34. SEDIMENTS AND SEDIMENTARY HISTORY OF THE MANIHIKI PLATEAU, SOUTH PACIFIC OCEAN

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ABSTRACT

The Manihiki Plateau is an elevated area of shallow oceanic crust in the South Pacific; it was formed during Early Cretaceous time by profuse outpourings of basalt. DSDP Holes 317, 317A, and 317B, on which this account is based, are situated in a basinal area on the so-called High Plateau, the shallowest part of the volcanic edifice, where depths range between 2000 and 3000 meters. The oldest sediments cored are Lower Cretaceous greenish-black volcaniclastics traversed by reddish-purple clay seams; these deposits commonly show evidence of redeposition, but the only possible shallow-water remains found are very rare bryozoan and echinoderm fragments. Olive-black sandstones rich in organic material were sampled at one level. Millimeter-scale spherules, with clay-rich rinds, occur locally. Two textural end members may be distinguished; one is a sandstone containing rounded grains of green ?nontronite and brown sideromelane altering to palagonite, analcite, and clay minerals; and the other is a hyaloclastite comprising fused glass shards set in a matrix of clay. These sediments have a chemical composition close to deep-ocean basalt, but are relatively enriched in K and depleted in Ca. Such changes are comparable with those described for basic igneous rocks undergoing submarine alteration to clay minerals, particularly K-rich smectites. Native copper occurs as blebs and strands within the volcaniclastic sediments up to 150 meters above basaltic basement; its origin is ascribed to "hydrothermal solutions" produced by basalt-sea water reactions that in turn were related to a geothermal system circulating through the top level of ocean crust and overlying sediments. The reddish-brown clay seams in the volcaniclastics may be the record of the conduits for the cupriferous solutions.

The volcaniclastic sediments grade upwards into limestone, chalk, oozes, and claystone which all locally contain cherts. These deposits bear witness to the increased importance of biogenous material as a sediment supplier when volcanic activity waned on the Manihiki Plateau during late Early Cretaceous time. The bivalves that colonized the surface of the edifice during this interval reflect increase in depth. Trace fossils, particularly *Planolites*, produced numerous fucoid burrows in the surficial sediments. Brown claystones, rich in fish material, may indicate high dissolution levels during Albian-Campanian time. Diagenetic processes included dissolution and precipitation of calcium carbonate to produce limestones and growth of replacement lussatitic and quartz chert. The youngest chert is of Oligocene age.

Above these variable sediments only chalks and oozes (Oligocene to Pleistocene) occur; these show some evidence of burrow mottling. Grayish-orange and bluish-white sediments alternate; these are ascribed to low and high sedimentation rates, respectively. The presence of grayish-orange sediments at the surface of the Manihiki Plateau today, together with the local occurrence of ferromanganese nodules, might suggest that net depositional rates, in some areas at least, are relatively slow at the present time.

INTRODUCTION

The Manihiki Plateau is an area of elevated ocean floor in the South Pacific (Figure 1). Depths in its shallowest parts range from 2400 to 3000 meters (Heezen et al., 1966). This plateau is characterized by a variable but considerable thickness of pelagic sediments of differing types (e.g., Figure 2). The topography of the igneous basement varies somewhat and Holes 317, 317A, and 317B were drilled in a basinal area of the so-called High Plateau (Winterer et al., 1974); the sedimentary pile is presumably thicker here than in neighboring regions. The water depth over the drill site is 2598 meters.

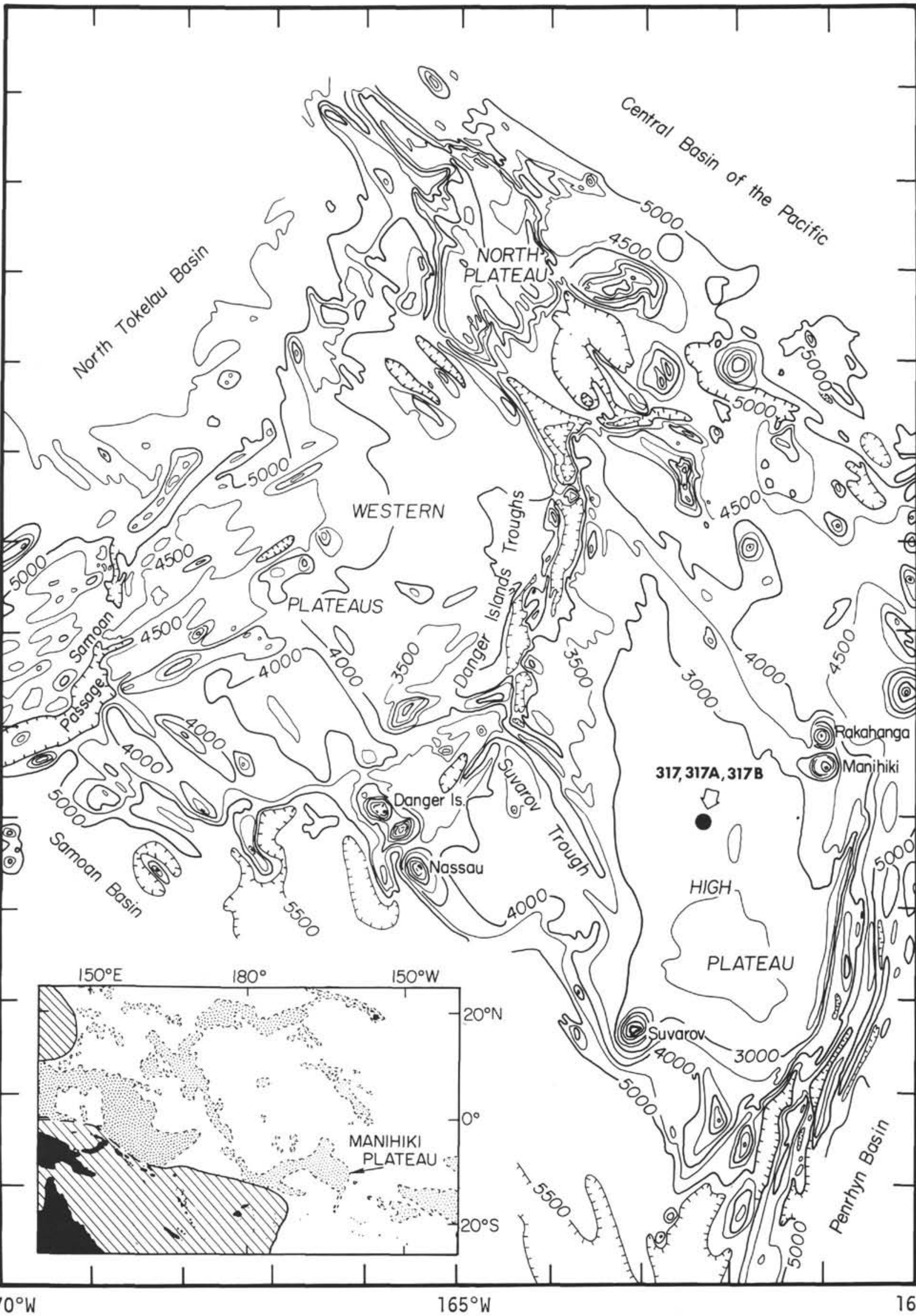

Figure 1. *Bathymetric chart of South Pacific showing location of Manihiki Plateau. Holes 317, 317A, 317B were situated at 11° 00.09'S, 162° 15.78' W. Contour interval = 500 m. Stippled areas are shallower than 4500m; ruled areas are outside the Pacific plate. (Map courtesy of E. L. Winterer).*

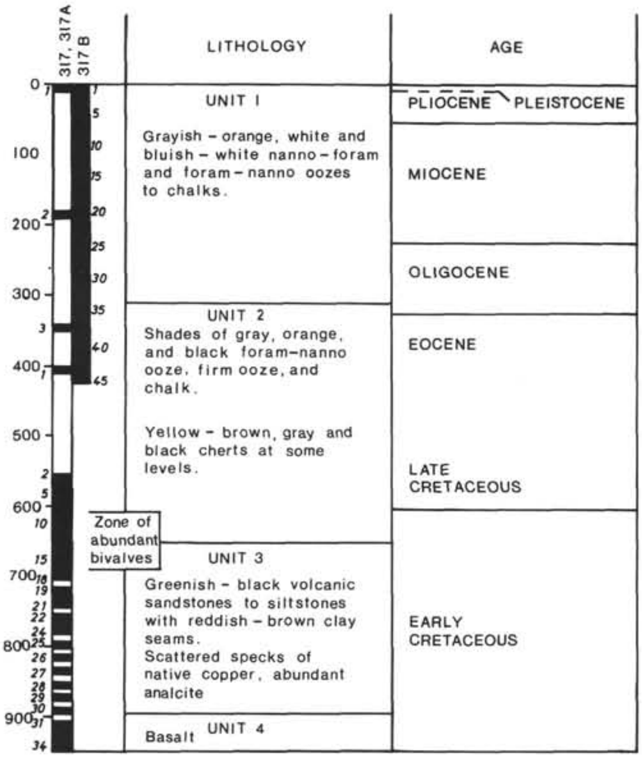

Figure 2. *Stratigraphical section of holes drilled on Manihiki Plateau (Holes 317, 317A, and 317B).*

According to Winterer et al. (1974) the Manihiki Plateau represents the result of Early Cretaceous volcanic effusions related to a melting spot in the mantle probably situated near a lithospheric triple junction—that is, that the plateau is a volcanic edifice roughly comparable in origin to Iceland.

VOLCANICLASTIC SEDIMENTS (UNIT 3)

The oldest rocks cored on the Manihiki Plateau comprise dark red, and grayish-brown to greenish-black volcaniclastic sandstones to siltstones. The basal sediments are intercalated with vesicular basaltic flows, and in this part of the section red-colored sediments are most abundant. Higher up, the sediments take on their more typical greenish-black hue and gradually become paler as the amount of carbonate increases, and bivalves and fucoid burrows become common. Sediments near the top of the unit include partially silicified nannofossil limestones.

Structures

Many of the volcaniclastic sediments exhibit grading, this being most pronounced in thin (ca. 1 cm) beds that are common in the basal part of the section. Higher up, where the beds are more calcareous, the grading is poor but cross-laminated units, with burrowed tops, are developed; within one sandstone bed thin layers of differing grain size are commonly present.

Conspicuous in much of the volcaniclastic sediment is evidence of hard- and soft-sediment deformation; this is manifested as a crinkling and angular distortion of beds and by the development of breccias (Figure 3). These

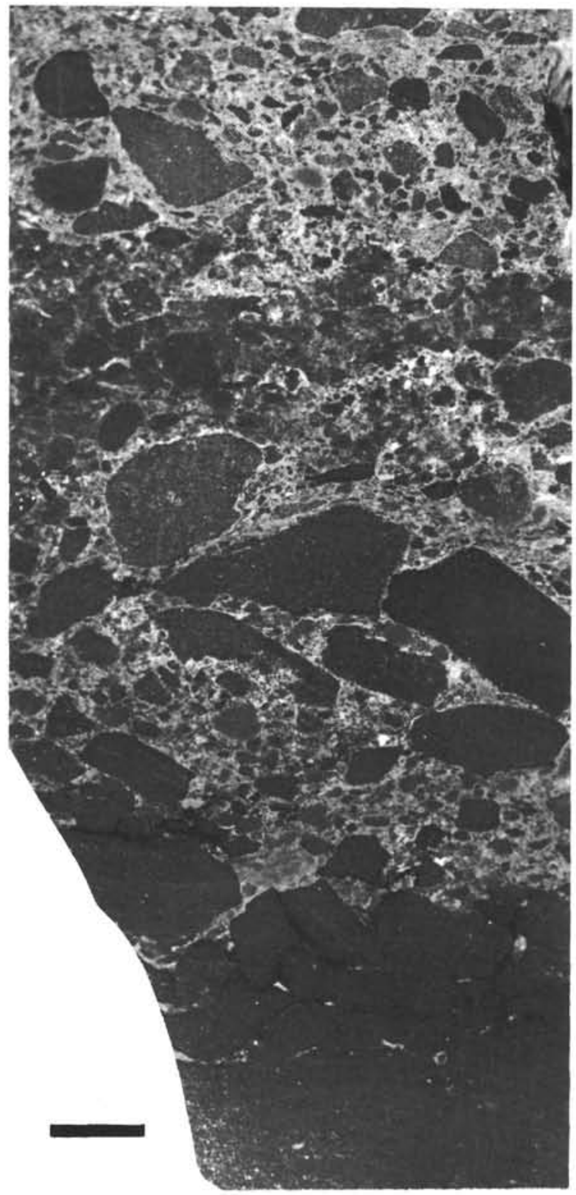

Figure 3. *Bed of breccia comprising greenish-black volcanogenic sandstone and siltstone in lighter clay-rich matrix. Various stages of brecciation are evinced, from simple rupture of blocks (at base) to complete separation and enclosure in matrix (higher in section). Polished face of 317A-16-5, 30-43 cm. Lower Cretaceous. Scale bar = 2 cm.*

breccias are characterized by the presence of dark greenish-black volcaniclastic siltstone to sandstone isolated by an anastomosing network of reddish-brown and purple clay-rich material (Figure 4). This latter material has apparently acted as a relatively fluid substratum and allowed differential movement of discrete consolidated sandstone layers. This, of course, presupposes early lithification of at least some of the volcaniclastic sediments. The angles of the brecciated zones (5°-10° to the bedding), and the change in angle across the core, suggest that these disturbed zones may in fact be small-scale manifestations of large slump folds.

