38. MESOZOIC CALCITURBDITES IN DEEP SEA DRILLING PROJECT HOLE 416A RECOGNITION OF A DROWNED CARBONATE PLATFORM

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ABSTRACT

The Tithonian-Hauterivian section of Site 416 contains about 10 per cent limestone and marlstone in well-defined beds between shale and sandstone. Depositional structures, grain composition, and diagenetic fabrics of the limestone beds as well as their association with graded sandstone and brown shale suggest their deposition by turbidity currents below calcite-compensation depth. Carbonate material was initially shed from a shallow-water platform with ooid shoals and peloidal sands. Soft clasts of deep-water carbonate muds were ripped up during downslope sediment transport. Shallow-water mud of metastable aragonite and magnesium calcite was deposited on top of the graded sands, forming the fine tail of the turbidite; it has been subsequently altered to tight micritic limestone that is harder and less porous than coccolith sediment under similar overburden. Abnormally high strontium contents are attributed to high aragonite contents in the original sediment. This suggests that the limestone beds behaved virtually as closed systems during diagenesis.

Drowning of the platform during the Valanginian is inferred from the disappearance of the micritic limestones, the increase in abundance of phosphorite, of ooids with quartz nuclei, and of the quartz content in the calciturbidites. In Hauterivian time, the carbonate supply disappeared completely. The last limestone layers are lithoclastic breccias which indicate that erosion has cut into well-lithified and dolomitized deeper parts of the platform rather than sweeping off loose sediment from its top.

Of all Tithonian sections recovered from the Atlantic, only the one at Site 416 contains abundant shallow-water debris and was deposited below the calcite-compensation depth. It is probably located on mid-Jurassic or older crust that had already deeply subsided and was close to the continental slope by Tithonian time. The other sites, located on younger crust, stood high on the flank of the mid-ocean ridge, and were consequently above the compensation level and beyond reach of shallow-water detritus shed down the continental slope.

INTRODUCTION

DSDP Site 416 is located on the outer continental rise off Morocco (Figure 1). During the Tithonian-Hauterivian interval considered here, the site remained below the calcite-compensation depth. The slow sedimentation of brown clay was frequently punctuated by influx of turbidity currents, depositing over 700 meters of graded beds with thin interbeds of ever-present shale. Carbonate material, in individual beds and (or) mixed with terrigenous sediment, accounts for about 10 to 25 per cent of the total turbidite contribution.

This material deserves special attention because: (1) it indicates a shallow-water carbonate source area now buried under thick continental-slope deposits; (2) the shale-sandstone-limestone sequence of Site 416 differs markedly from the cherty limestones recovered in the western Atlantic from the Tithonian-Hauterivian interval (Ewing, Worzel, et al., 1969; Hollister, Ewing, et al., 1972; Benson, Sheridan, et al., 1978); Site 416 adds a new facet to the spectrum of Jurassic ocean-floor deposits in the Atlantic; (3) the limestones close a gap in the spectrum of diagenetic case histories because they consist largely of shallow-water materials that, unlike many shallow-water limestones, were never exposed to fresh water; (4) the carbonate material at Site 416 forms hard, very tight limestones that profoundly influence the acoustic properties of the formation; our understanding of the origin and distribution of these limestones will improve our understanding of the nature and lateral variation of certain seismic reflectors in the deeply buried sequences of the Atlantic margins.

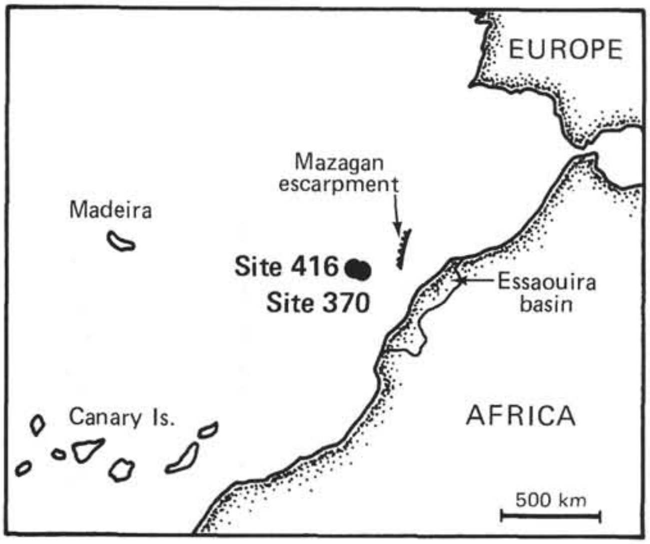


Figure 1. *Location of DSDP Sites 416 and 370. Jurassic Cretaceous shallow-water limestones which crop out along the Mazagan escarpment and onshore in the Essaouira basin are believed to be the source of the calciturbidites.*

The emphasis of this report is on standard petrographic methods with stained thin sections and peels, supplemented by work with the scanning electron microscope, X-ray diffraction, and atomic-absorption spectrometry. All thin sections were stained with a combination of Alizarin Red S and potassium ferricyanide to identify ferroan and non-ferroan calcite as well as ferroan dolomite (Evamy, 1963). A stain of 4 per cent ammonium molybdate in 25 per cent nitric acid was used to identify phosphorite.

TYPES OF TURBIDITES IN THE TITHONIAN TO HAUTERIVIAN SECTION

On the basis of the core descriptions prepared onboard ship and subsequent onshore examination, the turbidites are subdivided into three types (Table 1): terrigenous turbidites (mainly brown), composed of siliciclastic detritus and clay; carbonate turbidites (mainly gray) consisting of well-cemented carbonate sand-silt and mud (Figures 2,3,4A); and mixed turbidites composed of quartz-rich calcarenite or carbonate-rich quartz sand with a fine tail of clay and coccolith mud (Figure 4B; see also Price, this volume).

Only the terrigenous turbidites occur throughout the Tithonian-Hauterivian interval (939 to 1624 m). Pure carbonate beds are restricted to the lower part of the section (1430 tp 1624 m); mixed turbidites are common and well developed between 1435 and 1280 meters and rare between 1280 and 1190 meters (see also section on Downhole Variation of the Calciturbidites).

Depositional structures are more or less the same in all three types. The beds are size graded, have sharp bases and gradational tops, and commonly show Bouma's (1962) sequence of depositional structures; most cycles start with interval B or C.

TABLE 1
Types of Turbidites

Lithology	Color	Depositional Structures	Bouma Interval
Brown (Terrigenous), Cores 7-57			
4) Red zeolitic mudstone	Brownish red 2.5-5YR 3-4/3-4	Burrowed	D, E
3) Brown mudstone	Brown 5YR 3-5/2-4	Homogenous or faintly laminated	D
2) Siltstone	Brown to grayish green 5YR4/1 to 5/2	Parallel to cross laminated	C, D
1) Sandstone	As above	Massive; parallel or cross laminated	A, B, C,
Green (Mixed), Cores 7, 10-36			
3) Calcareous mudstone to marlstone carbonate mainly nannofossils	Pale green to grayish green, 5G 4-6/1-2 10G 6/2 10GY 5/2	Massive or faintly laminated, burrowed	D
2) Quartz-siltstone some carbonate detritus	Gray to greenish gray N6-5GY 6/1	Parallel or cross laminated	C, D
1) Calcarenite to lithic sandstone rich in phosphorite, some glauconite	Similar to (2)	Massive, parallel or cross laminated	A, B, C,
Gray (Carbonate), Cores 36-57			
3) Micritic limestone or marlstone carbonate mainly neomorphic (?) micrite	Gray or pinkish or greenish gray 10YR7/2 5Y6/1-2, 5Y5/3 5YR7/2 5G4-5/1-2	Burrowed faintly layered	D
2) Calcisiltite, quartzose, tightly cemented grainstone or packstone	Colors similar to (3)	Parallel or cross laminated	C, D
1) Calcarenite, tightly cemented grainstone or packstone with matrix of brown clay	Colors mostly gray, vary with grain composition	Massive, parallel or cross laminated	A, B, C,

In any particular core, carbonate and mixed turbidites differ from the terrigenous beds, not only by composition but also by their thicker, coarser beds and more complete Bouma cycles, suggesting deposition closer to the source. This difference in proximity, together with the lack of transitions between terrigenous and carbonate cycles below 1280 meters in the section, suggest different sources for carbonate and terrigenous turbidites.

CARBONATE MATERIAL

Because there are many graded beds almost entirely composed of carbonate material, I conclude that carbonate detritus of all sizes from granule to clay size was carried in from the source area. Because identification of grain kind depends on size, I will deal separately with the coarse fraction of granule to medium sand, with the fine sand to silt, and with the clay-grade material.

