive stress and resulting in the advance of pressure-solution processes; the second one operating in a perpendicular direction, i.e. tangentially to the ooid surface. The former is responsible for cracking either the whole ooid (Fig. 14*c*), or only its rigid envelopes and their in- or upward dislocation (Fig. 15a-c; *cf.* also Kettenbrink & Manger, 1971, Fig. 2C). The latter results in peeling-off the envelopes and their further disturbance (Fig. 15*d*). It may be noted that in the previously described similar mechanical disturbances of the pressure-solution contacts in ooids, only those of the latter group were observed as involving not only the rippling of external envelopes, but also their fracturing and tearing-off (Graf & Lamar, 1950; Radwański, 1965). A comparable spalling-off in ooids was regarded by Shearman & al. (1970) as originating during compaction; a schematic drawing (Shearman & al. 1970, Fig. 5) does not suggest, however, any possible mechanism of ooid deformation. A similar case was also referred to in this same way by Bathurst (1971, Fig. 245), while Knewtson & Hubert (1969, explanation to their Fig. 13) referred the spalling-off to a partial solution of some envelopes and their subsequent breakage due to compaction (although they noted the occurrence of pressure-solution contacts). But, in all cases where photographs of the peeled-off ooids and their neighbours were presented (Graf & Lamar, 1950, Fig. 14, and the same figure in Carozzi, 1960, Fig. 52; Radwański, 1965, Pl. 19, Figs 1—5; Knewtson & Hubert, 1969, Fig. 13; Coogan, 1970, Fig. 1F; Bathurst, 1971, Fig. 319), it is evidently visible that the fracturing and peeling-off was propagated from the pressure-solution contacts. Presumably, this is the only way in which the ooid peeling-off occurs during diagenesis in natural conditions; in experiments, however, it was obtained also by "dry"-pressure contacts (Fruth & al., 1966).

Within the Dunøyane ooids, it is noteworthy that the cracked ooids reveal traces of their radial texture which is manifested both in the shape of cracks in the external envelopes (Fig. 15*a-b*), and of those running throughout the whole ooid (Fig. 14*c*). Suggestions regarding the primary structure of some of the discussed cracked ooids are presented in a further chapter (*cf.* Fig. 19).

DISTORTED OOIDS

The analysis of pressure-solution contacts has also a bearing on the interpretation of the distorted ooids. As visible in these contacts, the distorted ooids are the succeeding or the final stages of subsequent disturbances in the ooid pairs or longer chains (*cf.* Figs 14, 15*a-b* and 16). The disturbances are of a striking similarity to boudinage structures, and may be explained by treating a pitted pair as subject to boudinage processes (*cf.* Fig. 17). The boudinage structures originate (Ramberg, 1955) due to stress, and result in the disruption of a competent (brittle) layer or other body embedded in a relatively incompetent (ductile) frame. In the



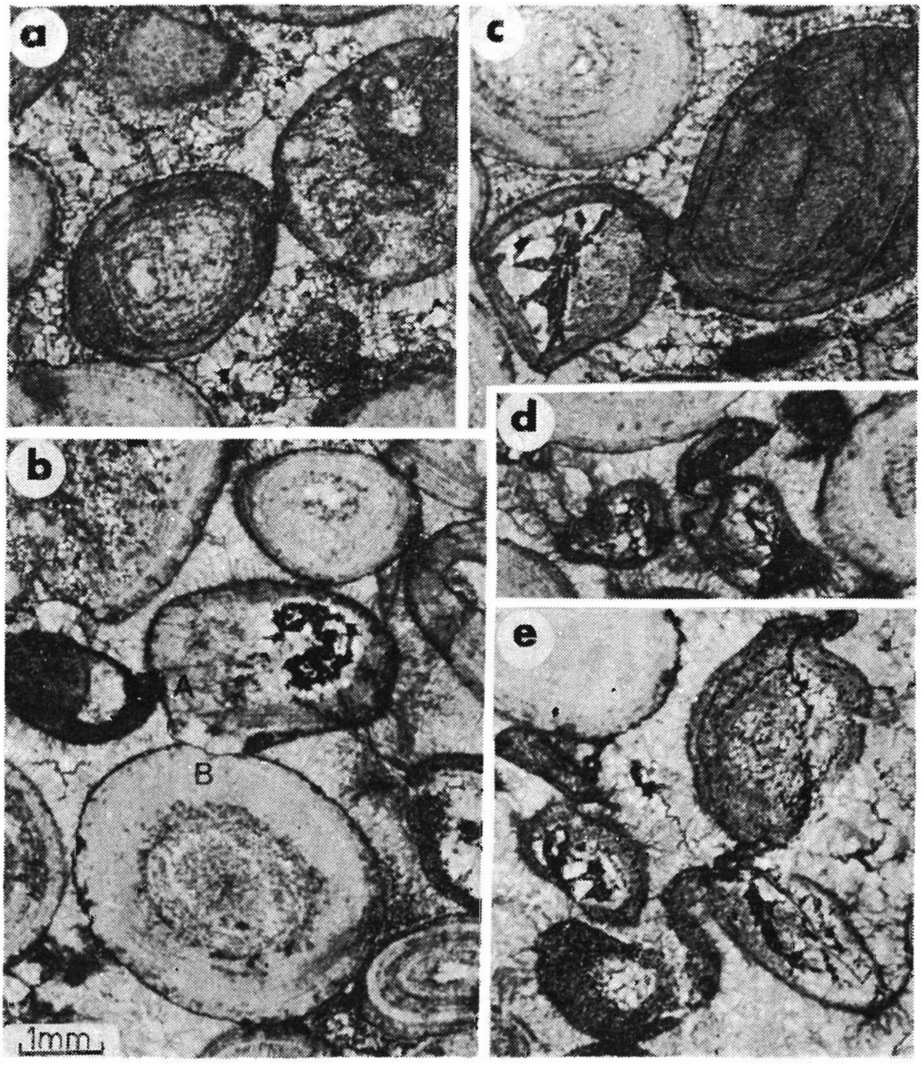
Ooids cracked by mechanical forces during pressure-solution action; locality Fanny-toppen

a — Ooid with the external envelope cracked in a few sectors, some of which are pressed-in to the interior, and some roofed upwards; arrowed are internal envelopes, either cracked or arched

b — Cracked ooid with a sector of the external envelope pressed inward (the envelope length shortened therefore by a distance of one pressed-in sector); pressure-solution contacts well visible, one of them (arrowed) being a dentate pit

c — Cracked ooid with a depressed flap, interpreted here as resulting from the stress of the overlying ooid, the contact of which is not intersected by the thin-section plane

d — Abraded ooid, the remnants of the last envelope of which were peeled-off by the contacting neighbours (their contacts slightly touched by thin-section plane). The resulting internal crack subsequently filled with fine-grained dolomite; the ooid also reveals a half-moon effect (its plane being almost vertical in the photo)



Ooids deformed by sedimentary boudinage — snouted and distorted ooids; locality Fannytoppen

a — Ooid pair with a pressure-solution contact and slightly pulled outwardly: the "snout"-contact effect appears. Another pressure-solution contact (bottom left) is undisturbed

b — Chain of three snouted ooids; snouts at points *A* and *B*

c — Pair of snouted ooids both of which are outwardly stretched at their contact. In the larger ooid (at right), an effect of its partial twisting is visible just at the contact, and it disappears progressively towards the more internal envelopes

d — Snout contacts in the ooids mostly elongated by tensile stretching: the shape typical of distorted ooids is realized

e — Typical distorted ooids: snout contacts are stretched into proboscid tips, more or less twisted, whereas the ooids themselves become split and contorted distorted ooids, a pitted pair or a longer chain is such a competent body, the adjacent matrix acting in an incompetent manner. A compressive stress that appears in a sedimentary sequence is the compaction pressure,