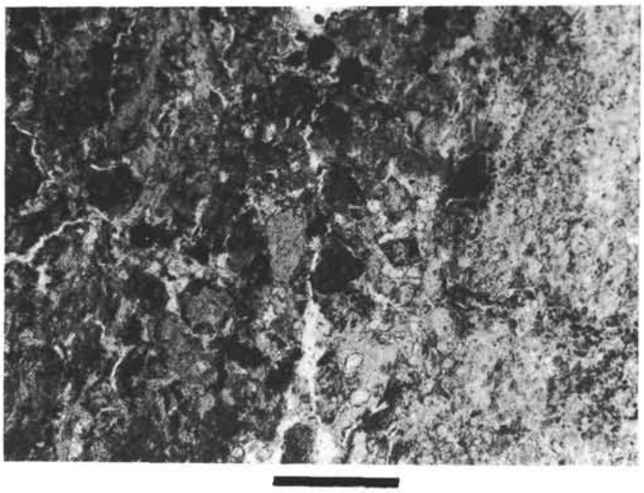

Figure 4. *Clay seam in contact with rounded volcanogenic grains, chiefly altered sideromelane. Contact between clay seam and volcanogenic sandstone is relatively sharp. Thin section from 317A-23-2, 124-128 cm. Lower cretaceous. Scale bar = 0.25mm.*

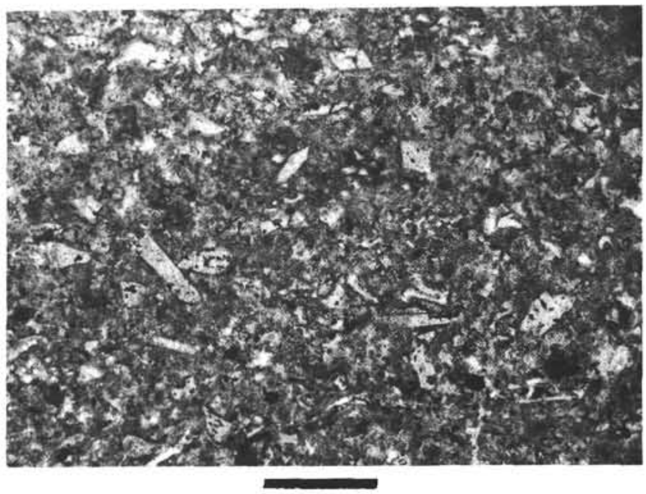

Figure 5. *Hyaloclastite containing glass shards of sideromelane partially devitrified to clay minerals, plus plagioclase and clinopyroxene grains. Matrix is clay and very fine-grained glass. Thin section from 317A-14-3, 130-140 m. Barremian-Aptian. Scale bar = 0.4 mm.*

Textures

The volcaniclastic sediments vary a good deal in texture; this variation is imposed by the size of the constituent grains, their degree of roundness, presence or absence of precipitated cement, and the relative abundance of clay-sized material. Two textural end members may be distinguished: one is clearly a hyaloclastite, containing small (0.1-0.2 mm), commonly curved shards of altered glass (sideromelane) changing to palagonite and smectitic clay minerals, plus feldspar and clinopyroxene crystals (Figure 5). All the constituents are welded together in a matrix of what is now clay. Many of the shards have clearly constituted the outer walls of vesicles, and some glassy fragments still contain vesicular hollows filled with clay-rich material. Silica-rich, probably radiolarian, spheres occur in discrete clay-ridden clasts. The other textural end member can perhaps best be described as a sandstone where rounded grains (0.1-0.3 mm, Figure 6a) of green material (?nontronite) commonly with a brown ferruginous rind, and brown sideromelane altering to palagonite, are set in a matrix of clay and fine-grained calcite. In the higher levels of the succession this facies is commonly associated with thick- and thin-shelled bivalves (Figure 7), echinoderm debris, rare bryozoan material, foraminifers, and sparse nannofossils. Additionally, Kauffman (this volume) records small gastropods and serpulid tubes.

In both of the above textures coarse-grained cements of subhedral calcite and euhedral analcite are common (Figures 6b, 8); analcite is also present as discrete detrital grains. In large void spaces (3-4 mm) calcite may occur around the edge with analcite as a central fill, but this relationship is not invariable. In some cases volcanic grains are enclosed in a poikilitic calcite cement where a number of intergranular voids are filled by what is optically one crystal, and this crystal is adjacent to several others of different crystallographic orientation. When space permits, calcite crystals are often relatively large (400-1000 μm) and in general do not show the size increase from void margins outwards typical of "normal" passive cements (cf. Bathurst, 1971, p. 417-425). Calcite may be a pseudomorph after ferromagnesian minerals.

As mentioned above reddish-brown and purple seams are common in the lower levels of the volcaniclastic sediments, and these have apparently acted as a substratum for mass movements. These seams are composed of clay minerals and opaque iron oxide-hydroxides that are streaked out parallel to the walls of the seam; volcanic grains are entirely absent (Figure 4). The boundary of the seam with the surrounding volcaniclastic sediment is relatively sharp.

Chemistry

A number of chemical analyses were carried out on crushed and washed samples of the volcaniclastic sediments (Tables 1 and 2). These analyses were undertaken on the PW1212 X-ray spectrometer at Durham University (see Jenkyns and Hardy, this volume); analytical techniques are fully described by Holland and Brindle (1966). These sediments are chemically very close to the mean composition of ocean-floor basalt as computed by Cann (1971) except for some notable departures (Table 3). K_2O, for example, is greatly enriched over its basaltic parent material. CaO is largely depleted. MgO has apparently behaved rather erratically but generally is enriched in the volcaniclastics. P_2O_5 and possibly Na_2O have been lost. The trends evinced by altered basalts have been described by many authors (e.g., Hart, 1970; Matthews, 1971; Melson and Thompson, 1973), although results on the behavior of some elements have been conflicting. All authors agree on gain of K_2O and loss of CaO, but the behavior of Na_2O, MgO and P_2O_5 in basalt-sea water interactions is more problematic. According to Hart (1970), on hydration of basalt, MgO is lost and Na_2O and P_2O_5 gained. Matthews (1971) also recorded loss of MgO, but whereas his sample of fresh basaltic glass contained 0.15% P_2O_5, the associated

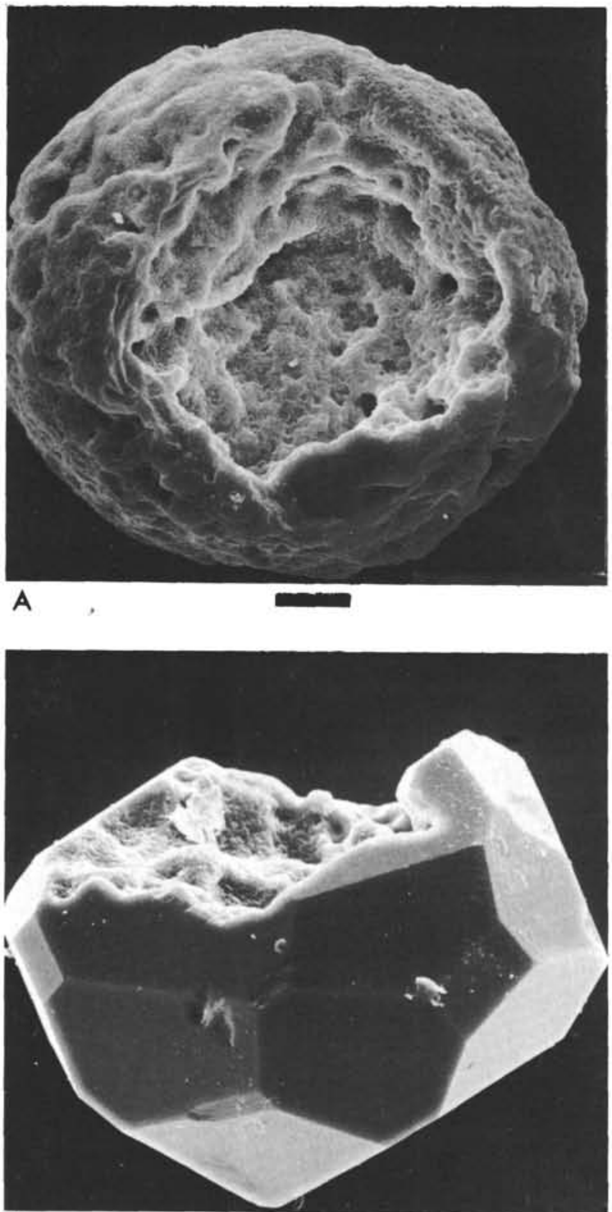

Figure 6. *Scanning electron micrographs of grains from 317A-25-2, 0-5 cm. (a) rounded and pitted volcanogenic particle. (b) euhedral analcite showing well-developed crystal terminations, and near-cubic symmetry. Lower Cretaceous. Scale bar = 25 µm.*

palagonite assayed at only 0.01%: this latter result is in agreement with the data on the Manihiki volcaniclastic sediments. Melson and Thompson (1973), however, studying the alteration products of hyaloclastites and pillow basalts from St. Paul's Rocks in the Atlantic, recorded a gain in MgO and loss of Na_2O; their analyses closely parallel those presented here.

The fact that the major-element analyses of the volcaniclastic sediments consistently add up to less than 100% suggests a considerable amount of hydration.