Coarse Fraction (granule to medium sand)

Ooids of two types, translucent radial calcite (''Bahamian'' ooid) or concentrically layered micrite (''pelag-

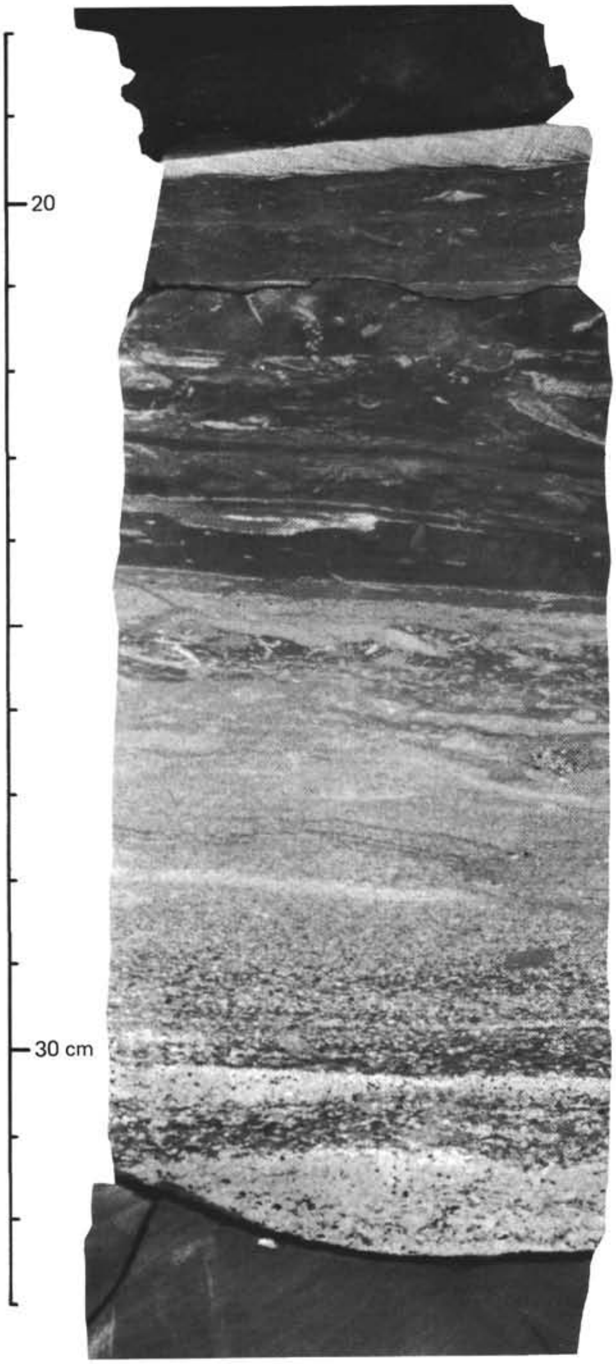


Figure 2. *Graded sand layer, starting with Bouma interval B. Note intensive burrowing in marlstone (interval D); nutrients and burrowers were probably swept in with the turbidity current. Cycle ends at base of cross-bedded sandstone at 19.5 cm. Sample 416A-37-3, 18–33 cm.*

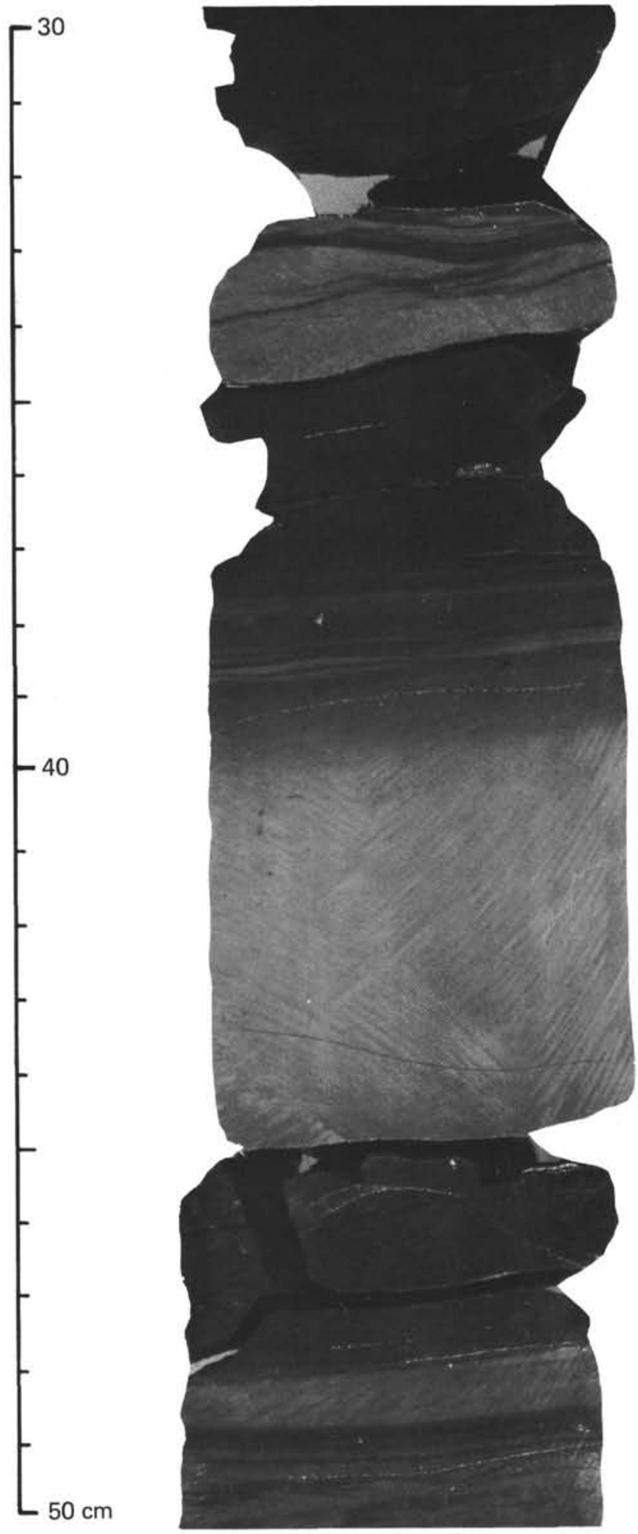


Figure 3. *Micritic limestone in brown shale with sandstone. A full-diameter core is recovered from the hard limestone, whereas the shale fractures and "caves." Note gradational top of limestone layer. Sample 416A-57-1, 30–50 cm.*

ic'' ooid, Jenkyns, 1972), are common to abundant. See Figure 5 and discussions below.

Peloids are circular, elongate, or irregular-shaped grains of structureless micrite or microspar. In our sample, we believe the peloids are in part hardened fecal pel-

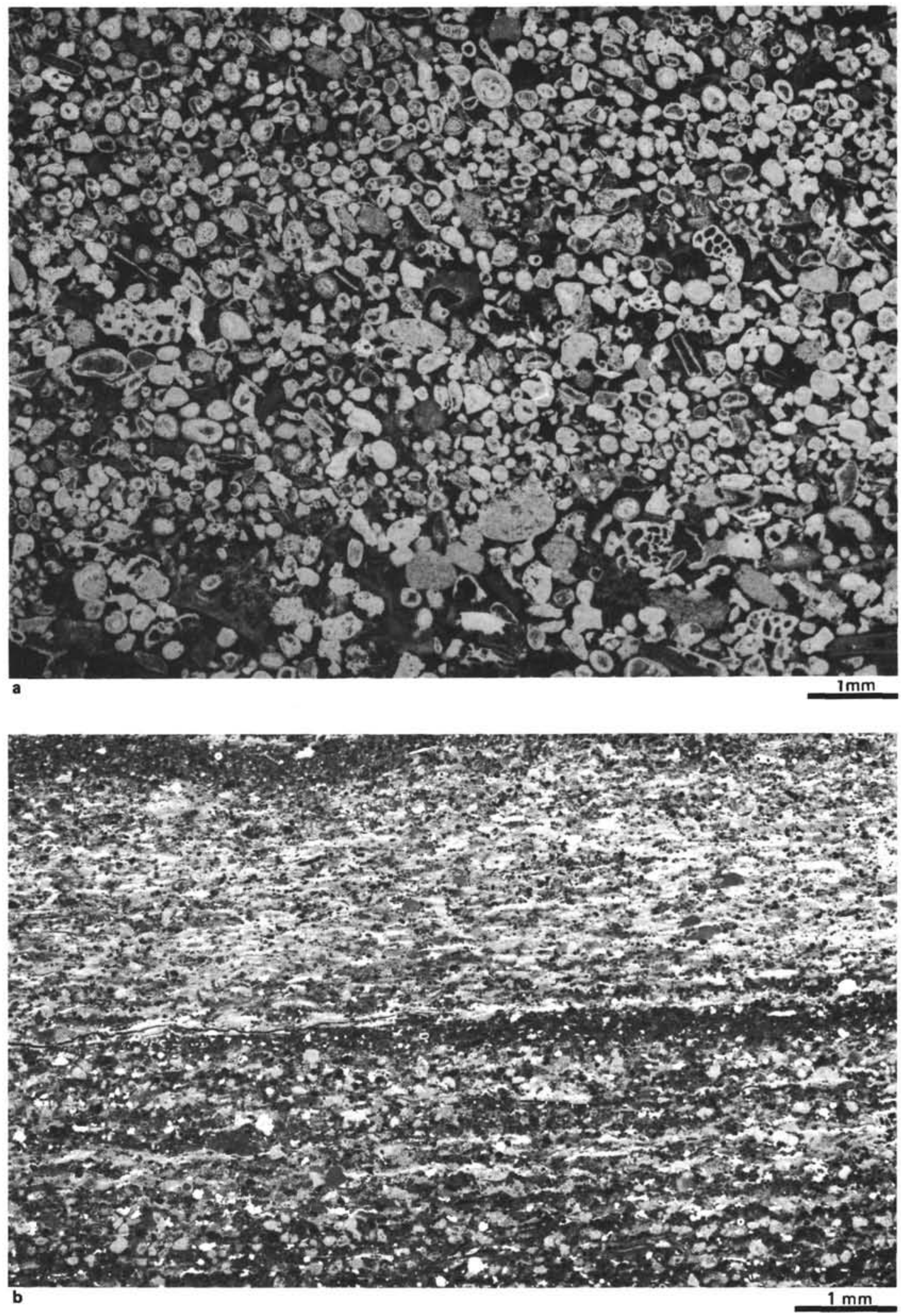


Figure 4. (a) *Grainstone from base of calciturbidite (gray type) with ooids, peloids, skeletal grains, and lithoclasts of pelagic mudstone. Sample 416A-50-2, 23–26 cm. Thin-section, negative print.* (b) *Graded interval of quartzose carbonate–phosphorite turbidite (green type); lower half shows grainstone of carbonate grains (light), quartz (dark), phosphorite (medium gray, arrows); upper half consists of flat intraclasts of brown clay (light), quartz, and phosphorite. Sample 416A-12-2, 1–6 cm; thin-section negative print.*