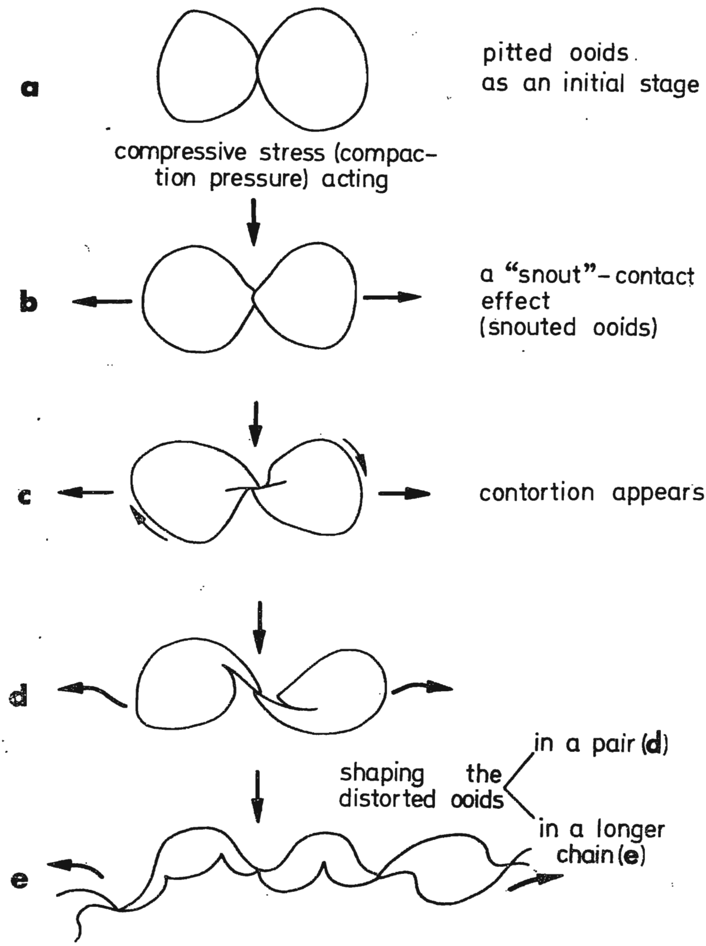


Fig. 17. Successive stages of the formation of distorted ooids

and the whole process should therefore be referred to "sedimentary boudinage" of McCrossan (1958).

When a compressive stress (compaction pressure) functions on the pitted ooids, the perpendicularly oriented tensile stress results in an outward-pulling of the pitted ooids (Fig. 17 *a-b)* and featuring them with a "snout" contact; the forms displaying this feature are called here "snouted ooids" (Fig. 16*a-c*). The pitting contact, having been firmly glued by pressure-solution, is stable enough not to be torn apart, and therefore only this contact remains relatively intact. The near-to--contact part of the ooids becomes initially misshaped (Fig. 17*b*). During further progress of the tensile stress, the outward-stretching of the pitted pair in the direction of this stress advances, but the pressure-solution contact remains furthermore relatively undisturbed; due to associated shear movement, more or less tangential to the ooid surface, a mutual rotation and contortion of both ooids is revealed (Fig. 17*c*). In the next stage, a strong disruption and contortion (Fig. 17*d*), or a splitting up of the ooids results (Fig. 17*e*), whereas the pressure-solution contact remains the point that firmly joins the ooids together. The final result, i.e. the distorted ooids (Fig. 16*d*-*e*), are homologous to the rotated boudins (*cf.* Ramberg, 1955, Pl. 3B and D) or some "pinch-and-swell" structures (also *cf.* Ramberg, 1955). It seems that the splitting of the ooids during the final disturbances may take place only when more than two ooids occur (*cf.* Fig. 17*e*); it results from the tearing-out and twisting of the ooids in various directions along the tensile-stress plane. The pressure-solution contacts become, during formation of the distorted ooids, also more or less stretched-out, but they remain not disrupted, and they then provide a characteristic linkage, often in a form of long, proboscid tips being a peculiar feature of the distorted ooids (*cf.* Cayeux, 1935; Carozzi, 1961a; Misik, 1968; Kettenbrink & Manger, 1971; Bachmann, 1973).

The presented mode of formation of the distorted ooids shows that the previous interpretation of these structures by Carozzi (1961a), as generated by reciprocal impacts of the ooids during sedimentation in agitated waters, is unacceptable. That interpretation, first postulated by Frantzen (1887), and subsequently held by Turnau-Morawska (1961) and recently by Sarkar (1973) has never been critically discussed; in the 1960's it seemed to be proved by Carozzi's experiments on artificial models.

The two of the four models presented by Carozzi (1961a, Fig. 12*a, b*) explain however only the shape resulting from the penetration of a depressed flap (*cf.* Carozzi, 1961a, Fig. 11). Such depressed flaps are really the mechanical structures and they are formed either by a sedimentary impact or by compaction pressure (*e. g.*, at the pressure-solution contacts such as presented in Fig. 15*c*) that pressed down a section of rigid external envelopes, but they are not in any way transitional forms to the distorted ooids. Such forms have commonly been observed, for example, in a set of synsedimentarily broken ooids reported from the Upper Triassic (Łabęcki & Radwański, 1967). Also mechanically destroyed during sedimentation are the ooids counterparting a model with a depressed flap that underwent curling (Carozzi, 1961a, Fig. 12c). The last of the four models (Carozzi, 1961a, Fig. 12d) is to some extent morphologically similar to the distorted ooids by its splitting into halves, but it cannot explain a proboscid shape of the ooid tips. Furthermore, the discussed models demonstrate neither *why*, nor *how* the distorted ooids were able to find each other during sedimentation or deposition and to match and join themselves exactly by these very proboscid tips ("apophyses" of Carozzi, 1961a). Neither can they explain how it was possible during rotation and twisting of the ooids that these linkage points were stable enough not to have been torn apart while the ooids themselves underwent rupture and strong deformation. The discussed proboscid tips and their linkage points, as revealed by illustrations in previous papers (see reproduced *in:* Carozzi, 1961a), are the most resistant parts of ooids during deformations whereas, if following Carozzi's interpretation, they should be expected the most friable and the weakest, as well as associated with underdeveloped forms (*i.e.* those which underwent distortion, but which were not hooked together with their neighbours); moreover, in strongly agitated waters such proboscid tips should easily be broken in a moment, and prior to a more serious damage of the remaining parts of the ooids. On the other hand, it is also instructive that no evidence has been presented as to the strongly agitated environment during deposition of the layers containing the distorted ooids which are commonly composed of such unresistant substances as iron ores (mostly chamosite), phosphate or clay minerals (*cf.* Cayeux, 1922, 1935; Déverin, 1945; Carozzi, 1961a; Jones, 1965; Gygi, 1969).