In terms of their trace-element composition the Manihiki volcaniclastic sediments also show strong

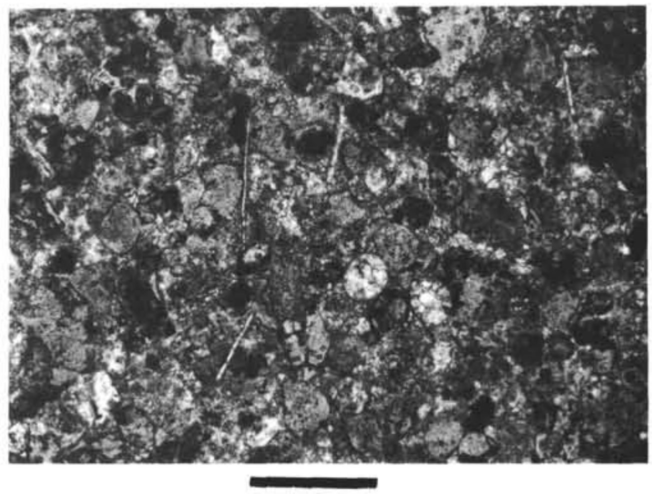

Figure 7. *Volcanogenic sandstone containing thin-shelled bivalves, commonly recrystallized, sparse echinoderm and bryozoan material with very rare foraminifers. Grains are generally well rounded; a few are distinctly angular. They comprise brown sideromelane, a green clay mineral (?nontronite) that usually has a dark brown ferruginous rind, some analcite, and calcite-filled spheroids that conceivably represent reworked vesicle fills. Matrix comprises clay and fine-grained calcite. Thin section from 317A-16-1, 140-150 cm. Lower Cretaceous. Scale bar = 0.4 mm.*

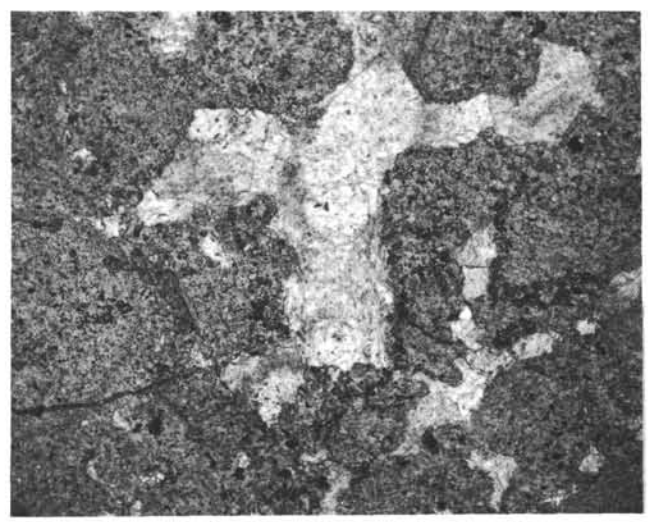

Figure 8. *Silt-sized volcanogenic grains of altered sideromelane and ?nontronite in clay-rich matrix. Millimeter-scale voids contain fills of calcite and analcite. Thin section from 317A-25-3, 145-150 cm. Lower Cretaceous. Scale bar = 1 mm.*

resemblances to oceanic basalt (cf. Philpotts et al., 1969; Cann, 1970; Nicholls and Islam, 1971), although some elements have been lost, while others have been gained. Ba, for example, seems to have been lost in most sedimentary samples but its behavior has not been consistent; this is in agreement with the results of Philpotts

TABLE 1
Major Element Analyses for Volcaniclastic Sediments of Hole 317A

Sample (Interval in cm)	SiO_2	Al_2O_3	Fe_2O_3	MnO	MgO	CaO	Na_2O	K_2O	TiO_2	S	P_2O_5	Total
18-1, 140-146	52.89	19.86	10.69	0.04	2.70	1.88	1.41	1.49	1.40	0	0	92.36
22-2, 103-109	53.40	15.30	13.58	0.09	5.59	2.16	1.40	1.48	0.97	0	0	93.97
23-3, 16-19	53.04	15.54	11.89	0.12	7.58	2.09	1.04	2.04	1.03	0	0.06	94.44
24-2, 40-43	51.12	14.24	13.37	0.15	10.03	1.63	1.16	2.41	1.26	0	0	95.37
25-5, 73-76	51.46	14.58	10.76	0.18	12.11	1.68	1.49	2.17	1.04	0	0.05	95.52
29-1, 135-140	51.87	13.85	10.76	0.14	13.85	1.60	2.26	1.27	1.08	0	0.05	96.73
29-5, 0-3	51.12	13.66	12.73	0.68	9.90	1.58	1.14	2.41	0.92	0	0	94.14
30-4, 124-133	51.61	14.85	10.81	0.16	10.42	2.94	1.61	3.00	0.98	0	0.13	96.51

TABLE 2
Minor Element Analyses (in ppm) for Volcaniclastic Sediments of Hole 317A

Sample (Interval in cm)	Ba	Nb	Zr	Y	Sr	Rb	Zn	Pb	Cu	Ni	Cr	Co
18-1, 140-146	22	3	76	9	833	26	129	0	257	100	572	40
22-2, 103-109	0	4	53	28	127	25	183	0	127	188	403	80
23-3, 16-19	0	4	55	11	113	33	260	8	4099	182	654	65
24-2, 40-43	10	2	69	18	94	46	100	12	176	123	244	60
25-5, 73-76	0	5	57	19	95	20	107	0	161	142	489	59
29-1, 135-140	0	3	63	17	76	16	81	0	159	264	768	65
29-5, 0-3	0	2	29	44	70	34	117	–	–	–	–	–
30-4, 124-133	0	4	56	31	75	35	117	0	113	156	622	59

TABLE 3
Average Composition of 317A Volcaniclastic Sediments (with Fe_2O_3 Recomputed to FeO) Versus Mean of Ocean-Floor Basalt as Given by Cann (1971)

	Mean of Ocean-Floor Basalt	Mean of Volcaniclastic Sediments
SiO_2	49.61	52.06
Al_2O_3	16.01	15.23
FeO	11.49	10.64
MnO	0.18	0.19
MgO	7.84	9.02
CaO	11.32	1.94
Na_2O	2.76	1.44
K_2O	0.22	2.03
TiO_2	1.43	1.08
P_2O_5	0.14	0.04

et al. (1969). Nb and Zr do not depart significantly from values in unaltered basalt (Cann, 1970); Y is a little more variable. Sr, apart from the anomalously high value of 833 ppm—possibly related to the presence of aragonite in a carbonate phase—does not depart significantly from values in fresh (or altered) basalt (Philpotts et al., 1969; Cann, 1970). Rb contents are low in fresh basalt and this element is known to be enriched in weathered alteration products (e.g., Philpotts et al., 1969; Cann, 1970) where presumably follows potassium. The higher Rb contents of the Manihiki Plateau volcaniclastics are comparable with those of certain weathered basalts. Zn and possibly Pb are present in values corresponding to basalt (Taylor, 1964). Cu is in general slightly higher in the volcaniclastics than in deep-sea basalt (cf. Nicholls and Islam, 1971) which is not surprising since native copper occurs at some levels (see below: Figure 9). The 4099 ppm Cu in one sample reflects the presence of the metal. Ni and Co have values comparable to those of basalt; Cr is marginally higher in the sediments than in basalt (cf. Nicholls and Islam, 1971).

In Sample 317A-16-2, 133 cm 28.7% organic carbon has been recorded from an olive-black sandstone containing pyritic burrows; this lithology is also present in the top part of Core 16, Section 3.

Mineralogy

Major minerals present in the volcaniclastic sediments are montmorillonite, including ?nontronite, clinopyroxene, sanidine, plagioclase feldspar, analcite, calcite, with traces of hematite and pyrite. A large amount of material visible in thin section is undoubtedly X-ray amorphous; this includes glass and a certain amount of iron oxide-hydroxide.

Native Copper

Native copper was observed as blebs and strands (Figure 9) in volcaniclastic sediments up to some 150 meters above basaltic basement. It is present in both red and greenish-black siltstones and sandstones. A greenish patina encases the metal and discolors the immediately enclosing sediment. Microprobe analysis of a sample (317A-25-2, 0-5 cm) revealed more than 99.5% metallic Cu, with a trace of sulphide (0.03% S).

Spherules

Small spherules (1-4 mm) are common in the volcaniclastic sediments (Figure 10). These bodies are

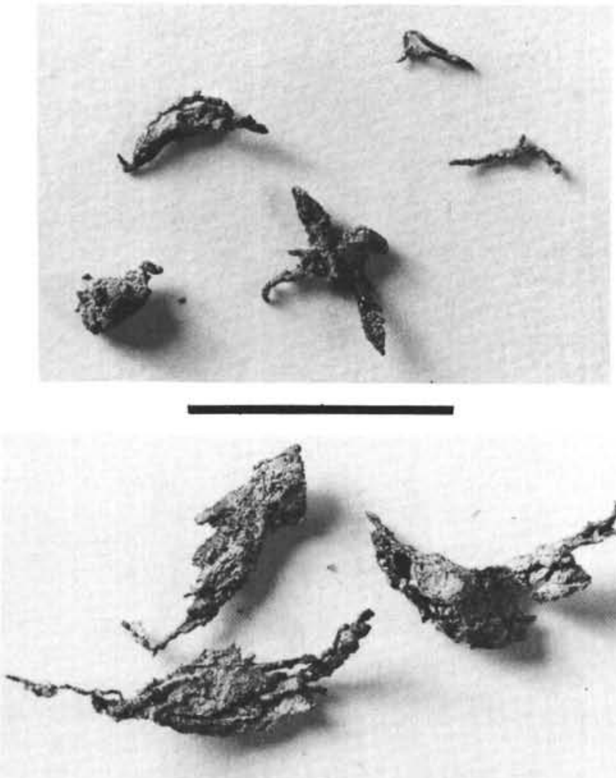

Figure 9. *Strands, wires, and blebs of native copper extracted from greenish-black to dusky red volcanogenic siltstone crossed by reddish-brown clay veins: 317A-23-3, 16-19 cm. Lower Cretaceous. Scale bar = 1 mm.*

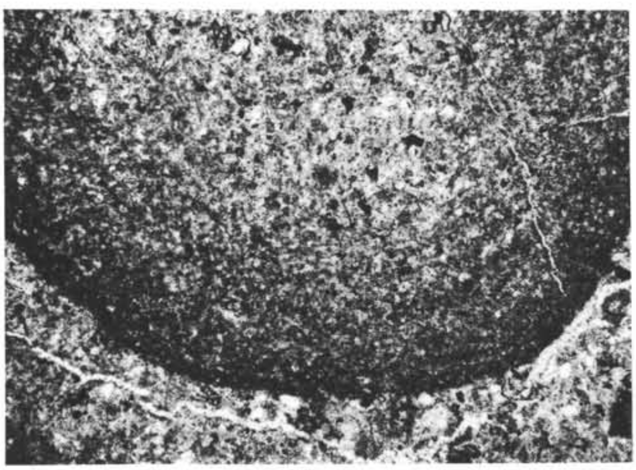

Figure 10. *Part of a spherule and matrix. The border of the spherule is finer grained and richer in clay than both the interior and the enclosing matrix. Volcanogenic particles, including abundant clinopyroxene and sideromelane, are chiefly of sand size and set in a matrix of clay. Thin section from 317A-29-2, 13-21 cm. Lower Cretaceous. Scale bar = 0.5 mm.*

characterized by cores of greenish-black volcanogenic material comparable to the matrix in every way, and darker, reddish-brown more clay-rich rinds. The rinds in some cases are surrounded by a smattering of calcite crystals that serve to distinguish further the spherules from the matrix. In other cases the spherules appear to blend into the host rock. Some spherules are present as hemispheres and thus are seemingly broken with only half present. The presence of the spherules is not linked to any particular grain size of matrix; they occur even in the finest fraction of graded layers.

Depositional Processes

Since Holes 317, 317A, 317B were located in a depression in the Manihiki Plateau, resedimentation processes were presumably more active in this region than they were on neighboring high ground. The cored section therefore should not be taken as a model for sedimentation over the whole of the High Plateau.

The volcaniclastic sediments clearly derive from the violent interaction of basaltic lavas and sea water. Those with hyaloclastic texture were presumably erupted in shallow water (e.g., Bonatti, 1967; McBirney, 1971), a conclusion supported by the highly vesicular nature of some of the underlying flows (Jackson et al., this volume). The eruptive force behind the formation of hyaloclastites is an obvious source for currents (possibly turbidity currents) that deposited the graded sandstone layers. Sandstones with rounded grains (e.g., Figures 6a, 7) are, however, as common as hyaloclastites. These are more of a puzzle as the mechanism whereby the grains become rounded is difficult to envisage in an environment where the amount of reworking they could have suffered was presumably very modest. Photographs of altered glassy abyssal basalts from St. Paul's Rocks on the Mid-Atlantic Ridge (Melson and Thompson, 1973) illustrate sideromelane with well-developed perlitic structure; many of these fractures are tightly curved. On disintegration and transport this altered glass would presumably liberate roughly rounded particles. Thus the rounded grains—which, in general, are further along the road to alteration than the more angular glassy material—could presumably be produced over a moderate time period by simple submarine weathering of disaggregated basaltic glass. This is in contrast to the hyaloclastites which were probably deposited from a turbidity current or suspension cloud directly after their eruption.