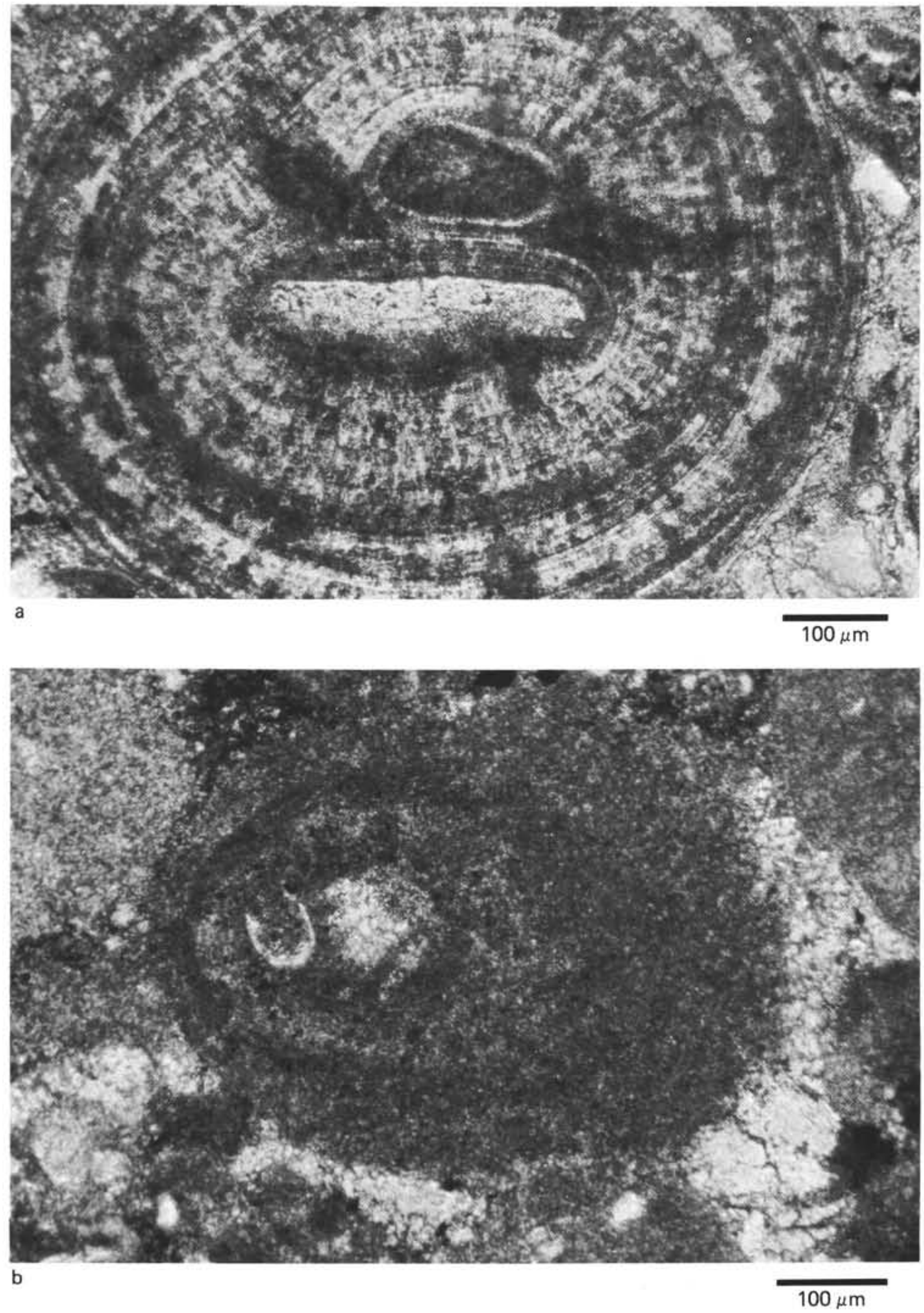


Figure 5. *Two types of ooids: (a) "Bahamian" ooid of radially oriented calcite with several micrite rims, thought to have formed as aragonite or magnesian calcite in very shallow water. Section 416A-27-4. (b) "Pelagic" micrite ooid with tintinnid, believed to have formed in slightly deeper water by sediment trapping on algal films. Sample 416A-37-2, 43–44 cm.*

lets, in part micritized ooids, and in part worm micritic lithoclasts. Very abundant. Figures 6, 11.

Grapestones are lumps of peloids, ooids, and (or) skeletal grains, loosely bound by micrite cement. Rare. Figure 7.

Lithoclasts are divided into two categories: (1) rarely occurring clasts of ooidal, peloidal, and (or) skeletal grainstones of shallow-water origin as hard, subangular to subrounded clasts; (2) very abundant clasts of mudstone, or skeletal wackestone with tintinnids, sponge

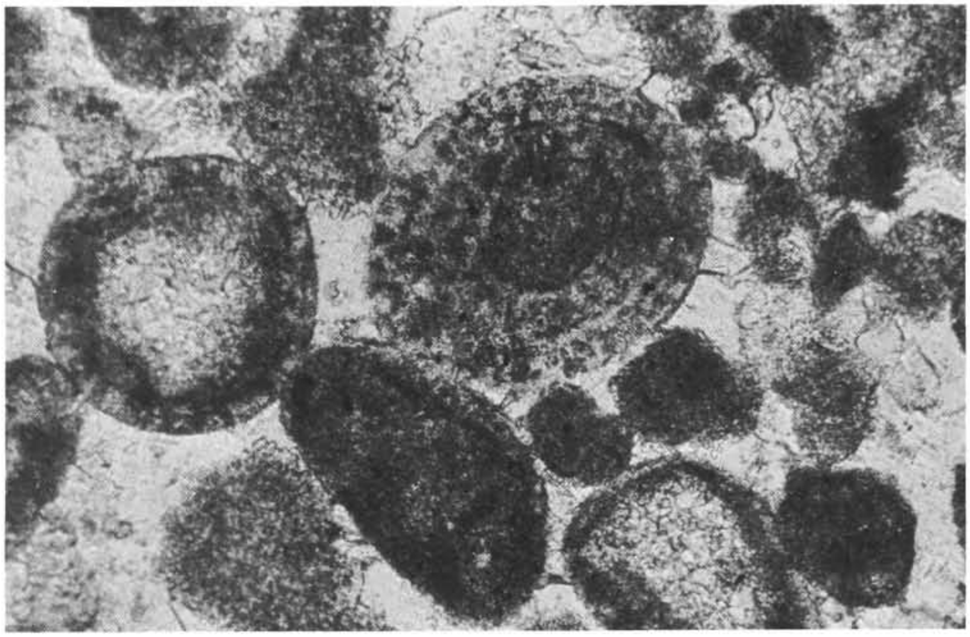


Figure 6. *Ooids and peloids cemented by blocky calcite. Some compaction indicated by fitted grain contacts. Sample 416A-50-2, 23–26 cm.*

spicules, and other pelagic biota. Most of these fine-grained clasts are severely deformed by compaction and were probably soft when being deposited. Grainstone clasts are rare, but soft clasts of mudstone or wackestone are very abundant. Figure 8.

Skeletal grains are about as abundant as ooids. They include dasycladacean green algae (cf. *Thaumatoporella*); questionable *Clypeina jurassica*, a pelagic alga; calcispheres; foraminifers with rotaliid shell structure (e.g., *Lenticulina*) and with miliolid structure; fragments of recrystallized, originally aragonitic hexacorals; bivalves with thick layers of neomorphic calcite (after aragonite); thick prismatic layers of *Inoceramus* (rare); ostreid shells with foliated structure; very thin-shelled ("pelagic") bivalves of the *Posidonia* type; aptychi; echinid spines; crinoid ossicles; including the bizarre shapes of the pelagic crinoid *Saccocoma*; and tintinnids.

Non-carbonate components make up between 1 and 50 per cent of the layers of carbonate sand (beds with over 50% non-carbonate material are classified as terrigenous sandstones). The grains are mainly quartz, intraclasts of quartz-bearing brown clay, phosphorite, and glauconite.

The clasts of brown clay are intensely deformed and act as a secondary matrix between the harder components. Clayey tintinnid limestones and clasts of tintinnid-bearing clay are common and were probably ripped up in the transition zone between carbonate sediments and brown clay.