The sedimentary-boudinage hypothesis for the formation of distorted ooids explains all their discussed features, answering the question why distortion, as known from most of the hitherto reported occurrences, is confined to non-calcareous ooids or pseudoooids embedded in the matrix of a different mineralogical composition. In such circumstances the differences in competence necessary to yield boudinage structures (*cf.* Ramberg, 1955) are the greatest (*e.g.*, differences between chamosite ooids and calcitic or sideritic matrix). In carbonate ooids with either micritic or sparry carbonate cement, the competence differences must be comparatively less, and distorted ooids are therefore much rarer or not well developed.

The sedimentary-boudinage hypothesis explains also another peculiar feature of the distorted ooids, viz. their distribution in small patches (*cf.* Fig. 16*e* and the same object in Fig. 7 at left upper corner) or in isolated portions of the oolitic layers, and appearing more or less parallel to the bedding, as recorded by Cayeux (1922, p. 171; 1935, p. 233) and subsequent authors (Carozzi, 1961a; Kettenbrink & Manger, 1971; Sarkar 1973) which would be unlikely when accepting a sedimentary origin for the disturbances in agitated waters. During development of the boudinage processes such a distribution is quite normal as it reflects areas of discrete physical conditions. The same kind of localized occurrence is recognized in pitted ooids (Radwański, 1965) that formed in a layer only in those places where local conditions were favourable for pressure-solution action (dense packing, few or no matrix). It should be mentioned that it was already Cayeux (1922) who stated that distorted chamosite ooids occurred in ironstone beds in those places where the ooids were densely packed and matrix was carbonate. The boudinage hypothesis can explain both these facts, as in densely packed parts the pressure-solution processes might have been functioning while a heterogenous matrix offered conditions of different competence; the sedimentary hypothesis could not demonstrate in this case how the distorted ooids were densely packed only in some places, and why these were restricted to heterogenous matrix.

A post-depositional origin for the distorted ooids has already been postulated, although very briefly by some previous students. A longer discussion was presented only by Cayeux (1935) who rejected a sedimentary origin for these ooids and considered them as formed by their being stretched into a layer, their hard-cemented contacts remaining intact. Déverin (1945) regarded distortions in chamosite ooids as diagenetic, which was, however, inevitable since he regarded the ooids themselves as post-depositional. In carbonate ooids, Scavnicar & Susnjara (1967) and Misik (1968) considered the distortions as formed by collapse during compaction. More recently, Kettenbrink & Manger (1971) and Zenger (1972a, b) suggest their origin as due to early plastic deformation attributed to compaction, whereas Bachmann (1973) refers them to a slip-shearing motion during deformation of the deposit. A similar possibility was also suggested by Jones (1965) for ferruginous ooids. To sum up, it may be concluded that the reasonable opinions, expressed in different words or paying attention only to some aspects, are recognizable in every one of these conclusions if the boudinage mechanism is kept in mind.

It may also be recalled that distorted structures in bioclasts and onkolites in a biocalcarenite were reported by Flügel (1966) who regarded them as possibly early diagenetic and comparable to the distorted ooids (Flügel, 1966, p. 22); in a very instructive photo (Flügel, 1966, Pl. 3, Fig. 1) which comprises pitted, snouted, as well as distorted bioclasts, it is clear that none of these granular components could have been distorted in a cracking-ball way during sedimentation, such as postulated by Carozzi (1961a). The same may be said about the distorted intra- and bioclasts ("pseudoooids") presented in his classical paper by Bornemann (1886, Pl. 7, Fig. 1).

The general conclusion in the hypothesis presented here is that the distorted ooids, the same as other distorted elements, result from the boudinage process that followed the pressure-solution. These two processes need however different conditions: the latter can develop only when the ooids contact each other and are not cemented, whereas for the former the matrix is a prerequisite for plastic deformation. The matrix which is the ductile frame must be, moreover, of different competence than the embedded ooids. Its appearance in volume sufficient to make the ductile frame may be explained as resulting from a decrease of the inter-ooid space when the pitting interpenetration of ooids progresses. It may be thought that at this very moment, when the inter-ooid space becomes filled with the matrix, the pitting processes have to vanish as they cannot result in a further decrease of free space. If the matrix is missing, the pitting process continues till the whole space is occupied by interpenetrating ooids (*cf.* Radwański, 1965). If the matrix occurs in an amount sufficient to stop the pressure solution, and this is the situation in the investigated oolites, further diagenetic events may set up the distortions, as follows (*cf.* Fig. 18: an "isolated" pair of pitted ooids is taken for demonstration).

If the matrix is of the same competence as the pitted pair of ooids, no distortion would develop (Fig. 18, *Ia-b*), but if it is of different competence, the ooids can undergo boudinage disturbances and distorted ooids may be produced (Fig. 18, *IIa-c*). At this phase, the deformations rise only in pressure-solution-contacted ooids, either arranged in pairs or chains or larger clusters, while the adjacent discrete ooids remain unaffected.

A commonly observed healing of the distortion structures by sparry carbonates (Carozzi, 1961a; Scavnicar and Susnjara, 1967, Pl. 3, Fig. 4; Misik, 1968, Figs 14—15; Kettenbrink & Manger, 1971, Fig. 3D-F; Bachmann, 1973, Fig. 61; Sarkar, 1973, Figs 5—7), which is also evident in the Dunøyane ooids (*cf.* Fig. 16), is a further stage in the diagenesis of the ooid-bearing deposits.

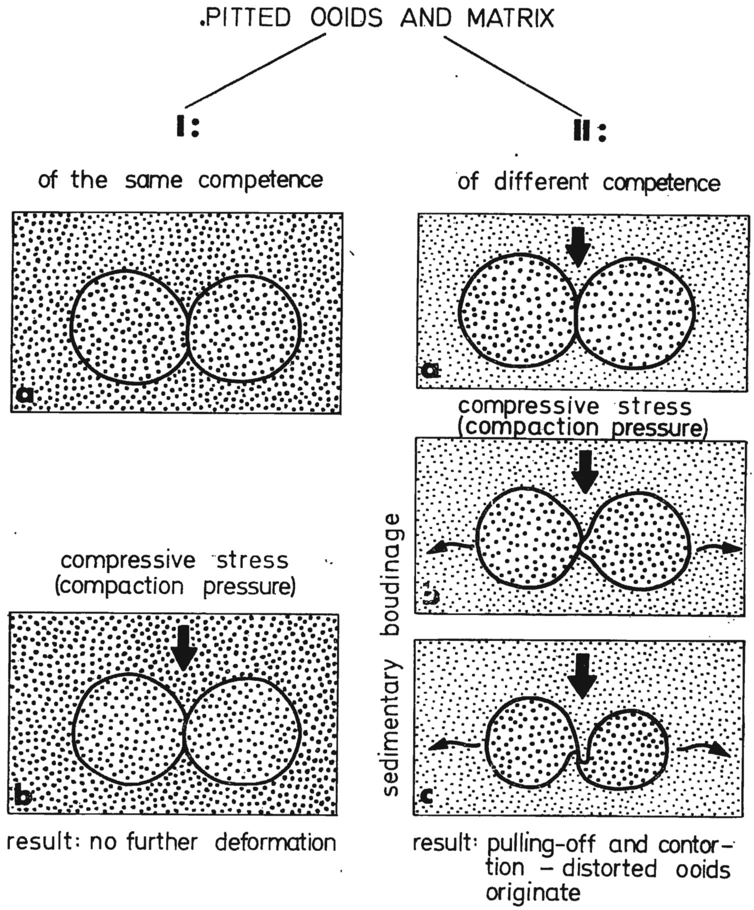


Fig. 18. Possible effects of a further advance of diagenetic processes in pitted ooids, related to the mutual competence of the pitted ooids and of the matrix

NATURE OF THE DUNØYANE OOIDS

The nature of the Dunøyane ooids and pisoids attaining an extreme size about or more than 1 cm in diameter should generally be discussed as most of the so large Precambrian ooids/pisoids owe their origin to the activity of blue-green algae, and they are therefore referred to as algal pisoids or spherical stromatolites (onkolites).