The absence of chert in this unit, apart from some slightly silicified layers toward the top, could either reflect dilution of siliceous organisms by volcaniclastics or reduced fertility over the plateau during the Mid-Cretaceous. Radiolarian-bearing clasts are, in fact, present in certain redeposited beds; and the plate-tectonic reconstructions of Winterer et al. (1974) suggest that the Manihiki Plateau was never more than 15° south of the equator during the whole of its lifetime. This latitude is within the present-day equatorial zone of abundant biogenic silica manufacture (e.g., Ramsay, 1973). However, the paleomagnetic results of Cockerham and Jarrard (this volume) suggest that the plateau originally lay in latitudes more southerly than this; if their data are accepted at face value, then the lack of chert in the earliest Manihiki sediments could be related to paucity of biogenic silica. The lack of abundant reefal and reef-associated organisms in the earliest Manihiki sediments may also be pertinent to paleolatitude, since the vesicularity of the basalt and the presence of hyaloclastites implies a shallow initial depth. The only possible shallow-water remains found are derived bryozoan and echinoderm fragments. Clearly, reefs were established on some parts of the plateau since they evolved into present-day atolls such as Manihiki and Suvarov. It seems therefore that the waters around the drill site location may initially have been too deep or too cold to allow reef growth, or that intermittent volcanic episodes during the early stages of subsidence rendered colonization by reefs impossible. Possibly only a few areas of submarine high ground were shallow enough to favor reef growth when northward motion of the Pacific plate caused the Manihiki Plateau to enter warm equatorial waters.

The presence of bivalves and gastropods of "shallow-shelf" aspect in the top part of the volcaniclastic unit suggests that the plateau lay, at this time, some hundred or so meters below the surface of the sea (see Kauffman, this volume).

The local presence in the bivalve-bearing volcaniclastic sediments of a layer rich in organic material is noteworthy. Its age is probably at least Barremian-Aptian which makes it correlative with carbonaceous sediments drilled on Leg 11 in the western Atlantic (Hollister, Ewing et al., 1972) and somewhat older than the comparable Upper Cretaceous horizons that have been cored in the Caribbean and Atlantic (e.g., Hayes, Pimm et al., 1972; Edgar, Saunders, et al., 1973) and which also occur on land in the pelagic-carbonate sections of the Umbrian Apennines, Italy (e.g., Renz, 1936; Bortolotti et al., 1970). Such organic layers imply deoxygenated bottom waters and may therefore be the record of some "event" related to poor oceanic mixing. Some comments on the significance of this type of event are given by Berger and von Rad (1972). Bearing in mind the widespread nature of the Caribbean-Atlantic-Tethyan sapropelic layer, it may be instructive to search for comparable Barremian or older horizons across the Pacific basin.

Diagenetic Processes

A considerable amount of reaction between basaltic glass and sea water clearly took place during the early sedimentary history of the Manihiki Plateau, since the chemical analyses quoted above show certain significant departures from the composition of deep-ocean basalts. Redeposition processes were apparently active while these basalt-derived sediments were undergoing alteration, as suggested by the presence of discrete analcite grains in certain volcaniclastics, so there was some overlap between depositional and diagenetic processes. Alteration of basaltic glass clearly was a relatively fast process.

The data of Melson and Thompson (1973) on the composition of clay minerals derived from Atlantic-Ridge hyaloclastites and pillow basalts are particularly significant to the present study. Their suggestion that enrichment of K_2O (and Rb) is due to the formation of potassium-rich smectites would seem to be highly relevant to the case of the Manihiki volcaniclastic sediments. They suggest formation of the potassium-rich smectite at moderate temperatures (<400°C) in a mildly alkaline solution such as sea water. Thus, marine waters, heated by volcanic effusions, and reacting with basaltic glass to give clay minerals, could account for most of the chemistry of the volcaniclastics. According to Hart (1973) this type of reaction is but one of a series that characterizes the chemical exchange between oceanic waters and young oceanic crust.

The origin of the spherules is presumably related to the transformation of volcanic glass into clay minerals, since the bodies commonly possess green volcaniclastic siltstone cores and reddish-brown clay-rich rinds. The apparently oxidized nature of this clay rind poses an intriguing problem given that the greenish color of the material within and without suggests that iron in and on clay minerals is held in the divalent state. Inasmuch as they appear to blend into the host rock, they seem at first sight to be small bodies that have been generated in situ. Their presence in the fine-grained fraction of a graded bed certainly belies hydraulic transport. On the other hand, the presence of some spherules that are apparently broken, with only the broken half present, suggests that they must have existed as free entities on the sea bottom. To resolve this one could suggest that the spherules formed by subsurface reactions whereby clay was expelled from siltstone nuclei to form a spherical rind, in which iron compounds were perhaps oxidized. Exhumation of certain spherules may then have taken place, during redeposition, with subsequent fracture, dispersal, and reburial. Oxidation of the rind could also have taken place during this disinterment. The physicochemical motive behind the segregation of clay is obscure; this does not, however, discount it as a viable mechanism since chamosite ooliths apparently form in a roughly comparable way (e.g., Schellmann, 1969).

An alternative possibility is that the spherules represent lapilli whose outer surfaces have reacted rapidly with sea water to form a clay rind. Although this is conceivable, it leaves a lot to chance to find such a lapillus "dropping in" to the finest fraction of a graded bed. On balance, therefore, I favor some kind of early diagenetic segregation mechanism for the formation of the spherules.

Associated with the weathering or hydrothermal alteration of these volcaniclastic sediments has been the formation and precipitation of calcite and analcite in available void space. Analcite has also been recorded by Melson and Thompson (1973) from the altered basalts of St. Paul's Rocks and by McKelvey and Fleet (1974) from Eocene pyroclastics on the Ninetyeast Ridge, Indian Ocean. Laboratory experiments on the formation of this zeolite from volcanic ashes show that it forms in neutral to alkaline solutions (Abe and Aoki, 1973; Höller et al., 1974). The temperatures at which this transformation takes place usually lie between 100° and 250°C (Höller et al., 1974). Flushing of heated sea water through volcanic sediments would thus easily account for the presence of analcite. This mineral could presumably also form at more modest temperatures given enough time. Calcite is often associated with zeolites and apparently may also be involved in the reaction of heated sea water with volcanic glass (Nayudu, 1964; Bonatti, 1966). The carbonate may derive from the interaction of volcanogenic carbon dioxide with various igneous minerals or perhaps from dissolution of calcareous nannofossils and subsequent precipitation as a cement; the fact that calcium carbonate is less soluble in hot water than in cold may also be significant in this context. The textures of the calcite—several large subhedral crystals in one void—might suggest that neomorphic processes had overprinted, perhaps several times, a sparry or micritic precursor (Bathurst, 1971, p. 484-503).

Data from the Reykjanes geothermal system, Iceland, are relevant in this connection, not only because the Manihiki Plateau was apparently a ridge-associated melting spot, but because the brines in this Recent system are apparently derived from sea water (Björnsson et al., 1972; Tómasson and Kristmannsdottir, 1972). These brines are particularly rich in carbon dioxide. The rocks through which the brines circulate are chiefly hyaloclastites, breccias, and some lava flows, and alteration products include calcite, epidote, zeolites (analcite), and brown- and green-colored montmorillonite. Montmorillonite is the dominant clay mineral up to temperatures of 200°C; at higher temperatures chlorite is developed. Epidote is also only formed at temperatures greater than 200°C. Applied to the Manihiki volcaniclastic sediments this would suggest that the sea water solutions that were flushed through were at temperatures less than 200°C: this is consistent with the experimental data on the formation of analcite.

What is a trifle problematic is that geothermal brines, where analyzed, are usually acid, with pH<6. Thus, on the one hand, there is the suggestion that the analcite, and the K-rich smectite, were formed by relatively alkaline solutions, whereas Recent hot brines are generally acid. The acid nature of these brines is due to dissolved CO_2, H_2S, HCl, and HF, gases which are probably volcanically derived. It seems probable, therefore, that the analcite was formed very early—as its presence as detrital grains would suggest—by simple interaction of glass with locally heated sea water, and that the geothermal system proper was not established within the sedimentary pile until later.

The origin of native copper is presumably to be sought in "hydrothermal solutions." The Red Sea geothermal brine deposits, for example, contain up to 6.5% copper, held as chalcopyrite (Bischoff, 1969). Native copper has been recorded in sedimentary sections at other Deep Sea Drilling sites (Leg 11, Site 105, some 20 m above basalt, Hollister, Ewing, et al., 1972), and, clearly, cupriferous fluids can be expelled high into the sediment column before precipitation takes place. Native copper is common in basalts (e.g., Cornwall, 1956; Nagle et al., 1973), often associated with calcite and zeolites, and is usually assumed to derive from the igneous rocks themselves via a leaching process. Copper averages between some 60 and 120 ppm in deep-sea basaltic rocks (e.g., Nicholls and Islam, 1971). Thus, the hydrothermal solutions that deposited the copper in the Manihiki Plateau sediments presumably also formed by the reaction between thermally driven sea water and fractured volcanic rocks. The weight of evidence now seems to favor this as the most likely mechanism for producing ore-forming fluids (e.g., Bischoff, 1969; Corliss, 1971; Ferguson and Lambert, 1972; Dymond et al., 1973; Natland, 1973; Spooner and Fyfe, 1973), despite the dissenting voice of Boström (1973).

The copper was probably carried as a chloride complex as is postulated for ore-forming solutions of this type (Helgeson, 1964; Corliss, 1971). Experiments on hydrothermal deposition of copper from chloride solutions thus have some relevance; the data of Wells (1925), Park (1931), and Page (1938) are significant. According to Wells and Page, at pH>5, native copper and ferric oxide precipitate together from boiling solutions of cuprous chloride, ferrous chloride, and sodium chloride. Straightforward cooling of sodium and calcium chloride solutions containing dissolved copper, above a critical temperature of 100°C, will also precipitate the metal (Page, 1933). Park (1931) precipitated some native copper from cuprous chloride in contact with calcite and mixtures of hematite and calcite.

Thus, introduction of acid solutions containing dissolved copper and iron would, on reaction with calcites and zeolites and consequent rise in pH, precipitate ferric oxide-hydroxides and native copper. Simple cooling of the ore-bearing solutions would probably accomplish the same end. The absence of significant sulphide phases in the deposit suggests that at these levels in the sediment the solutions were depleted in the requisite anions. The occurrence of chalcopyrite in certain of the basalts (Jackson et al., this volume) shows these anions were present at some stage, but were presumably precipitated out in the igneous basement before the ore-bearing solutions rose through the sedimentary column.

The reddish-brown and purple clay seams that traverse the greenish-black volcaniclastic sediments may be relevant to the emplacement of native copper, bearing in mind the experiments of Wells and Page. Clearly, these veins indicate former zones of oxidation where fluids have dissolved out the volcanic grains in the host rock and left a residue of homogeneous clay and iron oxide-hydroxides. It is possible that these seams are the record of the conduits along which cupriferous solutions travelled while ferruginous compounds were continually jettisoned. In 317A-29-1 volcaniclastic layers locally disrupt the red clay seams, suggesting that ore-forming fluids may have debouched on the sea floor at certain times.

OOZES, CHALKS, LIMESTONES, CLAYSTONES, AND CHERTS (UNIT 2)

The transition between the volcaniclastic sediments and more calcareous deposits is gradual, being registered by a lightening of the greenish-black color that characterizes the lower unit. Above the youngest volcaniclastic horizon biogenous sediments come into their own. Chert is common and is developed in sediments whose age ranges from upper Aptian to middle Oligocene. More clay-rich, darker-colored levels occur typically in the Albian to Campanian. Bivalves are common in the lower part of the unit.

Colors of these sediments vary from whites, if completely calcareous, through pinks and grays to the yellowish-brown and blacks of the claystones. Cherts are generally darker in color than their enclosing matrix.

Structures

Mechanical sedimentary structures are not well developed in these biogenous deposits. At the base of the unit there are traces of hard- and soft-sediment deformation, but this dies out up-section. Some parallel-laminated limestones occur locally, also near the base of the unit (Figure 11); these laminations are manifested by alternate foraminifer-rich and foraminifer-poor layers.