Fine Sand and Silt

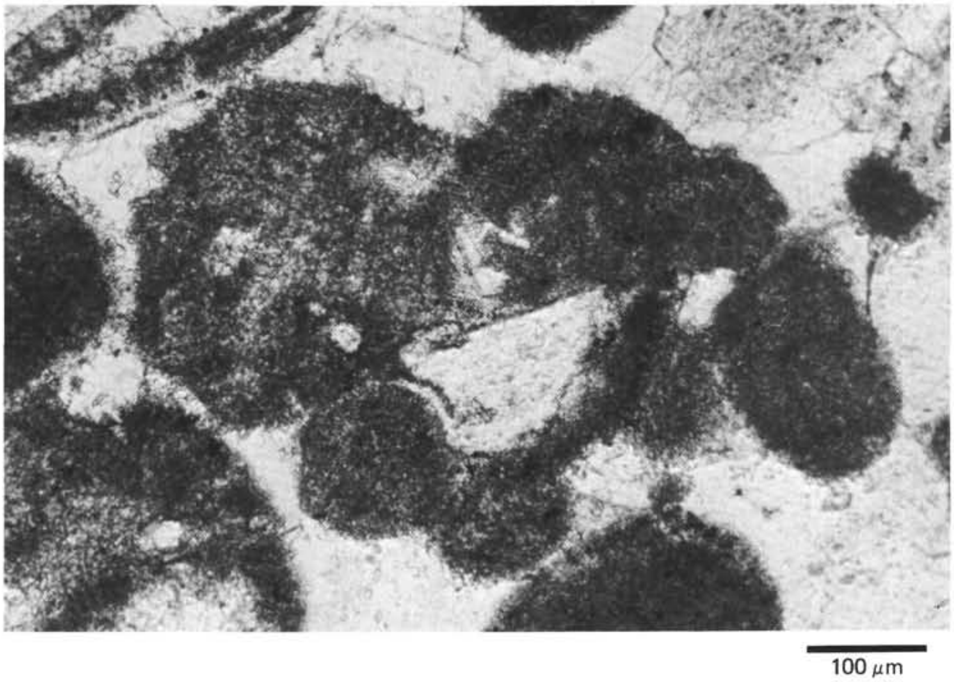
Only part of the grains of this fraction can be identified, because of their small size and because the grains are often deposited together with mud to form packstones in which the individual grains are difficult to recognize. Most of the grains appear to be micritic, probably derived from breakdown of peloids. Besides micrite, I noticed a fine shell hash, possibly of pelagic bivalves or tintinnids.

Mud

In the "green" turbidites of mixed terrigenous and carbonate material, the graded sand-silt interval is overlain by a marlstone. Its carbonate fraction consists mainly of nannoplankton, with a small portion of (neomorphic?) micrite.

In the gray calciturbidites the graded layer is capped by tight limestone with sparse fragments of tiny shells, probably of tintinnids. The carbonate is micrite with only traces of recognizable nannoplankton (Figure 9).

In both cases the carbonate layers are not perennial sediment, but are part of the fine tail of the turbidite. This is indicated by their position above the graded sand-silt interval of a turbidite and their passing upward into brown or greenish claystone. Between the tight limestone and the overlying claystone there are occasionally a few millimeters of gray marlstone, rich in arenaceous foraminifers. I interpret this band as an insoluble residue, formed when the top of the calciturbidite was dissolved by undersaturated sea water. There can be little doubt that both marlstone and tight limestone were primarily composed of clay-grade carbonate particles, because both rock types contain only negligible amounts of coarser debris and both occur in the turbidite sequence exactly where fine, muddy sediment is to be expected. I believe, however, that the carbonate fraction was of different composition. While the green marlstone is mainly composed of coccolith debris, the



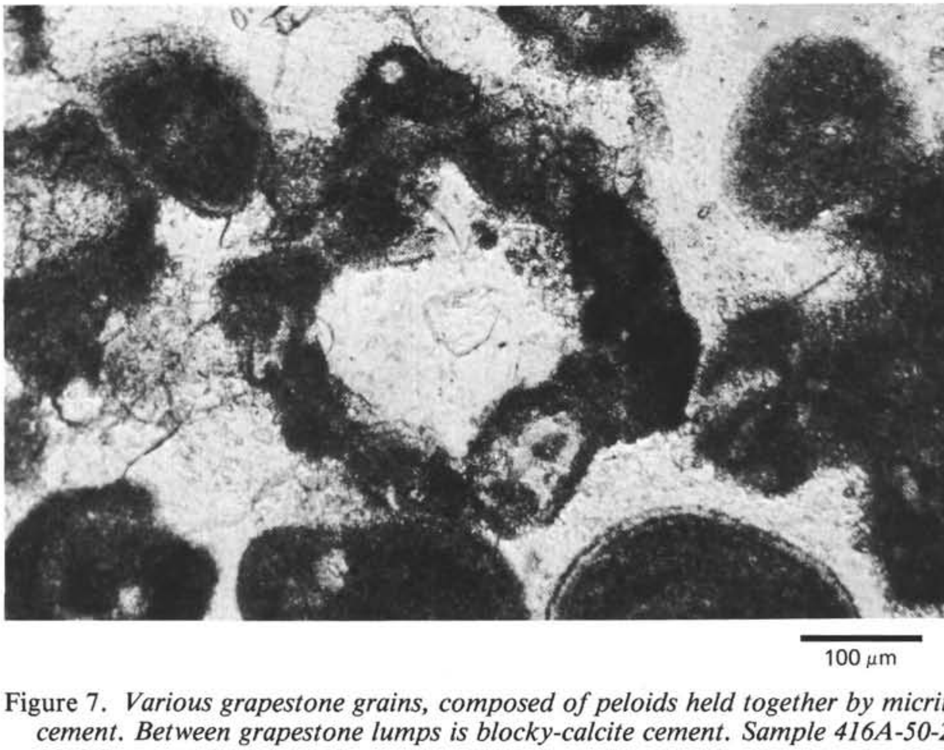


Figure 7. *Various grapestone grains, composed of peloids held together by micrite cement. Between grapestone lumps is blocky-calcite cement. Sample 416A-50-2, 23–26 cm.*

limestone was probably predominantly aragonitic mud of shallow-water origin. This is discussed further under Origin of the Micritic Limestone.

Shallow-Water and Deep-Water Grain Assemblages

Most of the carbonate grains mentioned above are fairly characteristic of a specific depositional environment.

Contemporary *ooids* form in warm, shallow water where they are continuously moved by waves and (or) tidal currents (Illing, 1954; Newell et al., 1960; Loreau and Purser, 1973; Bathurst, 1971; Lees and Buller, 1972). Jenkyns (1972) pointed out that certain types of micritic ooids with pelagic fossils may actually form below several tens of meters of water. Ooids of this kind do occur in our samples and may indicate a transition to deep-water facies (see below).

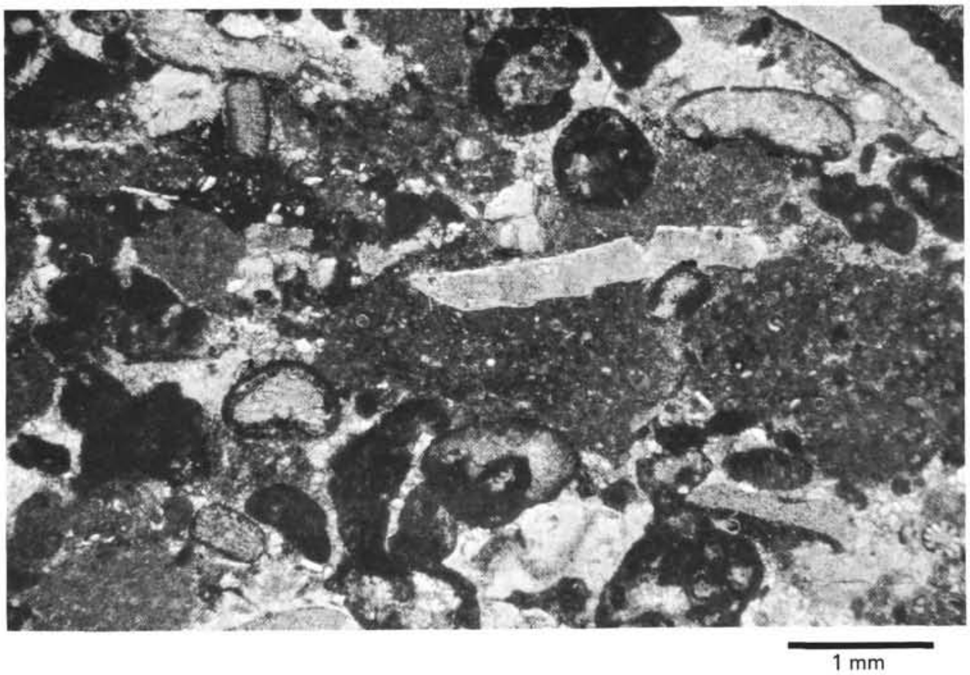


Figure 8. *Various shallow-water grains surrounding intraclast of deep-water limestone with tintinnids and aptychus valve (fractured during compaction). Large size of intraclast and its deformation by shallow-water grains suggest that clast was deposited as soft sediment of low density. Thin-section, Sample 416A-45-1, 24–28 cm.*

Peloids are either fecal pellets or micritized ooids and micritized, rounded skeletal grains. Note that in the Hole 416A material the peloids deposited by the turbidity currents were as hard as ooids and much harder than the clasts of tintinnid limestone. They must have been either lithified fecal pellets or initially hard grains such as ooids and skeletal fragments. This in turn argues for their shallow-water origin; contemporary hardened fecal pellets have been found to form only in shallow-water environments such as the Bahama Bank, where micrite cement or oolitic coatings harden the soft mud pellets (Illing, 1954; Bathurst, 1971, p. 299; Ginsburg and James, 1974, p. 137). Micritization is usually caused by photosynthetic algae and therefore is also most common in shallow water, although in exceptional cases it can be caused by fungi in deep water (Swinchatt, 1969; Bathurst, 1971, p. 389). The hard peloids are thus in all likelihood also derived from a shallow-water source, much the same as were the ooids.