The Dunøyane ooids display the features which are rarely or never met in onkolites, viz.: (*1*) Radial fractures during formation of both abraded and broken ooids (Figs 9—10), as well as of cracked ooids (Figs 14*c* and 15*a-c*); (*2*) Presence of concentric envelopes only continuous; (*3*) Half-moon texture which originated by a juxtaposed deposition of carbonate and supposedly sulphate envelopes; (*4*) Evidence of mechanical abrasion itself.

Within these features the most important is (*1*), as the primary radial texture is only met in physico-chemical ooids, as obtained in experiments by Monaghan & Lytle (1956) and Suess & Fütterer (1972), and as found in some hypersaline environments (Loreau & Purser, 1973; Friedman & al., 1973, Sandberg, 1975); of the same nature is suggested also the radial texture in some ancient ooids (Kahle, 1974; Sandberg, 1975), unlikely to the popularly accepted older opinions on its diagenetic origin (see discussion *in:* Eardley, 1938; Carozzi, 1962; Shearman & al., 1970; Friedman & al., 1973; Kahle, 1974; Sandberg, 1975). The remaining features are not definitely instructive, as: (*2*) is also typical of the onkolites with concentric lamination (mode "*C*" of Logan & al., 1964; *cf.* also Radwański and Szulczewski, 1966; Radwański, 1968; Hofmann, 1969) which however never display this feature in an ideal state, and never developed so selectively as in the investigated ooids; (*3*) may theoretically also happen in bluegreen algal structures, as these can develop in hypersaline environments (*cf.* Logan, 1961; Garrett, 1970; Hofmann, 1973); (*4*) is also possible in onkolites which are known as being hard and completely lithified on the sedimentary bottom (*cf.* Kutek & Radwański, 1965; Gygi, 1969; Monty, 1973). The up-to-date recognition of the spherical, carbonate chemical and bluegreen algal structures allow us to regard the set of the discussed features as indicative of an inorganic, physico-chemical origin for the Dunøyane ooids, and for their growth by a carbonate accretion in the sedimentary environment. The three characters of these ooids, namely their size, cement, and primary nature of the carbonate mineral, require more detailed consideration.

Various spherical concretionary bodies which may originate in caliche or other subaerial or subsoil crusts, are evidently of so different structure (*cf.* Nagtegaal, 1969; Bernoulli & Wagner, 1971; Castellarin & Sartori, 1973a, b) that they cannot be taken into account as comparable to the Dunøyane ooids.

OOID SIZE

The remarkable size of the Dunøyane ooids is their most outstanding and intriguing feature. In Recent marine environments, there are no records of such large ooids (*cf.* Bathurst, 1971); on the Bahama shoals, the median diameter ranges between about 0.25 and 0.42 mm, and the maximum is up to 1.0 mm (Newell & al., 1960). Larger forms are known only from some coastal and sabkha settings of the Persian Gulf (Loreau & Purser, 1973; Shinn, 1973); on the other hand, also from hot-water springs and travertines, as well as from caves and artificial circulation ducts (references *in:* Maubeuge, 1964; Donahue 1965, 1969; Kettenbrink & Manger, 1971; Gradziński & Radomski, 1967, 1976). In ancient sequences, there are only a few reliable records of small pisoids slightly exceeding in their average dimensions the upper ooid size (*e.g.*, Sarin, 1962; Akin & Graves, 1969; Kettenbrink & Manger, 1971; and aforementioned data presented by Usdowski, 1962; Krebs, 1968, 1974; and for half-moon ooids by Carozzi, 1963), and thus being much smaller than the Dunøyane bodies.

Regarding the ooid growth, it was assumed in previous years that the most important factor controlling the upper size is the water agitation (Cayeux, 1935; Illing, 1954; Carozzi, 1957, 1960; Newell & al., 1960). More recent data shows that this is not essential: Freeman (1962) reported quiet water ooids from Laguna Madre in Texas; Bathurst (1967, 1968) noted a growth of very thin oolitic coats in stagnant conditions of Bimini Lagoon, Bahamas; Suess & Fütterer (1972) obtained ooids in experiments either with very gentle agitation or without any water motion; non--agitated are also some cave pearls such as in rimstone dam occurrences (Donahue, 1969). Consequently, Bathurst (1968) supposes that the ooid upper size reflects rather a balance between precipitation on the ooids and their abrasion by mutual impacts. The abrasion increases when the ooids grow larger, and the resistance of the water to the motion of ooids becomes relatively smaller. If this is so, then this latter agent, the water resistance, should be postulated as a decisive factor responsible for enormous growth of the Dunøyane ooids: the precipitation is therefore suggested as being more plentiful, thus changing the water into a carbonate-milky solution, the resistance of which became greater. On the other hand, such a solution due to greater hydraulic buoyancy could keep the ooids in suspension longer when they were increasing in their size. These two combined factors might have promoted advanced growth of ooids and protected them both from abrasion and from being cemented together. The agitation itself, in such a case, was a subordinate agent and its value could certainly not be greater than in any Recent environment. A similar mechanism (*cf.* Donahue, 1965, 1969) is supposedly decisive for the enormous growth of hot-spring pisoids and some cave pearls.

The presented conclusion therefore regards the Precambrian carbonate precipitation as more plentiful than nowadays. It is highly likely that this was so, as there is abundant evidence of a greater role of carbonate sedimentation at that time. It is apparent, both from the thick carbonate sequences in many parts of the world, and especially from the enormous growth and geographic extent of such carbonate biosedimentary structures as stromatolites (*cf.* Monty, 1973; Gebelein, 1974; Cloud & al., 1974). As factors responsible for huge expansion of these bluegreen algal structures, particularly in the Late Precambrian, such varying agents have been suggested as, for example, greater tidal ranges or the resistance of the bluegreen algae to ultraviolet radiation. Recently, Garrett (1970) and Awramik (1971) claimed that a lack of metazoan browsers feeding on bluegreen algal mats, and of burrowers destroying the lamination, might have been important (*cf.* also Hofmann, 1973; Gebelein, 1974); the idea has however been objected by Monty (1973). Nevertheless, as shown by Logan (1961), widespread stromatolitic fields with domes a yard or so high, and thus comparable to the Precambrian forms, originate at present in a restricted environment of Shark Bay in western Australia where hypersaline conditions prevail and stromatolitic growth is limited by tidal range. It therefore seems that either the Precambrian stromatolites were also hypersaline, or more likely the stromatolite profusion in Precambrian time reflected a generally more plentiful precipitation of carbonate ooze which was trapped and bound in great quantities by bluegreen algae. These certainly played a much greater role in littoral niches than today (*cf.* Monty, 1973), but the greater and more profuse bluegreen algal vegetation should be able to involve considerably larger structures only when greater supplies of trapable ooze were available. The absence of the metazoan browsers or grazers, postulated by Garrett (1970) and Awramik (1971), should be at most regarded as responsible for the preservation of successive algal mats in stromatolites rather than for their origin and attaining enormous size (*cf.* also Monty, 1973).