Biogenic structures are, however, well developed, particularly in lower levels of the unit where they are conspicuous in clay-rich sediments. The dominant forms, the so-called "fucoids" of European authors, may be attributable to *Planolites* (Simpson, 1970; Warme et al., 1973); they are generally developed with dark burrow fills in a lighter matrix, more rarely vice versa. These burrows are variably compacted. *Zoophycos*, showing a characteristic chevron pattern in section, and oriented subhorizontally (cf. van der Lingen, 1973; Warme et al., 1973), occurs locally in basal (Aptian) levels.

Clay-rich pressure-solution veins are present locally and may cross-cut the burrows.

Ferromanganese-coated volcanic clasts occur in Sample 317A-7-1, 61-71 cm (Figure 12).

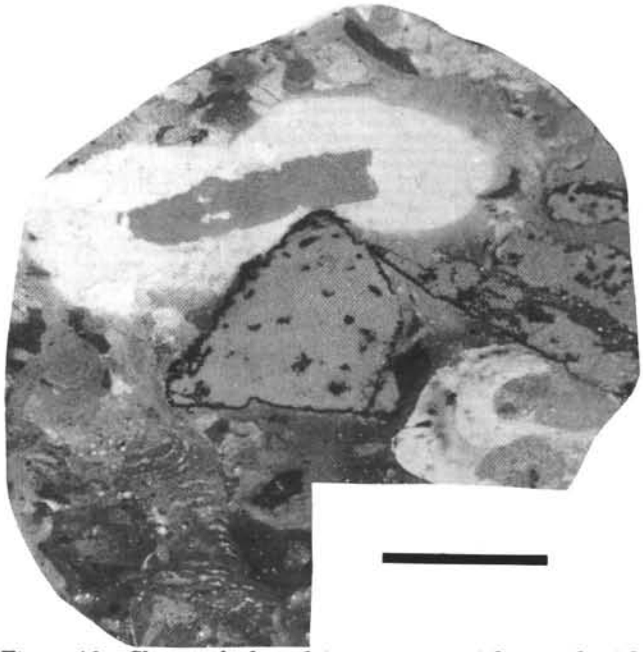

Figure 12. *Clasts of altered igneous material coated with iron-manganese oxide hydroxides in an irregularly colored gray marly matrix. This sample occurs where there is a paleontological gap in the section. Note the vertical burrow to the left of the largest clast; this may be attributable to the Ichnogenus* Teichichnus *(cf.* Warme et al., *1973). Polished face from 317A-7-1, 61-68 cm. Lower Turonian to Santonian. Scale bar = 1.5 cm.*

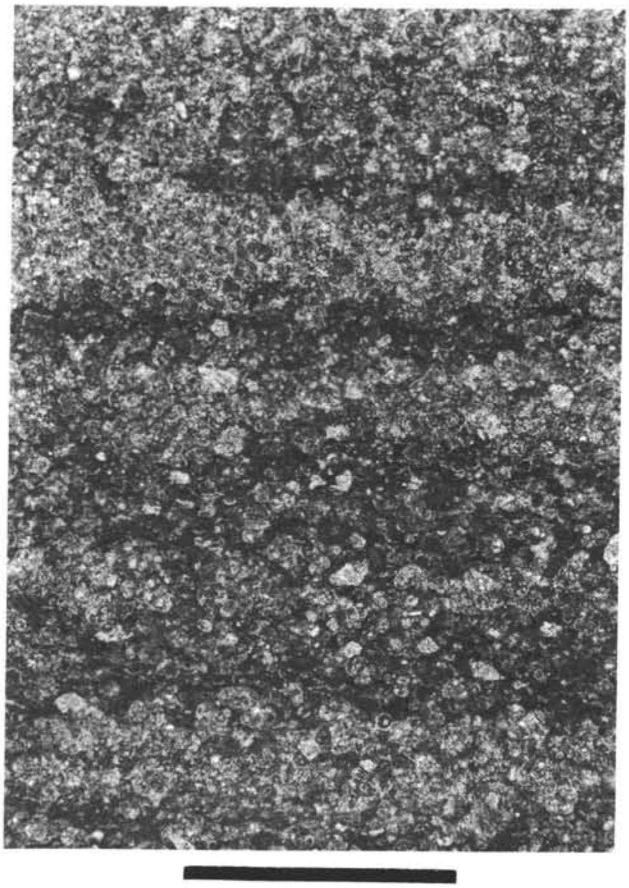

Figure 11. *Laminated foraminiferal biomicrite and biosparite. Foraminiferal chambers are filled with sparite and a little micrite. Inter-foraminiferal voids are also filled with sparite in the lighter colored, presumably winnowed layers. Thin section from 317A-11-2, 32-er cm. Aptian. Scale bar = 1 mm.*

Textures

The bulk of the sediment in this unit is best described as biomicrite with variable amounts of clay which may in extreme cases occur to the virtual exclusion of calcareous matter. Nannofossils are the prime purveyors of the micrite matrix, and foraminifers are common throughout. Radiolarians are present locally, as are sponge spicules. Thin-shelled bivalves (Figure 13) and *Inoceramus* (Figures 14 and 15) are widespread in the lower part of the section (Kauffman, this volume). Some bivalves are recrystallized, others exhibit cross-foliated structure; the shells are commonly broken. In 317A-11-1, 98-103 cm iron-manganese material encrusts certain shell fragments, either on one side only or on both sides.

Evidence of precipitation of calcite can be seen in the overgrowths on discoasters, in the sparry fills of some foraminifer chambers (Figure 16), and in the interparticle cement of winnowed foraminifer sand layers (Figure 17). The latter effect is particularly striking in that the appearance of the fills is relatively sudden. In the upper levels of the unit down to a subbottom depth of 575 meters—given that part of the section was not cored—the foraminifer chambers are empty or contain a trace of micrite. At a subbottom depth of 615 meters foraminifer tests, and certain molds of bivalved shells, are filled either with sparry calcite or with mixtures of micrite and sparry calcite. The sparry calcite crystals, occasionally euhedral, may show an increase in size from the test wall inwards; more commonly the crystals

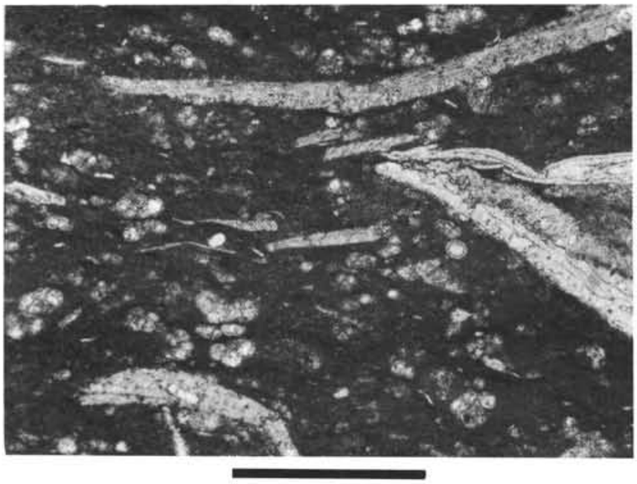

Figure 13. *Clay-rich molluscan biosparite, containing abundant broken thin-shelled bivalves and spar-filled foraminifers. Most bivalves retain their shell structure. Thin section from 317A-9-2, 114-117 cm. Lower to middle Albian. Scale bar = 0.5 mm.*

are large relative to the microfossil, and one crystal may entirely fill a foraminifer chamber. Small amounts of neomorphic spar replacing micrite occur locally around calcite-filled tests.

The dark yellowish-brown claystones that occur intercalated between the nannofossil chalks contain abundant fish debris (Figure 18) in which some teeth are recognizable. The phosphatic material may be concentrated in local "clots."

Textures become more variable when silicification—to whatever degree—has taken place. In early stages of silicification only the voids of foraminifer tests are filled with silica, either a salt-and-pepper arrangement of quartz blebs, or radiating fibers of length-fast chalcedony. Chamber walls of foraminifers generally remain as calcite; rarely they have been cannibalized by silica. Some foraminifer shells are filled with sparry calcite and are surrounded by fine-grained silica. Silica-filled, probably radiolarian, spheres occur at some levels; sponge spicules are abundant in 317A-7, CC (Cenomanian-upper Albian), but are otherwise rare; they are recrystallized but some still exhibit the central canal. Lenses of silica unrelated to the presence of microfossils are uncommon, but do occur around some bivalves (Figures 14 and 15). Where chertification has progressed to a considerable degree, calcite-walled foraminifers may be discernible in a matrix of fine-grained silica; in extreme cases only siliceous "ghosts" of former microfossils are visible in a completely chertified matrix of fine-grained silica.

Mineralogy

The most common mineral present in this unit is clearly calcite. Plagioclase, potash feldspars, and micas are tolerably common (Cook and Zemmels, this volume). Montmorillonite is present in variable amounts. Small quantities of barite occur locally. Clinoptilolite is locally abundant. High concentrations of palygorskite occur in a dark yellowish-brown claystone of Campanian age (317-6-2). These brownish-black claystones also contain apatite attributable to fish debris, quartz, potash feldspars, clinoptilolite, and montmorillonite; calcite is absent.

The silica phases vary. Some cherts are composed entirely of quartz; others contain additionally a mineral which gives peaks attributable to both cristobalite and tridymite and is probably best described as lussatite (e.g., Jones and Segnit, 1971; von Rad and Rösch, 1974). The relationships between silica phases and matrix are shown in Table 4.

Depositional Processes

As volcanic activity waned during the late Early Cretaceous, sedimentation on the Manihiki Plateau changed to dominantly biogenous in nature. The change in bivalve faunas, from those of "shallow-shelf" aspect in the volcaniclastics to those of deeper water character in the immediately overlying carbonates, suggests rapid sinking of the edifice at this same time (Kauffman, this volume). This is in accord with the subsidence of ocean ridges following the model of Sclater et al. (1971). Nannofossils, foraminifers, radiolarians, and to a lesser extent sponge spicules were the most important sedimentary contributors. Locally, hydraulic sorting produced lenses of foraminifer sand. The clay-rich sediments that occur in 317A, Cores 5 to 8 (Albian-Campanian) presumably indicate a decrease in the amount of nannofossil carbonate entering the burial stage or an increase in the supply of clay. The scarcity of foraminifers at these levels, known to be an index of solution (Berger, 1971; Berger and von Rad, 1972), might suggest that a shallow compensation level strongly influenced these sediments. Such a shallow compensation depth might have been caused by either a rise in the lysocline (Berger, 1970) or by decreased nannoplankton production in surface waters of low fertility. The northward passage of the Pacific plate with time necessitates that the sediment supply at any one spot will vary as differing zones of organic productivity are crossed (cf. Winterer, 1973; Berger and Winterer, 1974).

These clay-rich sediments apparently provided a suitable substrate for burrowing organisms so that they could mix darker clay-rich and lighter lime-rich sediments together to form typical "fucoids."