Grapestones indicate rapid cementation or organic binding and are characteristic of shallow and protected carbonate environments in modern seas (Illing, 1954; Bathurst, 1971, p. 316-319).

Of the biota, dasycladacean *green algae* are diagnostic of the shallow euphotic zone. Miliolid foraminifers are also most common in shallow water of shelves and platforms and occur in great abundance even in the restricted environments of the platform interior (Wantland, 1975; Rose and Lidz, 1977).

We can therefore recognize a *shallow-water assemblage* (euphotic zone) among the detrital carbonate material which includes ooids; peloids; grapestones; clasts of ooidal, peloidal, or skeletal grainstone; miliolid foraminifers; thick-shelled bivalves; and green algae.

Ooids and peloids were probably derived from the agitated margins of a carbonate platform; grapestone and miliolids represent the protected and more restricted facies of the bank interior. Lime mud for the fine tail of the turbidites might have been produced in several different environments of the platform, as is suggested by comparable modern platforms (Stockman et al., 1967; Neumann and Land, 1975).

The remaining components are clearly part of a *deep-water assemblage*. This category includes intraclasts of skeletal wackestones and mudstones with pelagic biota such as tintinnids, *Saccocoma*, and calcisphaerulids. Sponge spicules are also common in some of the clasts.

The clasts are larger than the associated shallow-water grains and are severely deformed by early compaction (Figure 8). I interpret them as pieces of soft sediment ripped up by the turbidity currents. The clasts seem to represent a spectrum of carbonate facies ranging from those deposited at wave base to those deposited at the compensation depth. The shallow end of the spectrum is marked by the occurrence of micritic ooids with tintinnid fragments and by superficially coated clasts of tintinnid wackestones ("pelagic ooids" of Jenkyns, 1972). The other, deep-water end of the spectrum

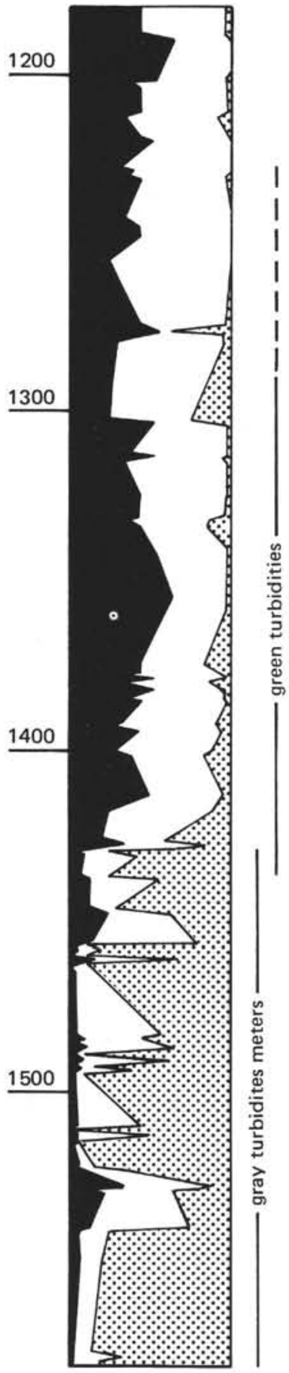


Figure 9. *Content of nannofossils (black) and micritic calcite of presumably neomorphic origin (dotted); non-carbonate fraction left blank. Estimates made aboard ship from smear slides. Scarcity of nannofossils in micritic limestones is one of several arguments for a shallow-water-carbonate source of the micrite in the gray turbidites. Carbonate in fine tails of green turbidites, on the other hand, is mainly coccolith material.*

is represented by the transition from clasts of tintinnid limestone to clasts of brown clays as mentioned above.

DIAGENESIS

Alteration of Grains

The preservation of *skeletal grains* clearly reflects their original mineralogy. Aragonitic shells have been replaced by neomorphic calcite, the original layering often indicated by brown (organic-carbon-rich) bands or by layered inclusions. Green algae, originally probably randomly oriented aragonite needles like the recent *Halimeda*, have been altered to calcite micrite, as were miliolid foraminifers (originally magnesian calcite). Echinoderms (originally magnesian calcite) show the usual pattern of occlusion of the intra-skeletal pore space, followed by syntaxial overgrowth of low-magnesium calcite. Particles composed originally of low-magnesium calcite, such as ostreid shells, do not appear to be altered at all. Others, such as aptychi, are heavily overgrown by syntaxial calcite, and their original porous texture is still visible through organic inclusions. Grains of all mineralogies frequently have micrite rims.

Ooids are preserved in two ways. Some consist of concentrically layered and radially oriented clear calcite, with brown luster and with many inclusions of 0.1 μm. Other ooids are concentrically layered calcite micrite. They lack the brown stain and any radial texture. By analogy with present-day Bahamian ooids, I interpret the clear, radially oriented grains as recrystallized ooids from an active shoal. The micritic, concentrically layered grains resemble the "pelagic ooids" described by Jenkyns (1972) from the Tethyan Jurassic. They are thought to have formed similarly to onkoids, by particle trapping on algal coatings, rather than micritization of clear ooids. The occasional occurrence of calpionellids in these micritic ooids (Figure 5) supports Jenkyns's view of a slightly deeper, more open marine environment of formation.

The original composition of the Hole 416A ooids could have been either aragonite or magnesian calcite. There is no evidence that the ooids have been leached selectively from the rock. Sorby (1879) and others have used the absence of selective leaching as an argument against aragonitic composition of many ancient ooids. In our case, this observation is of little diagnostic value, because the ooids in the calciturbidites were almost certainly never exposed to fresh water and therefore had little opportunity to be selectively leached. In the absence of fresh water, however, aragonitic ooids recrystallize to radial calcite ooids and preserve a large part of their primary texture. A good example of this sort are the Tertiary ooids Hesse (1973) described from a drowned Pacific seamount with radial and concentric texture and still some aragonite preserved in the cortex. The crystal inclusions in the ooids may eventually provide a clue to the original composition because calcite as a successor of magnesian calcite often betrays its origin by dolomite inclusions (Lohmann and Meyers, 1977). In the case of the Hole 416 ooids, however, the inclusions are only about 0.1 μm and could not be identified.

The peloids were deposited as hard grains of micrite on the sea floor, as can be inferred from their behavior during compaction. Many of them have the shape and size of ooids and are probably completely micritized ooids. The larger and more elongate peloids were probably fecal pellets, hardened at the sea floor prior to transport in the turbidity current. In contrast to the micrite of the fine tail of the turbidites, the calcite micrite of the peloids is not ferroan. It must have stabilized to calcite before the onset of burial diagenesis and did not recrystallize during later stages of diagenesis.

Cementation

Most of the sand and silt layers of the turbidites are free of mud ("grainstones" in the classification of Dunham, 1962). The original pores are filled by a simple sequence of cement (now all calcite): (1) a thin druse of calcite microspar occasionally forms the first layer around the grains and predates solution-compaction; this is followed by (2) slightly ferroan, blocky calcite that postdates or overlaps with pressure solution and completely fills the residual pore space (Figures 6,7,10,11). Around the echinoderm grains, aptychi, and prism layers of *Inoceramus*, syntaxial-overgrowth cement competes with the blocky-calcite. Most of this overgrowth cement is not ferroan; it must therefore predate the blocky cement.

The micrite cement in the grapestone lumps and relics of a fibrous druse in certain lithoclasts are thought to have been carried in from shallow-water environments and are not part of the normal sequence of cements in the turbidites.

Metasomatism

Dolomite was identified by staining and X-ray diffraction. It occurs in scattered rhombs in the grainstones and packstones, but not in the micritic limestones. (Authigenic dolomite is common in the brown clay surrounding the calciturbidites.) Lithoclasts of a sucrosic dolomite rock were found in the Hauterivian breccias of Core 7. I believe they are reworked older parts of the same shallow-water carbonate platform that provided the sand-size debris for the lower part of the section.

Phosphorite was found in the classic layers throughout the cores. Single grains were identified microscopically, concentrations of phosphorite of over 0.5 per cent (volume %) by a yellow stain with ammonium molybdate. Most of the grains are unidentifiable, angular to subrounded fragments which form part of the clastic portion of the turbidite. Some grains have the shape and size of shell fragments; others are round, with a faint concentric structure reminiscent of ooids. I assume that at least part of the phosphorite is a replacement of carbonate grains and was washed in from a drowned carbonate platform (see Downhole Variation of the Calciturbidites). Precipitation of phosphorite, coupled with metasomatic replacement of carbonate by phosphorite, is common in areas of slow deposition just below the photic zone (Ames, 1959; Gulbrandsen, 1969).