Under conditions of more plentiful carbonate precipitation in the sea water, the precipitation ratio was also high in a time interval, as compared to Recent environments — a greater amount of the precipitate could have been bound to the ooid surface, and much less abraded. This resulted in a more advanced growth of the ooids and their final dimension being much larger than those formed during Phanerozoic periods.

In the Dunøyane Member, the half-moon ooids which supposedly owe their structure to the presence of sulphate envelopes, do not generally attain the largest size; the enormous size of all the Dunøyane ooids cannot therefore be attributed to participation in the ooid growth of compounds others than the carbonates.

CEMENT IN OOLITIC DEPOSITS

The dolomite cement is now almost completely sparry, secondary in origin, and only in some samples are there either relics of a previous pelletal structure (Fig. 5*B*), or the fine-grained matrix (Figs 6*B* and 13). In one sample, the matrix is micritic and unaltered during subsequent processes (Fig. 12). When analysing the pitting processes, their slight advance shows that certainly the interspaces were filled with a deposit prior to the moment that the pressure-solution commenced. The same may be concluded from loosely packed oolites (Fig. 10) in which an immense interspace must have been filled with a deposit before the formation of sparry cement. It seems therefore reasonable to suggest the presence of primary, either micritic, or even pelletal and fine-grained matrix in most, or maybe all the Dunøyane oolites.

The sparitization of this primary matrix occurred when the pressure-solution and related processes (cracking and distorting of ooids) were completed. The structure of the primary matrix and its porosity caused an incomplete filling of the interspace during recrystallization: in many places the sparry dolomite is drusy, and the interiors of druses remain open. Small admixtures of iron compounds that had been dispersed in the primary ooze were then collected in the cavities of druses (*e.g.*, Fig. 4—5 and 10—11).

During appearance of the sparry cement, other voids were also filled with dolomite that precipitated from percolating solutions; these are various vugs in ooids, including the half-moon ones (Figs 10—13), and cracks, either in ooids (Fig. 14*c*) or in the deposit (small veins in Fig. 12).

Generally, the formation of the sparry dolomite cement was the last in the sequence of various post-depositional processes in the Dunøyane oolites, the diagenesis of which was then completed.

NATURE OF PRIMARY CARBONATE

A few lines of evidence as to the composition of the primary carbonate mineral in the ooids suggest that it was dolomite.

The first one is the varied, and strongly differentiated in detail, structure of particular ooids, and of particular envelopes in a single ooid (*cf.* Figs 7—8 and 10); if there was a general substitution of the oolitic deposit by allochthonous solutions, the structure of the substituted bodies should become more or less unified. The second is the absence of any traces of penetration of dolomite from the interspaces into the ooids; the ooids seem to have been dolomitic before the formation of the sparry cement, and thus reactions at the ooid/cement boundary could not take place. The third is the presence of half-moon ooids, the occurrence of primary sulphates in which is assumed to be responsible for their peculiar structure; the sulphates must have precipitated in a hypersaline environment where the potential of precipitation for dolomite is the highest of all the carbonates (*cf.* Friedman & Sanders, 1967).

If the primary mineral precipitated from the sea water was dolomite, a slight pronouncement of radial texture in ooids might had simply resulted from the mode of accretion different than that of calcium carbonates. On the other hand, if the abovediscussed plentiful precipitation was dolomitic, it might had controlled other specific conditions of the environment (buoyancy, resistance of ooids to abrasion). The final result of the rock-forming processes, i.e the oolitic Dunøyane dolostones should therefore be regarded as composed of the two generations of dolomite: primary in ooids and secondary in cement. In this respect, the Dunøyane dolostones closely resemble common calcitic oosparites.

In regard to the primary nature of dolomite in the south Spitsbergen sedimentary area, it is also instructive that primary calcareous deposits are well preserved in the Fannytoppen Member, the oldest unit of the Höferpynten Dolomite Formation. A similar situation appears moreover in other parts of Spitsbergen, e.g. in Ny Friesland (C. B. Wilson, 1961). It is therefore demonstrated that the general, world-wide tendency for replacement of limestones by dolomites with progressing age in the Precambrian is not evidenced here, and the limestone/dolomite succession in the investigated sequence (Fig. 3) may really be as it was in the sedimentary basin where both primary calcium-carbonate and dolomite could precipitate in various periods.

SEDIMENTARY CONDITIONS

The simple, undisturbed ooids of the Dunøyane Member certainly originated in a normal, oolitic environment which was characterized (*cf.* Donahue, 1965, 1969; Bathurst, 1968, 1971) by a supersaturated solution, its agitation, and by a splash-cup condition. The presence of detrital nuclei, regarded by Donahue, and by Bathurst, as the fourth of these prerequisities, is not well evidenced here, except for the half-moon ooids. In most of the ooids, if not obliterated during diagenesis, the nuclei are very indistinct and only slightly contrasted with the envelopes. It shows that the core was composed of the same material, and was of a very small size as compared with the total dimension of the ooids; it is possible also that air bubbles acted here as the initial locus of precipitation. If derived from another deposit, the nuclei were either micritic or containing more or less sulphates, and thus being the transitional forms to the cores of the half-moon ooids.

The half-moon ooids, if their soluble envelopes were really sulphate, must have formed in a more restricted area in which at least temporary evaporation occurred. Plenty of the half-moon ooids were redeposited in a normal oolitic condition, and it is assumed that the degree of the half-moon advance corresponds reversely to the distance of redeposition: in an intraclast (Fig. 12) derived from the proper half-moon-ooid facies, this effect is the highest; in the facies into which the intraclast was brought (Fig. 12, top), only the cores are half-mooned whereas their coating is composed of normal carbonate envelopes; finally, in normal oolitic facies (Fig. 10), the half-moon cores are smaller and smaller, and present in an insignificant number of ooids. It should therefore be interpreted that the deposits of the half-moon facies were temporarily subjected to erosion and redeposition. Their more lithified portions supplied intraclasts, while the loose ooids were swept out and brought into carbonate facies. In the latter they were either the dominant components (Fig. 12), or became more and more dispersed amongst normal ooids (Fig. 10) until they gradually disappeared (Figs 7—8).

The discussed range of conditions is attributable to a facies pattern from lagoonal hypersaline with periodical sulphate precipitation, to offshores of open sea regions (*cf.* Fig. 20). The latter are postulated here, as they better correspond to a continuous sequence of the deposits studied and to an absence of any evidence of subaerial processes which should be expected in tidal flats, widespread lagoons or other non-marine conditions.

Within the discussed sedimentary area, redeposition also occurred in the normal oolite facies, and resulted either in supplies of oolitic lumps (grapestones) or intraclasts derived from lithified oolitic or pelletal deposits (*cf.* Figs 7—8). All these components, if not too heavy to be suspended, were coated by common envelopes and became complex ooids (*cf.* Carozzi, 1964). The largest of the latter, some of which being multiphased (Fig. 6*A*) attain, or slightly exceed, the maximum size of simple ooids (*cf.* Figs 5 and 10), while the largest intraclasts and grapestones remain uncoated (Fig. 5*A-B*). It therefore appears that the agitation that kept the bottom particles in motion was here responsible for a maximum accretion of oolitic envelopes (as postulated by Carozzi, 1957, 1960), the enormous size of which was however due to other agents (see discussion on ooid size). A situation where some of the largest complex ooids slightly exceed the size of simple ooids, if not accidental in the thin-section planes, may be explained by a smaller weight of some more porous intraclasts or lumps.