The presence of palygorskite in 317A-6-2 (Santonian) is notable. The origin of this mineral has been discussed by Bonatti and Joensuu (1968), Bowles et al. (1971), and von Rad and Rösch (1972). It is generally assumed to derive from the interaction of magnesium-rich fluids and excess silica with montmorillonite-group clays. Submarine volcanism and hydrothermal solutions are frequently invoked to explain the genesis of this mineral. In this part of the Manihiki section, however, there is no trace of the passage of hydrothermal solutions, and the only evidence of volcanism is the small amount of montmorillonite. The presence of abundant fish remains and the absence of calcium carbonate, and the relative condensation of this part of the section, suggest slow rates of deposition and it seems likely, following von Rad and

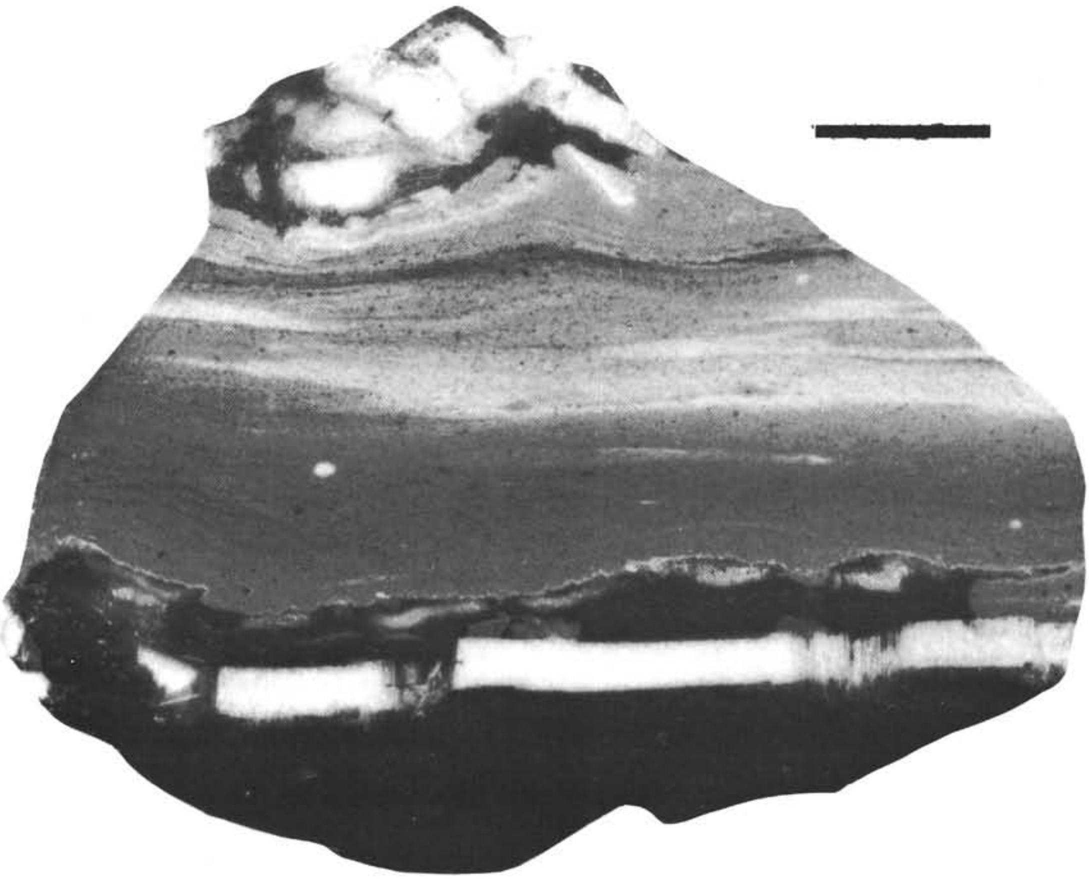

Figure 14. *Partially silicified claystone showing laminated "fluidal" structure and containing black chert in which unreplaced* Inoceramus *prisms occur. Polished face from 317A-8-1, 136-140 cm. Upper Albian to Cenomanian. Scale bar = 1 cm.*

Rösch (1972), that an essentially nonaccumulational environment favored the formation of palygorskite. Thus, its origin was presumably related to slow alteration of montmorillonite with excess silica supplied by subsurface solution of opaline tests and magnesium supplied by sea water. Thus, reaction may have been governed by the chemical gradients that pertained at the sediment-water interface.

The pinkish clasts of volcanic material coated with ferromanganese rinds in 317A-7-1, 61-71 cm (Figure 12) are significant in that they coincide with a paleontological gap between the lower Turonian and Santonian. In the pelagic sedimentary record on land iron-manganese nodules are frequently associated with condensed sequences that bear witness to considerable submarine erosion (Jenkyns, 1970; Wendt, 1970). Thus the encrusted material in the Manihiki section presumably also formed during the nondepositional episode.

The presence of cherts in sediments ranging in age from upper Aptian to middle Oligocene shows that the Manihiki Plateau lay beneath waters favorable for siliceous plankton during this time: this is presumably linked to a location close to equatorial zones (e.g., Ramsay, 1973; Winterer, 1973).

Diagenetic Processes

Diagenetic processes operating within this dominantly calcareous sedimentary pile include precipitation of calcite cement to produce lithified limestones at certain horizons, and the growth of replacive chert.

Large nannofossils such as discoasters are frequently overcalcified, a phenomenon described by many workers (e.g., Schlanger et al. 1973; Matter, 1974; Schlanger and Douglas, 1974), and it is apparent that the well-established pattern of dissolution of foraminifer crystallites and concomitant precipitation of calcite on large nannofossils has taken place. The most dramatic manifestation of the production of cement is the sparry calcite fills in and around foraminifer tests (Figures 16 and 17; see also Matter, 1974) which appear at subbottom depths of 615 meters.

The production of cements in chalks has been considered by Neugebauer (1973, 1974), who has stressed the role of the magnesium ion. With abundant magnesium of sea-water origin in the pore fluids, pressure solution-precipitation is inhibited since interstitial fluids are already supersaturated with respect to the solid phase. Following Neugebauer (1974), only after overloads of more than 300 meters does the amount of cement released become significant. Thus the relatively sudden appearance of spar-filled chambers at a subbottom depth of 615 meters, when they are absent at 575 meters, could be pointing to a critical depth of about 600 meters for abundant pressure solution-precipitation. This theory, however, does not account for the known departure from the ideal of a direct depth-of-burial

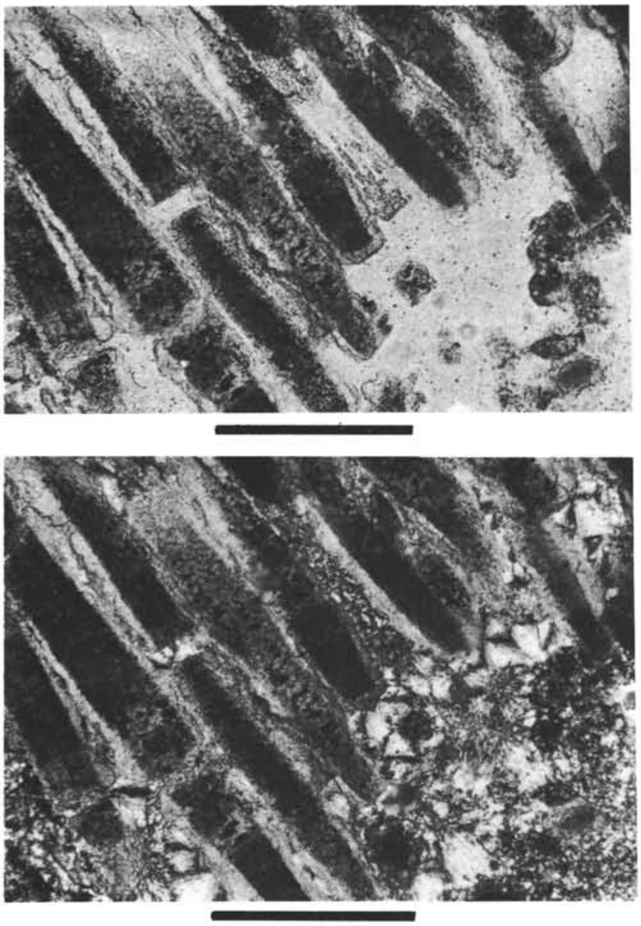

Figure 15. Inoceramus *prisms partly enveloped and invaded by silica. Above: plane polarized light. Below: crossed nicols. Under crossed nicols silica exhibits salt-and-pepper structure with the development of some length-fast chalcedony. Thin section from 317A-8-1, 134-137 cm. Upper Albian-Cenomanian. Scale bar = 0.5 mm.*

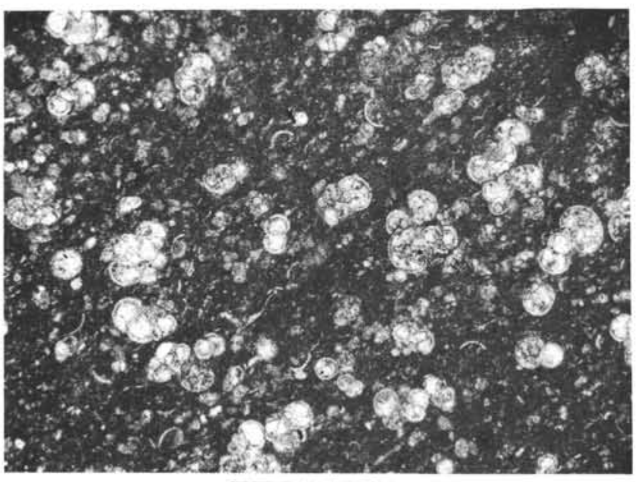

Figure 16. *Foraminiferal biomicrite. Chambers are generally filled with sparry calcite, which may be only one or two crystals. Neomorphic spar occurs immediately outside some foraminiferal tests. Thin section from 317A-11-1, 98-103 cm. Aptian. Scale bar = 0.5 mm.*

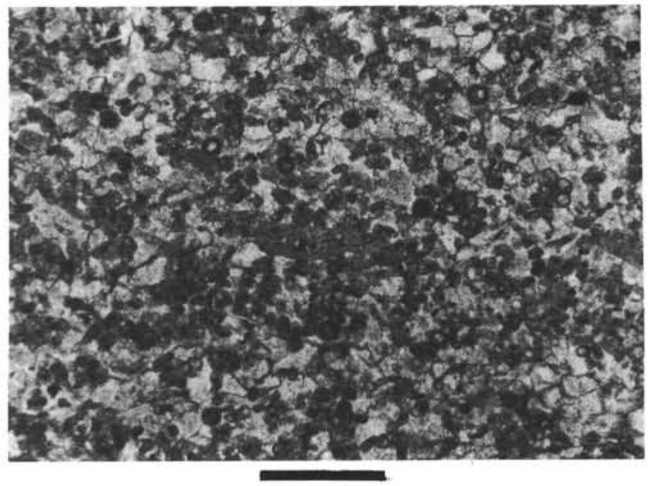

Figure 17. *Winnowed foraminiferal sand, partly cemented by void-filling sparry calcite. Thin-shelled bivalve fragment is visible in center of picture. Thin section from 317A-9-1, 50-58 cm. Lower-middle Albian. Scale bar = 0.5 mm.*

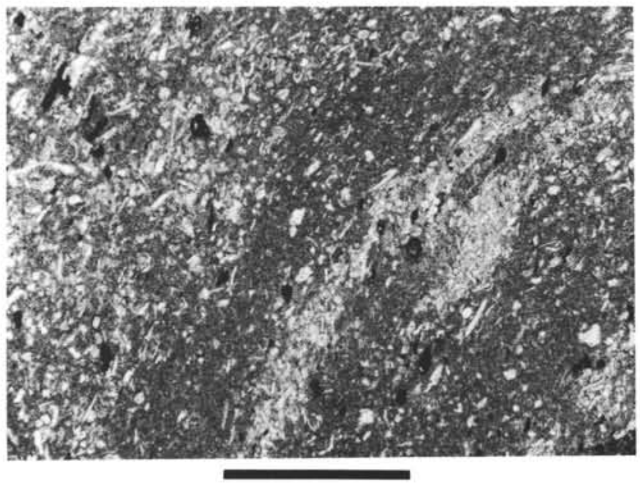

Figure 18. *Claystone containing abundant fragmental fish remains. Some of the phosphatic material occurs in "clots" and may have a fecal origin. Section from which this claystone derives is stratigraphically condensed. Thin section from 317A-6-2, 67-69 cm. Campanian. Scale bar = 1 mm.*

hardness dependence manifested by the Manihiki and other DSDP sites. To resolve this Schlanger and Douglas (1974) have suggested that some ancient oozes contain more diagenetically soluble particles than others, and the former possessed therefore an enhanced capability to form chalk or limestone. Of these two theories, that of Schlanger and Douglas perhaps finds most support from the Manihiki section in that this highest level of widespread carbonate cementation corresponds with the top of the zone of abundant bivalves (Figure 2). It seems likely, therefore, that below a sub-bottom depth of around 615 meters, extending down into the volcaniclastic sediments, molluscan aragonite entered the burial stage and was available, via solution-

TABLE 4
Silica Phases Found in Differing Matrix Rocks in Holes 317A and 317B

Subbottom Depth (m)	Matrix	Silica Phases
406.9	Light olive-gray chalk	Quartz Lussatite
579.6	Dark gray clayey nannofossil chalk	Quartz
585.0	Dusky yellowish-brown nannofossil claystone	Quartz
595.0	Dark yellowish-brown claystone	Lussatite Quartz
602.4	Dark yellowish-brown claystone	Lussatite Quartz
621.6	Olive-gray and pinkish-gray chalk	Quartz
623.3	Pale gray clayey limestone	Lussatite Quartz

precipitation, for diagenetic production of calcite cement.