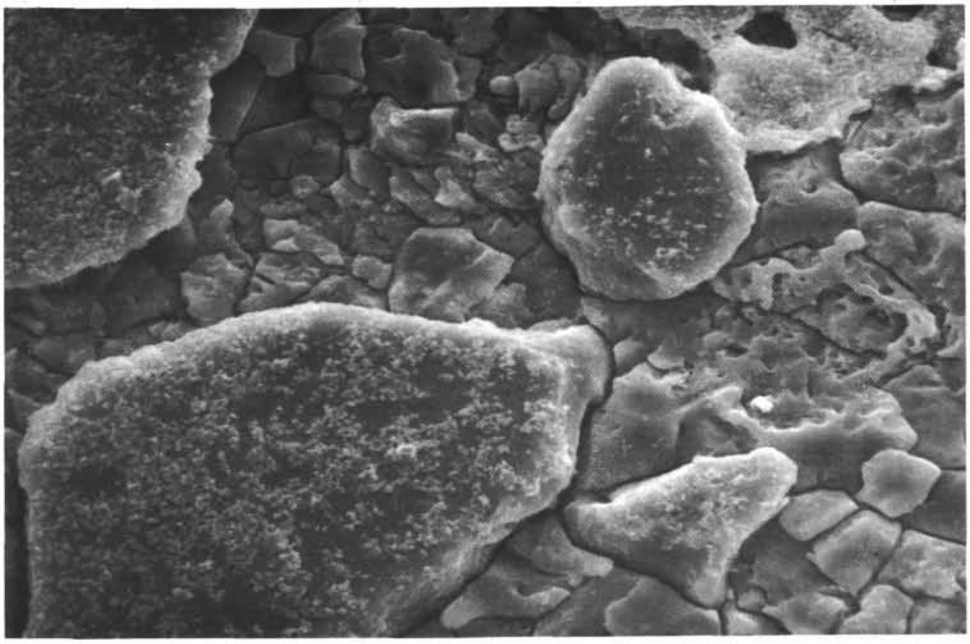


Figure 10. *Tightly interlocking mosaic of calcite cement between quartz grains and unidentified skeletal fragment (upper right). Coarse layer of calciturbidite, Sample 416A-57-1, 147–14 cm. SEM micrograph of polished and etched surface.*

Figure 11. *Interlocking mosaic of blocky cement around micritic peloid. Cement is ferroan calcite of burial diagenesis; peloid is non-ferroan calcite of presumably early diagenetic origin. Sample 416A-57-1, 147–149 cm. SEM micrograph of polished and etched surface.*

Lithification Of Mud

The fine tails of the calciturbidites are either marlstone composed of cocoliths and clay, or micritic limestone, probably largely recrystallized aragonite mud of shallow-water origin (see Origin of the Micritic Limestone). In both cases, the fine sediment has been lithified, but the diagenetic overprint is more severe in the limestone than in the marlstone. The limestone has a residual porosity of 1 to 7 per cent and consists almost exclusively of neomorphic(?) calcite micrite with less than 5 per cent recognizable cocolith material (Figure 9). The texture is an interlocking amoeboid mosaic, generally considered a final stage of micrite diagenesis (Fischer et al., 1967; Bathurst, 1971, p. 502; see Figure 12). Staining reveals an iron content in the micrite similar to that of the blocky calcite cement. The marlstone, on the other hand, has a porosity of 10 to 25 per cent and most of its carbonate fraction is still-recognizable coccolith material; only a small portion is neomorphic calcite micrite.

Pressure Solution

Three groups of pressure-solution features can be recognized in the calciturbidites:

1. *Microstylolites,* horse-tails (Roehl, 1967), and sutured grain contacts occur frequently throughout the Hauterivian-Tithonian interval.

2. Well-developed *stylolites* with amplitudes of 5 mm or more occur from Core 40 (1460 m) down into the micritic limestones. They are vertical, and when taken apart their surfaces show slickenside-like striations rather than the common pattern of interlocking cones. I conclude that these vertical stylolites were caused by strike-slip movements with lateral compression rather than by simple overburden compaction.

3. *Pervasive pressure solution* along crystal boundaries (Wanless, 1977; Scholle, 1977, p. 994) must have affected the micritic limestone as well as the grainstone in order to form the tight, interlocking mosaics of micrite and sparry cement shown in Figures 10, 11, and 12. The porosity of the micritic limestones decreases from about 12 per cent at 1450 meters to less than 5 per cent at 1600 meters. I attribute this loss of porosity to an increase in pervasive pressure solution.

Diagenetic Environments

Burial diagenesis is by far the most prominent diagenetic environment recognizable in the rocks. It caused not only the compaction and pressure solution in the grainstone and mudstone, but also provided over 90 per cent of the pore-fill in the form of blocky, slightly ferroan calcite cement that postdates compaction and at least incipient pressure solution. The similar iron content in blocky calcite cement and neomorphic micrite of the fine turbidite tails suggests a burial diagenetic origin for most of this micrite too. The micrite of peloids is non-ferroan and thought to be the result of shallow-water lithification. The iron content of the burial cements indicates reducing conditions within the turbidite layer in spite of the high oxidation state of the perennial sedi-

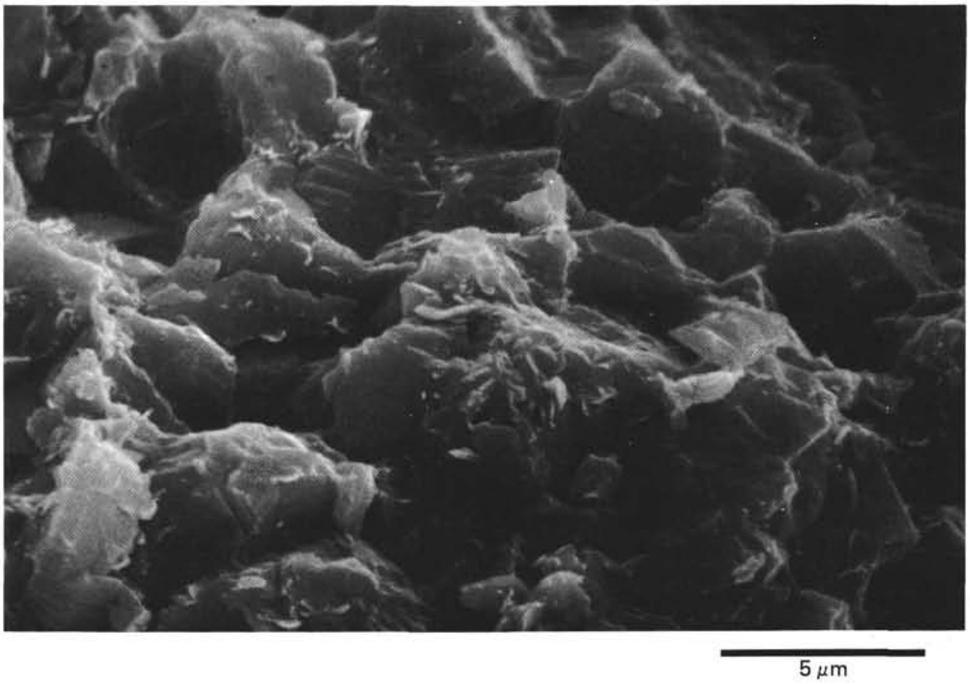


Figure 12. *Tightly interlocking mosaic of ferroan micrite from fine tail of shallow-water turbidite. Limestone is rich in strontium and magnesium and is thought to be a neomorphic alteration of aragonite and magnesian calcite. Sample 57,CC. SEM micrograph of broken surface.*

ment. The allochthonous material was probably buried with abundant organic matter swept in from the shallow source area.

Diagenesis at the deep-sea floor was of minor importance. It caused some dissolution at the tops of the calciturbidites, indicated by a drop in carbonate content. The thin druse of microspar that predates compaction probably also formed at least close to the sea floor.

Shallow-water diagenesis is represented by micrite rims on skeletal grains and ooids (characteristically lacking on deep-water aptychi and *Saccocoma* fragments), by micrite cement in grapestones and peloids, and by ghosts of fibrous cement in lithoclasts. With the exception of a few doubtful cases of selective leaching of aragonite, no indications of fresh-water diagenesis have been found.

ORIGIN OF THE MICRITIC LIMESTONE

Several independent arguments suggest that the micritic limestone was originally not coccolith sediment but was composed largely of shallow-water carbonate mud:

1. The total content of recognizable nannoplankton in the micritic limestones is less than 10 per cent, much lower than in the nannofossil marlstone that forms the fine tails of the calciturbidites in Cores 36 and 37 (Figure 9), notwithstanding the higher total carbonate content of the limestones.

2. The micritic limestone occurs always as the fine tails of calciturbidites of mainly shallow-water detritus. The abrupt decrease in shallow-water detritus and the change to phosphorite-rich relict sands above Core 37 also marks the disappearance of the micritic limestone. This correlation suggests that the micritic limestone was derived from the same shallow-water source as the carbonate sand, and was not, or only partly, cocolith mud picked up on the way downslope.

3. The limestone is considerably less porous and better lithified than pure coccolith ooze under comparable overburden and temperature conditions. Its porosity at 1500 meters is between 3 and 10 per cent, while the trend established by Scholle (1977, p. 992) and Schlanger and Douglas (1974) would predict values of 15 to 25 per cent (Figure 13).

4. Atomic-absorption analyses of the limestone (Table 2) show abnormally high contents of strontium. It seems impossible that this amount of strontium can be accommodated in the lattice of low-temperature calcite. Rather, it may occur as strontianite and (or) celestite formed during the transformation of strontium-rich aragonite to strontium-poor calcite. Neither strontium mineral could be identified by X-ray diffraction, however; only the presence of dolomite could be verified. Dolomite accounts for the high magnesium content of the limestone and may have formed as a by-product of the recrystallization of magnesian calcite to calcite.