Another possibility which should be taken into account here is a redeposition of the material after its complete oolitization; as a result some samples may be interpreted as composed of more allochthonous material, and therefore not so well sorted (*e.g.*, Figs 5*A* and 7—8) as the other, more autochthonous ones (*e.g.*, Figs 5*B* and 10). An advanced mixing of various oolitic components which have been stirred up and transported is evident only in some samples (*e.g.*, Fig. 6, mostly 6*B*).

A general accordance of the oolitization processes with those recognized by Carozzi (1957, 1960) in Recent environments is an additional premise for the physico-chemical, accretional nature of the Dunøyane ooids. On the other hand, the numerous intraclasts and lumps or grapestones evidence a common reworking, often multiphased (*cf.* Figs 4 and 6), and therefore a high-energy environment, more comparable to that occurring at present on the Bahamian shoals (*cf.* Illing, 1954; Newell & al., 1960; Purdy, 1963; Ball, 1967) than to those of quiet-water oolitic growth (*e.g.* that presented by Freeman, 1962).

Concerning the absence of the bluegreen algal facies in the sequence of the Dunøyane oolites, it may be noted that this facies, so typical of related nearshore facies of the Precambrian (*cf.* Hofmann, 1969, 1973; Garrett, 1970; Awramik 1971; Monty, 1973; Truswell & Eriksson, 1973; Eriksson & Truswell, 1974; Gebelein, 1974), is well developed both in the underlying Wurmbrandegga Member and in the top part of the Dunøyane Member above the oolites, though not necessarily in the same section (*cf.* Birkenmajer, 1972; and Figs 2—3 herein). It is therefore suggested that during the formation of the Dunøyane oolites the situation was similar to that occurring at present on the Bahamas where the oolite and the bluegreen algal facies are mutually exclusive as a result of different required hydrodynamic conditions (*cf.* Newell & al., 1960; Purdy, 1963; Bathurst, 1967; Monty, 1967, 1972). Consequently, the intermediate spherical structures which might have been formed by a simultaneous function of both physico-chemical (oolitic) precipitation and of bluegreen algal activity are unknown, and the role of these algae on the ooid surface is passive here (or negative, as it is mostly perforating — *cf.* Illing, 1954; Newell & al., 1960).

In the Dunøyane oolites, no structures are recognized which might have been attributed to the bluegreen algal activity, either in the formation of some bodies or of their particular envelopes. This allows us to infer a temporary expelling of the bluegreen algae from the sedimentary areas of the Dunøyane Member.

In connection with the above statements it should be noted that there are recently reported (Friedman & al., 1973) algal mats associated with oolites, as well as the transitional forms between ooids and onkolites, from a hypersaline lagoon in the coastal region of the Gulf of Aqaba, northern Red Sea. These transitional forms are regarded by Friedman & al. as precipitated by algae. Although no specimens are available for comparisons, it may be stated that either there are quite exceptional conditions there, or the presented interpretation is unjustifiable, and the discussed transitional forms are really the oolitized onkolites or the ooids containing temporary onkolitic envelopes of the same kind as reported from some oolitic deposits both modern and ancient (*cf.* Wood, 1941; Newell & al., 1960; Radwański, 1968).

EARLY LITHIFICATION PROBLEMS

A common occurrence of oolitic intraclasts and lumps in the Dunøyane oolites shows that the originating oolitic deposits were here more or less lithified prior to their temporary reworking and redeposition. The intraclasts were hard, as their boundaries are sharp and particular ooids are distinctly abraded (*cf.* Fig. 5*A*), whereas in the lumps the matrix was not so resistant, being partially washed out which resulted in shaping them like bunches of grapes. The broken or abraded ooids also show distinctly that at least some simple ooids were hard, rigid bodies during reworking.

A more intricate pattern is revealed by half-moon ooids, some envelopes of which were supposedly composed of sulphates. The settling of their envelopes appears to the quite plastic, pointing to their being in a soft state. But, the half-moon cores of simple and complex ooids were certainly rigid when formation of common envelopes, composed of carbonates, progressed: in the half-moon grapestones used as a core, the common envelopes match tightly the particular ooids, although these have a very thin shell covering the sulphate-bearing interiors (*cf.* Fig. 13). In the intraclasts with half-moon ooids, the marginal ooids are either sharply cut or even incised (*a* in Fig. 12) evidencing their firm consistency. In all these cases, the plastic settling of envelopes occurred after final deposition of the ooids on the bottom, as is evident from the same orientation of the half-mooned parts both in the intraclasts and in the ooids of the background (*cf.* Fig. 12). The same happened to half-moon ooids brought into a normal oolite facies (Fig.10), and it should be also inferred for the half-moon grapestones (Fig. 13). The most probable interpretation is therefore that the half-moon ooids, loaded with sulphate laminae, were generally hard and rigid during reworking, but the embedded carbonate interlaminae were then unlithified. During diagenesis, when the sulphates become dissolved due to their not being in equilibrium with the carbonate solutions percolating through the host deposits, these unlithified envelopes began readily to collapse and plastically deform. It is hardly recognizable if their lithification appeared soon thereafter or much later when the drusy dolomite began to precipitate in the voids. The latter does not obliterate the collapsed envelopes and looks as if introduced after lithification of these envelopes.

The suggested sequence of events explains also a feature of some ooids which settled complete (Fig. 11). As previously discussed, the particular envelopes of these ooids seem not to have been composed of, but rather interwoven with sulphates. It is therefore probable that they were really soft and began to collapse softly due to their own weight or that of the overburden when the sulphates were being dissolved. Only in these ooids might the pure compaction effect have therefore been pronounced in any of the Dunøyane oolites.

The structure of the sulphate-bearing ooids suggests that if the carbonates of the envelopes were interbedded or interwoven with sulphates, they became prevented from early lithification. A sure explanation of this feature cannot however be offered.

In respect to the sulphate-bearing ooids, the structure of the cracked ooids is also instructive. These ooids are interpreted as resulting from the secondary effects of the pressure-solution processes (*cf.* Fig. 15). These processes can only develop in hard, rigid bodies, such as the external envelopes of the cracked ooids (*cf.* Fig. 15*a-b*). Some of the presented ooids bear fragments of internal envelopes plastically deformed (arrowed in Fig. 15*a*). Certainly the complete interior of these ooids must have been disturbed in the same way, because otherwise they could not contain the cracked and depressed segments of the last envelopes. In the rocks investigated, the interiors are filled (*cf.* oomoldic porosity of Friedman, 1964) with drusy dolomite, and it is thus possible that the sulphates were here substituted in a fashion similar as in the half-moon ooids. The sulphates might have been acting plastically when the cracked envelopes were pressed inwards in the ooids. It is therefore assumed that such ooids (Fig. 15*a-b*, to smaller extent also 15*c*) with a sulphate or sulphate interwoven core were those in which the cracking and depressing of the outer shell could develop (Fig. 19). Consequently, the leaching out of the sulphates is regarded