The mechanics of chert formation in deep-sea calcareous oozes are now known in some detail (e.g., Heath and Moberly, 1971; Wise and Kelts, 1972; Heath, 1973; Lancelot, 1973; von Rad and Rösch, 1974; and Wise and Weaver, 1974). There does, however, remain a controversy over the controls on the silica phases precipitated. According to von Rad and Rösch (1974) and Wise and Weaver (1974), dissolution of biogenic opaline silica followed by precipitation of lussatite lepispheres in available void space is the first step; expulsion of host-rock carbonate then takes place. The lepispheres gradually accrete and finally invert to quartz. Mineralogy is assumed to be primarily a function of age and temperature gradients. However, according to Lancelot (1973) and Greenwood (1973), lussatite is favored in clay-rich sediments, whereas quartz is a primary precipitate in calcareous sediments. The mineralogical data on cherts from the Manihiki section, albeit limited, at least indicate that age is not the only, or indeed the most important, factor governing mineralogy (Table 4). Indeed, the oldest chert analyzed (cored at a subbottom depth of 623.3 m) contains dominantly lussatite with subordinate quartz. Cherts at 597.6 meters and 585.0 meters contain only quartz. Thus it seems that, as Lancelot and Greenwood suggest, host sediments must dictate to some extent the mineralogy of precipitated silica phases. The data from the Manihiki section suggest that where abundant lussatite occurs the matrix is clay rich; this in no way precludes the formation of quartz in clay-rich sediments (as at the subbottom depth of 585 m). However, lussatite can clearly form in fairly pure chalks (e.g., Wise and Kelts, 1972; Wise and Weaver, 1974), so the role and extent of sediment control is complex.

OOZES AND CHALKS (UNIT 1)

Above the youngest chert in the Manihiki section, the sediments are dominantly calcareous with carbonate contents greater than 90%. The most striking feature of these sediments is the alternation in color from pale orange to bluish-white and white.

Structures

The only structures present in this part of the section are those manifested by differentially colored zones of centimeter scale; such mottling was presumably produced by burrowing organisms. These mottles extend from surface cores down to the base of the unit.

Textures

These sediments consist of nannofossil-foraminifer ooze and foraminifer-nannofossil ooze to chalk (Figure 19). The chambers of the foraminifers are empty or contain a trace of micrite (Figure 20). Some white streaks in darker grayish-orange sediment are pure nannofossil ooze. These deposits can be referred to as foraminiferal biomicrites. The fauna and flora includes, in addition to foraminifers and nannofossils, radiolarians and scarce sponge spicules with very sparse silicoflagellates. Diatoms have not been found; nor have pteropods. Clay constitutes the only other significant sediment contributor.

Chalk is present below a subbottom depth of 149 meters (middle Miocene) and is typically present as "biscuits" produced by drilling disturbance; above this the sediments comprise firm ooze and ooze. Below a subbottom depth of 45-50 meters, overgrowths on discoasters are intermittently recognizable with the petrographic microscope; these overgrowths are generally more pronounced in whiter than in grayish-orange sediments. Solution traces on foraminifers are, however, more obvious in the orange sediments.

Mineralogy

The upper sedimentary unit of the Manihiki Plateau comprises dominantly calcite with traces of other minerals (Cook and Zemmels, this volume). Acid-insoluble fractions contain quartz, potassium feldspar, plagioclase, clay minerals, plus relatively abundant barite, some clinoptilolite, and local amounts of psilomelane and apatite.

Depositional Processes

During the later, Tertiary to Recent history of the Manihiki Plateau abundant low-magnesian calcite was supplied to the bottom sediments by nannofossils and foraminifers. Sedimentation did not, however, continue at the same rate throughout this period but alternately accelerated and decelerated as the sediment accumulation rate curve shows (Figure 21). It is interesting to attempt a correlation between the color changes in the sedimentary section and the inferred accumulation rates (Table 5). Clearly, the times of lower accumulation rate correlate roughly with formation of grayish-orange sediments. This may be explained as follows. Assuming roughly constant input of iron-manganese oxide hydroxides and clays from aeolian and intra-oceanic volcanic sources, reduced deposition of calcareous sediment, due either to reduced primary supply, or solution, will result in a relative increase in pigment-carrying material, thus coloring the sediment grayish-orange. The fact that foraminifers show etching effects in certain

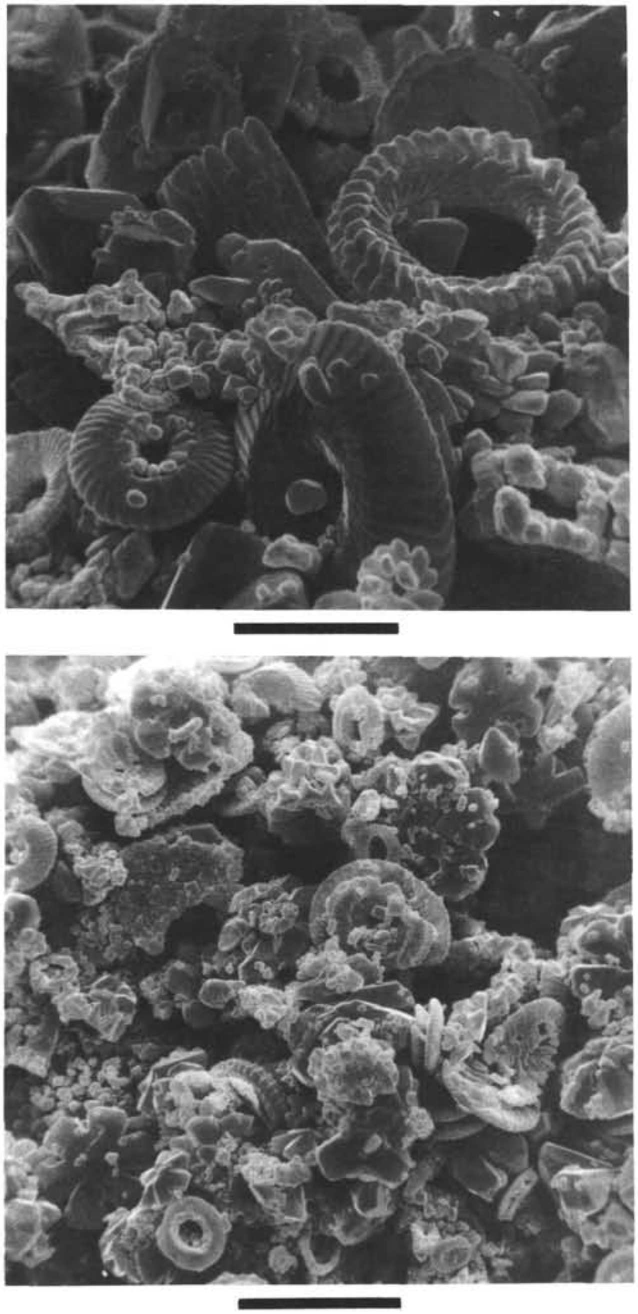

Figure 19. *Scanning electron micrographs of grayish-orange foraminiferal nannofossil firm ooze: from 317B-19-4, 32-34 cm. Coccoliths are well preserved and show some evidence of overgrowth. Euhedral crystals visible in top left of upper photomicrograph may represent void-filling calcite. Lower Miocene. Above: scale bar = 5µm Below: scale bar = 12.5µm.*

of these grayish-orange sediments suggests that solution was an important agent in reducing the amount of sediment that became buried. Also significant in terms of coloration is the influence of bacterial oxidation of organic matter; in areas of slow net deposition this process takes place on the sea floor and the sediment is not therefore reduced at depth with consequent loss of color.

The presence of orange-colored sediments at the surface of the Manihiki Plateau today suggests that depositional rates are now slow; some support for this is given by the record of iron-manganese nodules that cover parts of the surface of this edifice (Heezen and Hollister, 1971). Slow rates of deposition could also explain the absence of pteropods in the bottom sediments; the depth of the drill site (2598 m) was well within the present-day accumulation level of these aragonitic molluscs (e.g., Murray and Hjort, 1912; Chen, 1964). However, in a nondepositional environment exposed to carbonate-undersaturated bottom waters pteropods could easily be dissolved.

Certain of the minerals that occur in this upper part of the sedimentary column are probably aeolian; this

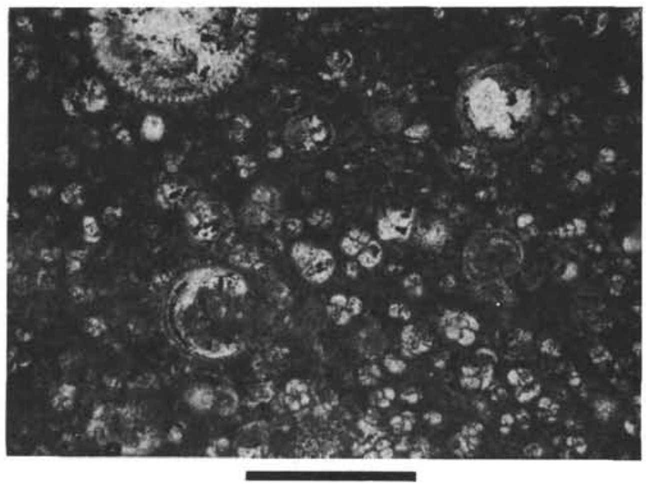

Figure 20. *Nannofossil ooze containing foraminifers whose chambers are either completely empty or contain a trace of micrite. Thin section from 317B-9-1, 18-80 cm. Upper Miocene. Scale bar = 0.4 mm.*

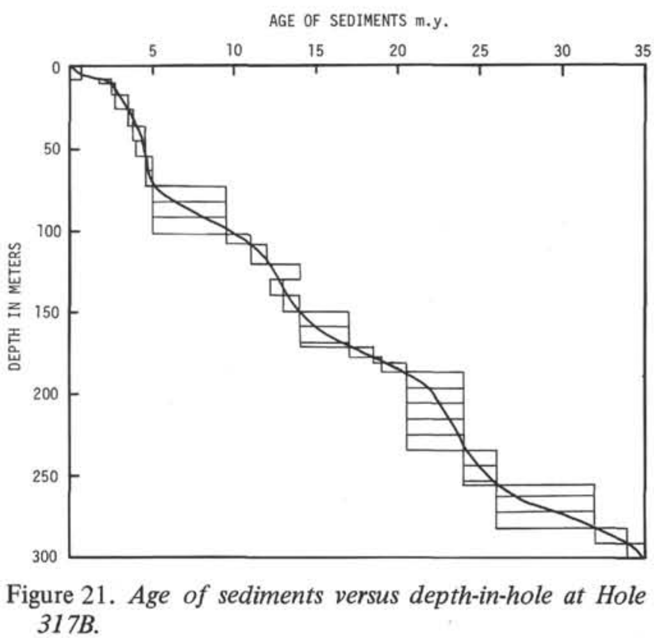

Figure 21. *Age of sediments versus depth-in-hole at Hole 317B.*

TABLE 5
Color of Calcareous Oozes in Hole 317A Versus Subbottom Depth

Depth-in-Hole Location of Pale Orange Ooze to Chalk (m)	Depth-in-Hole Location of White and Bluish-White Ooze to Chalk (m)
0-10	10-114
114-118.5	118.5-139
139-234.5	234.5-258
258 (base of unit)	

origin could account for some of the quartz and feldspar (e.g., Rex and Goldberg, 1958). The gypsum presumably is also wind transported, being derived from oceanic islands where evaporation of sea water has locally taken place. Barite may be supplied by both biogenic and volcanic processes (e.g., Boström et al., 1973). Clinoptilolite is invariably associated with montmorillonite, which is in accordance with its proposed origin from the low-temperature devitrification of volcanic glass under alkaline conditions (e.g., von Rad and Rösch, 1972 and references therein). The apatite is presumably derived from fish remains; the psilomelane may record the presence of a ferromanganese micronodule.