All these observations have led me to believe that the micrite limestone—contrary to the nannofossil marlstone above Core 37—consisted largely of platform-derived mud with a minor portion of calcareous nannoplankton. If the recent is in any way a key to the past, then this mud must have consisted of aragonite and some magnesian calcite. In modern carbonate platforms, the mud consists of aragonite and magnesian

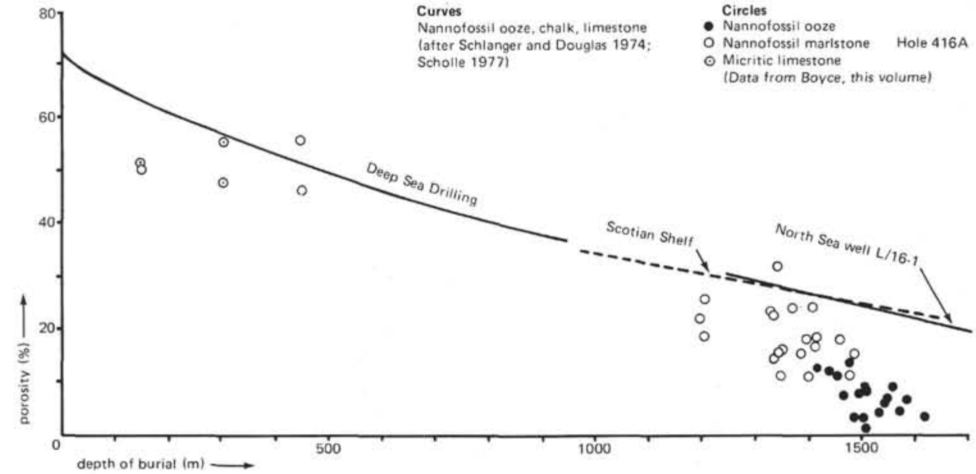


Figure 13. *Curves of porosity versus burial depth for coccolith sediments in DSDP holes and oil wells. Coccolith sediments of Hole 416A follow this trend down to about 1350 meters. Influx of shallow-water carbonate mud below this level is thought to have caused the remarkably low porosity.*

TABLE 2
Atomic Absorption Analyses of Some Limestone Samples

Sample (Interval in cm)	Insoluble Residue	MgO	CaO	Sr	FeO	MnO
37-1, 10	22.48	1.9	37.5	0.67	1.55	0.12
46-2, 18-21	10.1	0.79	47.5	0.93	0.78	0.12
47-1, 12-15	11.1	1.05	46.9	0.91	0.80	0.065

calcite in a ratio of 1:1 to 9:1 (Husseini and Matthews, 1972; Matthews, 1966; Taft and Harbaugh, 1964). Most, if not all, of this lime mud is of organic origin (Stockman et al., 1967; Neumann and Land, 1975). Green algae, the main mud producers in the recent, occurred in abundance throughout the Mesozoic (Wray, 1977, p. 153).

The question remains, why are the micritic limestones so tight and so highly altered compared to nannofossil ooze under similar overburden (Figure 13). The micritic limestone is now a tightly interlocking mosaic of calcite crystals with sutured boundaries. It seems that the inevitable merry-go-round of pressure solution and reprecipitation that controls burial diagenesis of limestone (Scholle, 1977; Logan and Semeniuk, 1976) has gone to completion earlier than in coccolith sediments. The original content of aragonite and magnesian calcite is the most likely cause of this diagenetic "prematurity." The path of diagenesis to this final stage of tight micrite may have followed either one of the following two alternatives:

1. The mud was deposited so rapidly that the metastable minerals were little affected by dissolution at the sea floor. During burial diagenesis, the turbidite sandwiched between clay was a closed system in which some aragonite may have survived until the onset of pressure solution. Because of the higher solubility of aragonite over calcite, these aragonite-bearing rocks were more susceptible to pressure solution and matured to tight limestone earlier than purely calcitic rocks.

2. The metastable minerals stabilized rapidly at or near the sea floor and were replaced by calcite of 4 to 5 mole per cent $MgCO_3$, which seems to be the least-soluble form of calcite in sea water (Plummer and McKenzie, 1974; Füchtbauer and Hardie, 1976; Schlager and James, 1978). With change in pore fluids during burial diagenesis, this magnesium-bearing calcite may have become less stable than pure calcite and may have been converted to pure calcite by pressure-solution reprecipitation. In this way, the last pore space may have been closed.

DOWNHOLE VARIATION OF THE CALCITURBIDITES

Change In Carbonate Input

The layers with carbonate detritus vary considerably throughout the hole. The most prominent change is upward disappearance of the gray carbonate turbidites and their replacement by the green, phosphoritic type (Figure 14). Aboard ship, this change was used to define the boundary between lithologic Units VI and VII (see Site 416 Report this volume). Another, more gradual change is the upward disappearance of the green turbidite. The topmost beds of carbonate clastics are two breccias in Core 7 which differ from all other beds in their coarse grains (up to 2 cm in diameter). They contain mainly lithoclasts of shallow-water limestone and of sucrosic dolomite, as well as finer layers of quartz sand with some carbonate grains and phosphorite.

The most important changes throughout the turbidite sequence are summarized in Figure 14. Grain types and grain sizes were point counted in thin section. Mean bed thickness was measured aboard the ship. The section has a minimum mean bed thickness at 1420 meters, where the gray turbidites with their thick layers of micri-

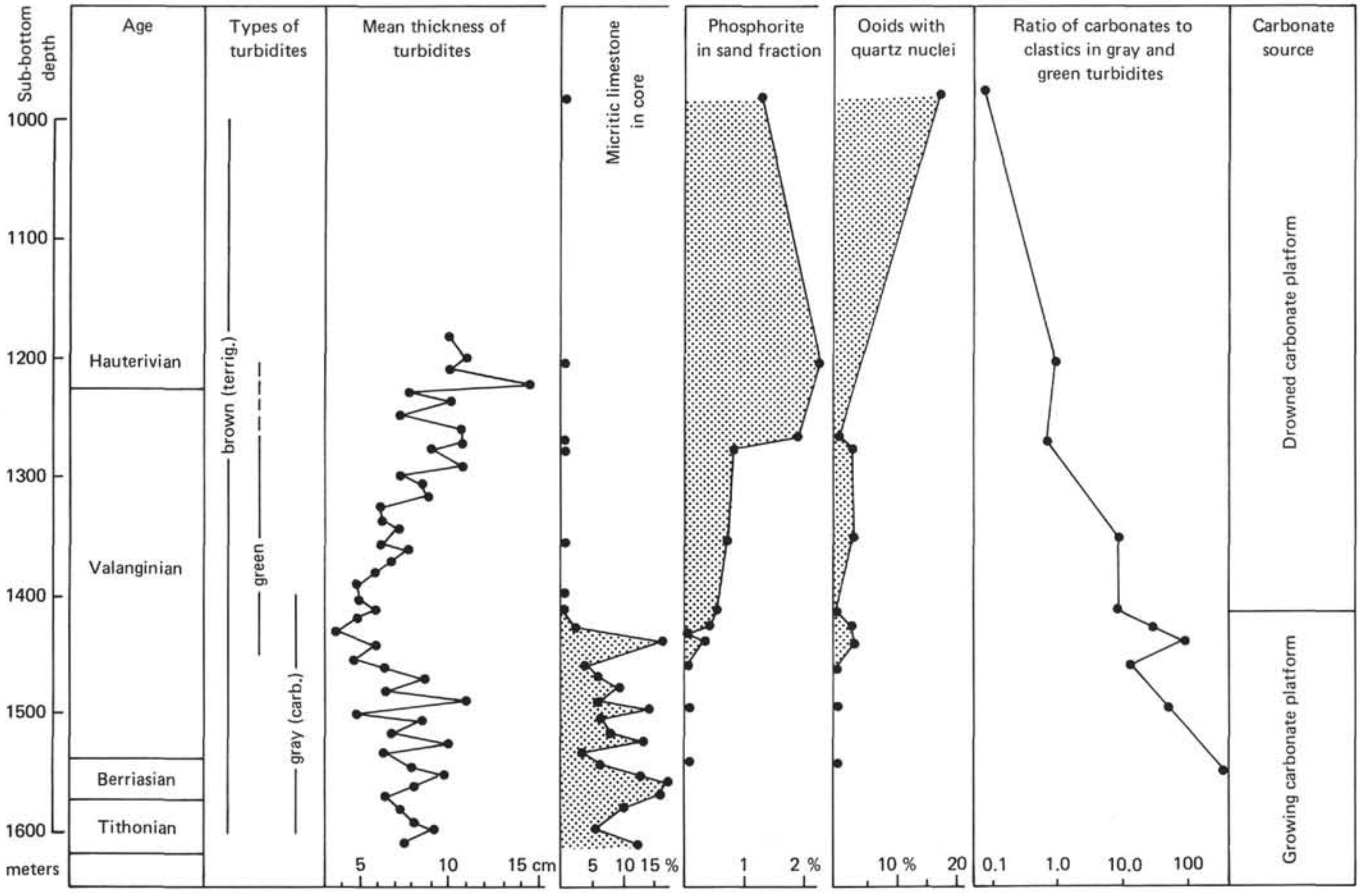


Figure 14. *Growth and drowning of the carbonate platform as interpreted from the turbidite record. Period of presumed platform growth (Tithonian–Valanginian) is marked by abundant micritic limestone of shallow-water(?) mud, by high carbonate-to-quartz ratio and by scarcity of phosphorite. Drowning of platform during Valanginian causes disappearance of micritic limestone, decrease in carbonate-to-quartz ratio, and increase of phosphorite and of number of ooids with quartz nuclei. Bed thickness of total core drops with drowning of platform because of disappearance of thick beds of micritic limestones. Increase in bed thickness above 1400 meters is probably caused by greater terrigenous supply. Bed thickness and thickness of micritic limestone are on basis of shipboard core description; other columns on basis of point counts in thin sections: 100 points per sample for carbonate-to-quartz ratio, 100-300 points for ooids with quartz nuclei, 2000 points per sample for phosphorite.*

tic limestones disappear. (The upward increase in bed thickness above this minimum is probably the result of an increase in terrigenous supply, rather than a change in carbonate contribution.) The disappearance of the micritic limestone coincides with the increase in phosphorite in the sand fraction from traces below Core 37 to values around 1 per cent and higher in Cores 36 to 7. At the same time, larger amounts of quartz appear in the green turbidites, reducing the ratio of carbonate to siliciclastic grains. The increase in ooids with quartz nuclei, and the persistent and clear separation of green and brown terrigenous turbidites in the cores suggest encroachment of clastics on the former carbonate source area, rather than increased mixing during downslope transport. None of the major lithologic changes shown in Figure 14 correlates with a biostratigraphic boundary. It is thus unlikely that the changes in lithology are caused by hiatuses in sedimentation.