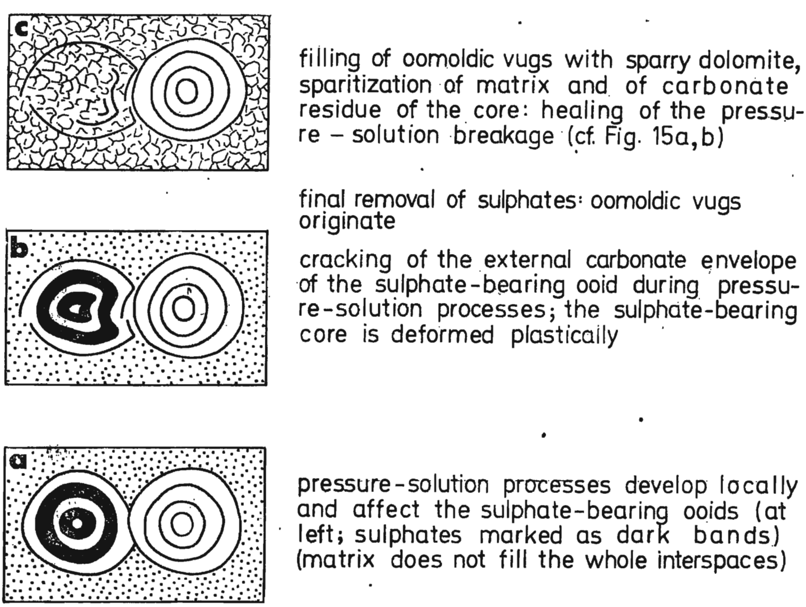


Fig. 19. Suggested stages (*a*, *b*, *c*) of the cracking of sulphate-bearing ooids

as not being completed when the pressure-solution processes began. It means that the pressure-solution appeared very early in the Dunøyane ooids, a feature typical of many other oolitic deposits (*cf.* Radwański, 1965).

AN ENVIRONMENTAL MODEL

The sedimentation of the Dunøyane oolites is attributed to a pattern of facies ranging from shallow subtidal, through intertidal, to the supratidal lagoons. In all of them the oolitic facies was dominant during the formation of the sequence.

In supratidal lagoons, hypersaline conditions were temporarily established and they resulted in the precipitation of sulphate cores or envelopes in ooids. The lagoons were periodically submerged when larger marine floods encroached: under such conditions agitations became stronger, the reworking of deposits progressed, and either intraclasts or loose ooids with sulphate contents were swept into the marine facies (*cf.* Figs 12—13). In the latter case, it is also possible that the carbonate growth over sulphate-bearing bodies occurred in the lagoons themselves when the encroaching waters diluted the brine and carbonate precipitation could return.

The marine facies were favourable for a continuous growth of ooids over wide areas of the intertidal and shallow subtidal zones where the conditions of a splash cup were met, and they protected the region against the diluting of supersaturated waters by currents and the input of non--saturated water. In these areas, the high-energy conditions often resulted in reworking of the oolitic deposits, and in the formation of intraclasts or lumps, some of which became the cores for further, even multiphased ooid growth (*cf.* Figs 4—6). When the agitation was not so strong, oolitic sand with simple ooids was formed (*cf.* Figs 7—8 and 10); but, also in these areas the presence of either half-moon (*cf.* Fig. 10), or broken ooids (*cf.* Figs 7—8) is apparent. The former were evidently derived from hypersaline lagoons and, as discussed before, the latter probably were also. The lowest energy were certainly the pelletal and micritic (mud) facies, both of them however being influenced either by supplies of coarser material (*cf.* Figs 5*B*, 6*B* and 12—13), or by reworking (*cf.* Figs 4 and 7—8). It is therefore reasonable to consider all the discussed facies as an interfingering pattern, the particular facies of which were non-uniformly distributed along the shore, and could also superpose and variously overlap each other through time. The presented environmental model, pictured by a cross--section (Fig. 20), exhibits a conjectured pattern of facies in their temporary lateral succession.

A similar facies pattern of carbonate sedimentation, from shallow subtidal to inter- and supratidal, has recently been recognized in other regions of the Precambrian epeiric seas, viz. in South Africa, as documented by Truswell & Eriksson (1973, 1975) and Eriksson & Truswell (1974) for the famous stromatolite-bearing sequences of the Transvaal Dolomite and Malmani Dolomite; most of these environments were however coloni-

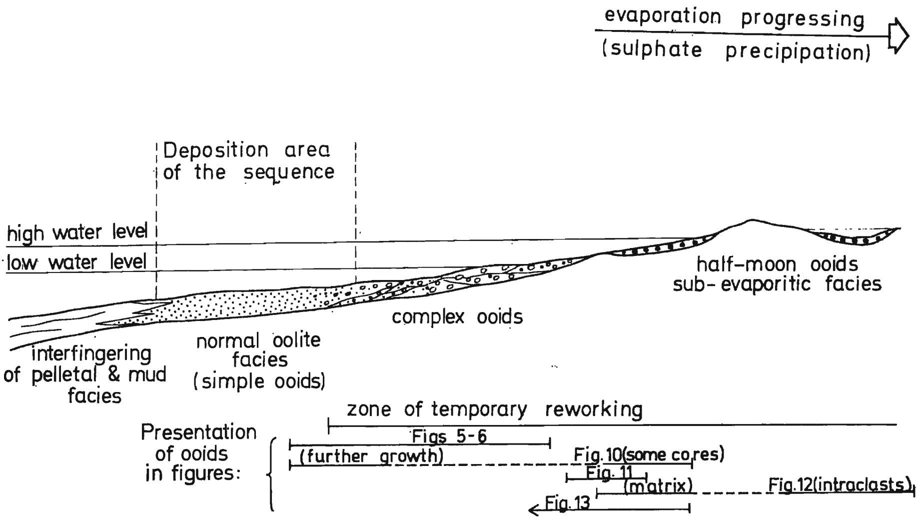


Fig. 20. Palaeoenvironmental model of a temporary lateral succession of the Dunøyane facies, and conjectured zones of origin of some ooids (before their sweeping into the final deposition area)

Note the conclusive features in the presented figures, as follows:

Figs 5—6: few intraclasts and grapestones, some of them oolitically coated, brought into normal oolite facies;

Fig. 10: sulphate-bearing cores brought into normal oolite facies for further growth;

Fig. 11: ooids with envelopes interwoven with sulphates, formed at the external margins of sub-evaporitic facies;

Fig. 12: intraclasts with half-moon ooids, as well as the discrete half-moon ooids brought into carbonate facies for further growth;

Fig. 13: half-moon ooids brought into oolite facies for further growth of common envelopes (complex ooids).

zed by bluegreen algae acting in the formation of dome-shaped stromatolites or, along the more agitated shoreline conditions, of onkolites, while the oolites were very local and spatially limited. As appears from the references, the south Spitsbergen sequence of the Dunøyane Member remains the only one known from the Precambrian in which the littoral facies were dominated by true physico-chemical, oolitic precipitation.

In the other parts of Spitsbergen and in northern Norway (Finnmark), the oolitic facies is subordinate, and it is developed within the stromatolite-bearing units. An accurate appraisal of its significance cannot be judged as the earlier reports are very scant, and an interpretation of imperfect illustrations may be misleading.