Interpretation—Drowning Of A Carbonate Platform

Clearly during the Valanginian the carbonate contribution ("green" and "gray" turbidites) changed in character, whereas the siliciclastic contribution ("brown turbidites") varied only in distance from source. In Cores 57 through 37 the high carbonate content of the gray turbidites in combination with the abundance of green algae and shallow-water mud indicate that they were derived from a carbonate source area in warm shallow water, within the euphotic zone, in which almost no siliciclastic material was present. The green layers of Cores 37 through 12 are probably derived from a winnowed shelf below the euphotic zone, as indicated by the scarcity of green algae, the lack of shallow-water mud, and the abundance of phosphorite—a rock that deserves special attention in this discussion. The precipitation of calcium phosphate in the marine environment is favored by the same conditions as the formation of calcium carbonate, but in addition requires a high concentration of phosphate ion in sea water. Consequently, phosphates form in shallow waters of high fertility but, unlike carbonates, in depths just below the euphotic zone, where the phosphate ions are not depleted by living organisms, and rates of carbonate deposition are sufficiently low so as not to mask the relatively slow deposition of phosphate (Gulbrandsen, 1969; Blatt et al., 1972, p. 551–555).

This change in carbonate contribution could have been brought about either by a shift to a different source area or by a change in the depositional environment within the same source area. The abruptness of the change and the fact that the carbonate-rich "green" turbidites as a group remain distinctly separate from the terrigenous "brown" turbidites argue against a shift to another source area and point to an environmental change within the carbonate source.

Drowning of the shallow-water carbonate source area and its transformation into a current-swept, deeper-water shelf below the euphotic zone best explains the disappearance of green algae and shallow-water-mud contribution, as well as the appearance of phosphorite. The ooids, hardened peloids, and lithoclasts of Cores 36 through 12 may be relict sands that can persist in such an environment for considerable time (e.g., the relict sands on present-day open shelves; Pilkey et al., 1966; Fabricius et al., 1970; Ginsburg and James, 1974, p. 137–144). On isolated highs of the former platform, active ooid shoals may have persisted long after the main part of the platform had subsided below the zone of maximum carbonate production.

The drowned platform continued to shed carbonate detritus, albeit at a steadily decreasing rate, until the late Hauterivian. The uppermost beds of carbonate clastic sediments are two breccias in Core 7. Their granule-size clasts of shallow-water limestone and of sucrosic dolomite, and the predominance of quartz in the terrigenous finer layers, indicate that by Hauterivian time erosion had cut into the lithified and dolomitized part of the platform and that terrigenous sediments were encroaching on it.

The picture of a carbonate continental margin fits well with the regional geology. Throughout the Jurassic the shallow seas in northwestern Africa were largely a domain of carbonate deposition (Dresnay, 1971). Upper Jurassic shallow-water carbonates crop out on the Mazagan escarpment, 100 km east of Site 416 (Renz et al., 1975) and are widespread in the subsurface of the continental slope (Lehner and de Ruiter, 1977; Todd and Mitchum, 1977).

On shore, in the Essaouira basin, the Souss trough and the Tarfaya basin, the shoreward part of these carbonate platforms is exposed. It consists of dolomitic limestone with red beds (Ambroggi, 1963, p. 123-134; Societe Cherifienne, 1966; Wiedmann et al., 1978; Behrens et al., 1978). Sparse biota and occasional evaporites indicate their deposition in restricted shallow-water environments. During late Berriasian and Valanginian time, the open marine environment expanded rapidly eastward, and the lagoonal deposits were covered by ammonite-bearing marl, sandstone, and detrital limestone. Condensed faunas, hardgrounds, and hiatuses indicate slow and intermittent deposition (Ambroggi, 1963, p. 123-124; Societe Cherifienne, 1963; Wiedmann et al., 1978; Behrens et al., 1978; Price, this volume). The rapid change from a highly restricted to an open deeper-water marine environment in the Western Atlas and the Essaouira Basin coincides with the postulated drowning of the carbonate platforms on the continental slope. One is tempted to link the transgression in the onshore basins with the drowning of the offshore platforms and assume that a rise in sea level of a period of increased subsidence was the common cause of both. The intermittent sedimentation in the onshore sections may indicate that sea level rose in steps, possibly separated by episodic lowering. A slight drop followed by a rapid rise in sea level is probably the most efficient way to drown a carbonate platform. I believe that healthy platforms can keep up with very high rates of subsidence and rises of sea level; their tolerances is reduced, however, in the critical time span when carbonate production is slowly resumed after a period of exposure.

On the basis of West African and global data, Todd and Mitchum (1977) propose a rapid and deep drop of sea level in the Valanginian, separating periods of rise in the Portlandian-Berriasian and in the late Valanginian-Hauterivian. Our data on Hole 416A are compatible with this hypothesis, inasmuch as they, too, point to an important intra-Valanginian event. However, the continuity of the turbidite record, as well as the onshore stratigraphy by Ambroggi (1963) are not easily reconciled with an extreme Valanginian low stand. With Price (this volume) I favor the interpretation of a stepwise rise in sea level during the whole Portlandian-to-Valanginian interval, interrupted only by minor low stands.

SITE 416—THE FIRST TITHONIAN SEQUENCE BELOW THE CCD IN THE ATLANTIC

The shale-sandstone-limestone alternation of the Tithonian-Hauterivian interval at Site 416 was deposited below calcite-compensation depth, with brown clay as perennial sediment and turbidite input of varying provenance in episodic pulses. This statement, made tentatively after shipboard examination of the cores (see Site 416 Report, this volume), has been fully confirmed by the present study. The carbonate occurs in graded beds, deposited in one pulse, with all the characteristic features of turbidites. Even the fine carbonate fraction consisted largely of shallow-water material, and the coarse layers contain all transitions from pelagic limestone to brown clay in the form of "rip-up" clasts, picked up during transport downslope. Site 416 lay below the compensation depth, but the continental slope adjacent to it was covered with carbonate sediment and crowned by carbonate platforms at the continental shelf.

Other DSDP holes that reached Tithonian strata in the Atlantic include those drilled at Sites 4 and 5 (Ewing, Worzel, et al., 1969); Sites 99, 100, 101, and 105

(Hollister, Ewing, et al., 1972), and Site 391 (Benson, Sheridan, et al., 1978) in the western Atlantic; and Site 367 (Lancelot, Seibold, et al., 1978) in the eastern Atlantic. In all these holes the Tithonian sequence comprises pelagic carbonate sediments; turbidites with shallow-water debris are scarce, although extensive carbonate platforms existed on both sides of the young ocean.

The difference of facies between the sediments of Site 416 and the rest is consistent with the location of Site 416 far within the magnetically quiet zone, on crust that could be as old as Early Jurassic (Pittman and Talwani, 1972; Lancelot, Seibold, et al., 1978). By Tithonian time this crust must have subsided below the calcite-compensation depth; it lay adjacent to the continental rise and was thus within reach of shallow-water debris from the continental shelf. The other sites, on younger crust (mostly Callovian-Oxfordian), still stood high on the flanks of the mid-ocean ridge, above the compensation depth, and beyond reach of turbidity currents sweeping down the continental slope. This explanation holds for all sites except for Site 391 in the Blake-Bahama basin of the western Atlantic. It is located within the magnetically quiet zone, at about the same distance from its boundary as Site 416, and very close to what has been shown to be a Neocomian (and possibly older) carbonate-platform margin (Benson, Sheridan, et al., 1978). The setting of Site 391 is therefore an almost exact mirror image of Site 416. Yet the Tithonian-Neocomian of Site 391 is almost purely pelagic limestone, deposited above calcite-compensation depth and without shallow-water-carbonate influx. This discrepancy needs further investigation.

CONCLUSIONS

Calciturbidites, although only a fraction of the terrigenous contribution in the Jurassic-Cretaceous sequence of Site 416, provide a considerable amount of information about the adjacent continental shelf and slope. In particular, they seem to record fairly accurately the growth and the drowning of a carbonate platform.

Carbonate platforms are common along passive continental margins and are probably the most sensitive indicator of sea-level changes along the seaboard of continents. Their initiation, the rate of their growth, and the timing of their death can often provide a key to both the subsidence history of passive margins and eustatic sea-level fluctuations in the geologic record.

Turbidite sequences not only allow us to recognize these events at some distance from the platform, but also to date them fairly accurately by means of the intercalated pelagic sediments.

ACKNOWLEDGMENTS

I thank my colleagues of the scientific party aboard the *Glomar Challenger* for stimulating discussions and exchange of ideas and observations during Leg 50. I thank O. Joensuu (University of Miami) for help with atomic-absorption and X-ray analyses; R. N. Ginsburg (University of Miami) and E. A. Shinn (United States Geological Survey) kindly reviewed the manuscript.

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