Various spherical structures, some of which being evidently ooids and small pisoids, were reported by C. B. Wilson (1961) from very shallow-marine or intertidal facies of Ny Friesland in northern Spitsbergen (Backlundtoppen Formation, correlated with the Höferpyten Formation — *cf.* Birkenmajer, 1972, Table 2); other forms, poorly illustrated and called by this author "oolitoids", defy identification. On the other hand, true ooids are recognizable in pictures presented by Milstein (1967) who calls these structures "gelwackes" and classifies them as onkolites; the enclosed illustrations, however, display more or less recrystallized ooids, either simple or identical with Carozzi's (1964) complex ooids (simple reworked, or with two and more nuclei). In northern Norway, the oolitic facies occurs locally in some units of the Porsanger Dolomite Formation, interpreted as inter- to supratidal (White, 1969), or also partly subtidal (Tucker, 1975), and correlated with the Höferpynten Formation (*cf.* Spjeldnaes 1964, Fig. 3; Birkenmajer, 1972, Table 2). Similar reports are also presented from the Late Precambrian of other regions in northern Europe: Spjeldnaes (1967) recognized a facies with calcareous pisoids (some of them interpreted however as algal) from pebbles in the lowest part of the Biskopåsen (= = Biri) Conglomerate in the Lillehammer Subgroup of southern Norway, while Smit & Swett (1972) reported the same type of facies from central East Greenland, viz. from some stromatolite-bearing units of the Nökkefossen Formation (Eleonore Bay Group); both these formations are also broadly equivalent to the Höferpynten Formation (*cf.* C. B. Wilson, 1961 pp. 89—90; Spjeldnaes, 1964, Fig. 3; Birkenmajer, 1972, Table 2). The presented review of the up-to-date records shows that Precambrian oolitic sedimentation attains greater role only in the Dunøyane Member of south Spitsbergen.

The suggested environmental model for the Dunøyane oolites may be regarded in its general composition as matching the present-day Bahamian facies. Previous records of such facies (*cf.* Illing, 1954; Newell & al., 1959, 1960; Purdy, 1963; Ball, 1967) from ancient times are known only from the Phanerozoic, e.g., from the Ordovician of Argentina (Serpagli, 1973); Ordovician (Black River Group), Devonian (Palliser Formation) and Mississippian (Rundle Group) of Canada (Beales, 1958); Upper Triassic (Rhaetian) of the Tatra Mts, Carpathians (Radwański, 1968); Upper Oxfordian (Osmington Oolite Series) of England (R. C. L. Wilson, 1968); this same age (Reynolds Oolite) of Arkansas (Akin & Graves, 1969); and Oxfordian/Kimmeridgian boundary of the Holy Cross Mts in Poland (Kutek & Radwański, 1965; Kutek, 1969). The differences consist in an absence of skeletal organisms and their rock-forming communities, greater supplies of deposits from nearby lagoonal facies, and mainly in the enormous size of the ooids in the Dunøyane Member. The latter feature could have resulted from a greater carbonate content in the sea water and its much greater and profuse precipitation, primary dolomitic during the Precambrian time. Taking into account the discussed differences, it is considered that the Dunøyane oolites demonstrate a Precambrian counterpart of the Phanerozoic Bahamian-type oolites.

PALAEOGEOGRAPHIC SIGNIFICANCE OF THE DUNØYANE OOLITES

The sedimentary environment of the whole Dunøyane Member has previously been discussed by Birkenmajer (1972), based on field observations. The shallow-marine sedimentation is evidenced by the occurrence of small-scale trough-cross-strafication in some oolitic/pisolitic dolostones, of dolostone-flake conglomerates (sedimentary breccias), and stromatolitic structures. The character of the cross-bedding points to the presence of possibly intertidal flats where the scouring of the bottom by ebb currents produced small rills, and ripple drift formed during flood tide. The origin of dolostone-flake conglomerates can be explained by reworking of laminated, already partly lithified carbonate which quickly desiccated and cracked when exposed to relatively warm air at low tide, then was redeposited by tidal currents. The varying rate of carbonate deposition controlled the mode of growth and shapes of algal stromatolites which grew in slightly deeper waters, possibly just below the low-tide mark. Fragmentation of algal structures also occurred, and stromatolite clasts are rather frequent.

Variation in thickness and lithology of the Dunøyane Member with coarsest oolitic and pisolitic dolostones present at Fannytoppen and Höferpynten-Wurmbrandegga which are the sites of minimum thickness of the Member (Fig. 3), suggests that the coastline was situated east of the area studied.

The petrographical and sedimentological analysis of the Dunøyane oolites postulates the same facies pattern. A comment will however be offered on some particular features. The Dunøyane oolites are poorly bedded, and the layers attain up to 3 m in thickness, but they are well stratified. It may be therefore assumed that the same facies persisted through a long span of time, and no greater erosional processes, recognizable in the adjacent areas, occurred in the deposition area (*cf.* Fig. 20). The graded bedding noted in the field (Birkenmajer, 1972) may be attributed to local strirring up of the deposits by stormy agitation, and their fractional settling thereafter (*cf.* Kuenen & Menard, 1952); such storm-orginated layers may be referred to as the tempestites *sensu* Ager (1974).

The carbonate column of the Höferpynten Dolomite Formation (Fig. 3) seems to correspond to a continuous change of facies from moderately neritic with the limestones at the base (Fannytoppen Member), through massive dolostones with chert nodules in the lower part (Andvika Member), and with shallower dolostones with stromatolitic structures at the top (Wurmbrandegga Member). The stromatolites may really be regarded as formed here in the subtidal environment, the same as that recently postulated for some Precambrian stromatolites (Hofmann, 1973; Monty, 1973; Truswell & Eriksson, 1973, 1975; Tucker, 1975). If so, the oolitic facies of the overyling Dunøyane Member (*cf.* Fig. 3) indicates the shallowest subtidal and progressively, inter- and supratidal environments in which the evaporation gradually progressed and hypersaline conditions appeared. They resulted in the first place in primary dolostones (*cf.* Friedman & Sanders, 1967) and, during their further advance within restricted lagoons, in sulphates. The best pronounced evaporite facies, characterized by strongly half-mooned ooids, is not however preserved *in situ;* it has completely been reworked during marine oscillations and is recognizable only in the intraclasts deposited in a permanently submerged area (*cf.* Figs 12—13 and 20).

The main differences between the Dunøyane oolites and the present-day Bahamian oolitic facies may be interpreted as resulting from a different structure of Precambrian epeiric seas and their shorezones, influencing different chemistry and precipitation conditions in the shallow marine, near-to-shore facies. An inset of oolitic facies within supposedly extensive, carbonate tidal flats and subtidal zones of the Late Precambrian stratigraphic column of south Spitsbergen may be compared to that recognized in some Mesozoic sequences (*cf.* Usdowski, 1962; Simone, 1971; Castellarin & Sartori, 1973 a), and the oolitic structures of which, as discussed previously, bear close resemblances to those of the Dunøyane Member.

The presence of stromatolites at the top of the Dunøyane Member is evidence of a gradual deepening of the sedimentary area that followed maximum epeiric upheaval recorded by the oolite/pisolite facies. A further deepening is also visible in non-carbonate facies that subsequently developed (Gåshamna Phyllite Formation — *cf.* Fig. 3).

A comparison of sedimentary environments of the Dunøyane Member and of contemporaneous carbonate sequences of other regions in Svalbard (Spitsbergen and Nordaustlandet) and of northern Norway (Birkenmajer, 1972, 1975; see also Kulling, 1934; C. B. Wilson, 1961; Spjeldnaes, 1964; Winsnes, 1965; Flood & al., 1969; Harland, 1969; White, 1969; Tucker, 1975), as well as of southern Norway and East Greenland (Spjeldnaes, 1964, 1967; Haller, 1971; Smit & Swett, 1972) shows that in the Late Precambrian similar environmental conditions prevailed over extensive areas in the epeiric seas of the northern hemisphere.

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