Diagenetic Processes

The major diagenetic processes affecting this pile of sediments have been dissolution, transfer, and precipitation of calcium carbonate; large nannofossils have grown at the expense of smaller nannofossils and crystallites of foraminifers. These processes have been discussed above.

SEDIMENTARY HISTORY

During Early Cretaceous time the Manihiki Plateau was formed by abundant outpourings of basaltic lavas that grew into moderately shallow water. The site was probably near an ancient triple junction. Despite a relatively shallow depth reefs were apparently developed only locally on the plateau, presumably because volcanism was active throughout the early stages of subsidence, depths were generally too great, or waters were initially too cold. The earliest sediments deposited, of possible Barremian or greater age, were volcaniclastic, directly derived from explosive hyaloclastite eruptions and from weathering of fine-grained and glassy volcanic products. Redeposition processes were active, transporting sediments into ponds on the plateau surface. These sediments rapidly reacted with sea water to produce potassium-rich smectitic clays and analcite. Geothermal systems, involving reactions between volcanic gases, heated sea water, and basalt were probably engendered at an early stage during the growth of the volcanic edifice and, as the volcaniclastic sedimentary pile built up, the circulating solutions entered it. The more mobile of these, specifically iron-bearing cupriferous solutions, were expelled high into the sediment column to precipitate as iron oxide-hydroxides and native copper.

Towards the end of Early Cretaceous time the supply of volcanically derived sediment waned; an episode of stagnant deoxygenated bottom-water conditions is recorded by the presence of a sapropelic sandstone. During the Barremian-Aptian bivalves, including *Inoceramus* and various other thick- and thin-shelled forms, colonized the surface of the Manihiki Plateau as it rapidly sank. During Aptian time the calcareous material derived from nannoplankton and foraminifers became increasingly important. Siliceous macrofossils were also delivered to the sediment; this opaline silica was eventually redistributed to form chert. During the Albian and Campanian, however, the calcite compensation depth apparently rose so that dark clay-rich sediments, rich in fish material, were laid down. Burrowing organisms flourished in the sediment during this time and throughout much of the Late Cretaceous. Between the early Turonian and Santonian a non-depositional episode took place—presumably caused by submarine erosion—and ferromanganese-coated clasts were formed.

The later, Tertiary to Recent record of the Manihiki Plateau was dominated by calcareous sedimentation which nevertheless varied somewhat in its rate of accumulation, such that orange-hued oozes were laid down slowly and bluish-white and white oozes relatively rapidly. Depositional rates on the Manihiki Plateau are probably slow at the present time.

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REFERENCES

Abe, H. and Aoki, M., 1973. Synthesis of analcime from volcanic sediments in sodium silicate solution: Contrib. Mineral Petrol., v. 42, p 81-92.

Bathurst, R.G.C., 1971. Carbonate sediments and their diagenesis: Amsterdam (Elsevier).

Berger, W.H., 1970. Planktonic foraminifera: selective solution and the lysocline: Marine Geol., v. 8, p. 111-138.

_______, 1971. Sedimentation of planktonic foraminifera: Marine Geol., v. 11, p. 325-358.

Berger, W.H. and von Rad, U., 1972. Cretaceous and Cenozoic sediments from the Atlantic Ocean. *In* Hayes, D.E., Pimm, a.C., et al., Initial Reports of the Deep Sea Drilling Project, Volume 14: Washington (U.S. Government Printing Office), p. 787-954.

Berger, W.H. and Winterer, E.L., 1974. Plate stratigraphy and the fluctuating carbonate line. *In* Hsü, K.J. and Jenkyns, H.C. (Eds.), Pelagic sediments: on land and under the sea: Int. Assoc. Sedimentol. Spec. Publ. 1, p. 11-48.

Bischoff, J.L., 1969. Red Sea geothermal brine deposits: their mineralogy, chemistry and genesis. *In* Degens, E.T. and Ross, D.A. (Eds.), Hot brines and recent heavy metals in the Red Sea: Berlin (Springer Verlag), p. 368-401.

Björnsson, S., Arnôrsson, S., and Tômasson, J., 1972. Economic evaluation of Reykjanes thermal brine area, Iceland: Am. Assoc. Petrol. Geol. Bull., v. 56, p. 2380-2391.
Bonatti, E., 1966. Deep-sea authigenic calcite and dolomite: Science, v. 153, p. 534-537.
_______, 1967. Mechanisms of deep-sea volcanism in the South Pacific. *In* Abelson, P.H. (Ed.), Researches in geochemistry: 2. New York (John Wiley), p. 453-491.
Bonatti, E. and Joensuu, O., 1968. Palygorskite from Atlantic deep sea sediments: Am. Mineral., v. 53, p. 975-983.
Bortolotti, V., Passerini, P., Sagri, M., and Sestini, G., 1970. The miogeosynclinal sequences. *In* Sestini, G. (Ed.), Development of the Northern Apennines geosyncline: Sediment. Geol., v. 4, p. 341-444.
Boström, K., 1973. The origin and fate of ferromanganoan active ridge sediments: Stockholm Contrib. Geol. v. 27, p. 149-243.
Boström, K., Joensuu, O., Moore, C., Boström, B., Dalziel, M., and Horowitz, A., 1973. Geochemistry of barium in pelagic sediments: Lithos, v. 6, p. 159-174.
Bowles, F.A., Angino, E.E., Hosterman, J.W., and Galle, O.K., 1971. Precipitation of deep-sea palygorskite and sepiolite: Earth Plant. Sci. Lett., v. 11, p. 324-332.
Cann, J.R., 1970. Rb, Sr, Y, Zr and Nb in some ocean floor basaltic rocks: Earth Planet. Sci. Lett., v. 10, p. 7-11.
_______, 1971. Major element variations in ocean-floor basalts: Phil. Trans. Roy. Soc. London, ser. A, v. 268, p. 495-505.
Chen, C., 1964. Pteropod ooze from Bermuda Pedestal: Science, v. 144, p. 60-62.
Corliss, J.B., 1971. The origin of metal-bearing hydrothermal solutions. J. Geophys. Res., v. 76, p. 8128-8138.
Cornwall, H.R., 1956. A summary of ideas on the origin of native copper deposits: Econ. Geol., v. 51, p. 615-631.
Dymond, J., Corliss, J.B., Heath, G.R., Field, C.W., Dasch, E.J., and Veeh, H.H., 1973. Origin of metalliferous sediments from the Pacific Ocean: Geol. Soc. Am. Bull., v. 84, p. 3355-3372.
Edgar, N.T., Saunders, J.B., et al., 1973. Initial Reports of the Deep Sea Drilling Project, Volume 15: Washington (U.S. Government Printing Office).
Ferguson, J. and Lambert, I.B., 1972. Volcanic exhalations and metal enrichments at Matupi Harbour, New Britain, T.P.N.G.: Econ. Geol., v. 67, p. 25-37.
Greenwood, R., 1973. Cristobalite: its relationship to chert formation in selected samples from the Deep Sea Drilling Project: J. Sediment. Petrol., v. 43, p. 700-708.
Hart, R.A., 1970. Chemical exchange between sea water and deep ocean basalts: Earth Plant. Sci. Lett., v. 9, p. 269-279.
_______, 1973. A model for chemical exchange in the basalt-seawater system of oceanic layer II: Canadian J. Earth Sci., v. 10, p. 799-816.
Hayes, D.E., Pimm, A.C., et al., 1972. Initial Reports of the Deep Sea Drilling Project, Volume 14: Washington (U.S. Government Printing Office).
Heath, G.R., 1973. Cherts from the Eastern Pacific. *In* van Andel, T.H., Heath, G.R., et al., Initial Reports of the Deep Sea Drilling Project, Volume 16: Washington (U.S. Government Printing Office), p. 609-613.
Heath, G.R. and Moberly, R., Jr., 1971. Cherts from the western Pacific Leg 7, Deep Sea Drilling Project. *In* Winterer, E.L., Riedel, W.R., et al., Initial Reports of the Deep Sea Drilling Project, Volume 7: Washington (U.S. Government Printing Office), p. 991-1007.
Heezen, B.C. and Hollister, C.D., 1971. The face of the deep: New York (Oxford University Press).
Heezen, B.C., Glass, B., and Menard, H.W., 1966. The Manihiki Plateau: Deep-Sea Res., v. 13, p. 445-458.
Helgeson, H.C., 1964. Complexing and hydrothermal ore deposition: New York (Pergamon Press).
Holland, J.G. and Brindle, D.W., 1966. A self-consistent mass absorption correction for silicate analysis by X-ray fluorescence: Spectrochim. Acta, v. 22, p. 2083-2093.
Höller, H., Wirsching, U., and Fakhuri, M., 1974. Experimente zur Zeolithbildung durch hydrothermale Umwandlung: Contrib. Mineral Petrol., v. 46, p. 49-60.
Hollister, C.D., Ewing, J.I., et al., 1972. Initial Reports of the Deep Sea Drilling Project, Volume 11: Washington (U.S. Government Printing Office).
Jenkyns, H.C., 1970. Fossil manganese nodules from the west Sicilian Jurassic: Eclog. Geol. Helv., v. 63, p. 741-774.
Jones, J.B. and Segnit, E.R., 1971. The nature of opal. I. Nomenclature and constituent phases: J. Geol. Soc. Australia, v. 18, p. 57-68.
Lancelot, Y., 1973. Chert and silica diagenesis in sediments from the Central Pacific. *In* Winterer, E.L., Ewing, J.I., et al., Initial Reports of the Deep Sea Drilling Project, Volume 17: Washington (U.S. Government Printing Office), p. 377-506.
Matter, A., 1974. Burial diagenesis of pelitic and carbonate deep-sea sediments from the Arabian Sea. *In* Whitmarsh, R.B., Ross, D.A., et al., Initial Reports of the Deep Sea Drilling Project, Volume 23: Washington (U.S. Government Printing Office), p. 421-469.
Matthews, D.H., 1971. Altered basalts from Swallow Bank, an abyssal hill in the N.E. Atlantic, and from a nearby seamount: Phil. Trans. Roy. Soc. London, ser. A, v. 268, p. 551-571.
McBirney, A.E., 1971. Oceanic volcanism: a review: Rev. Geophys. Space Phys., v. 9, p. 523-556.
McKelvey, B.C. and Fleet, A.J., 1974. Eocene basaltic pyroclastics at Site 253, Ninetyeast Ridge. *In* Davies, T.A., Luyendyk, B.P., et al., Initial Reports of the Deep Sea Drilling Project, Volume 26: Washington (U.S. Government Printing Office), p. 553-565.
Melson, W.G. and Thompson, G., 1973. Glassy abyssal basalts, Atlantic sea floor near St. Paul's Rocks: petrography and composition of secondary clay minerals: Geol. Soc. Am. Bull., v. 84, p. 703-716.
Murray, J. and Hjort, J., 1912. The depths of the ocean: London (Macmillan).
Nagle, F., Fink, L.K., Boström, K., and Stipp, J.J., 1973. Copper in pillow basalts from La Desirade, Lesser Antilles Island Arc: Earth Planet. Sci. Lett., v. 19, p. 193-197.
Natland, J.H., 1973. Basal ferromanganoan sediments at DSDP Site 183, Aleutian abyssal plain, and Site 192, Meiji Guyot, northwest Pacific, Leg 19. *In* Creager, J.S., Scholl, D.W., et al., Initial Reports of the Deep Sea Drilling Project, Volume 19: Washington (U.S. Government Printing Office), p. 629-640.
Nayudu, Y.R., 1964. Palagonite tuffs and the products of post-eruptive processes: Bull. Volcanol., v. 27, p. 1-20.
Neugebauer, J., 1973. The diagenetic problem of chalk: the role of pressure solution and pore fluid: Neues Jb. Geol. Paläont., Abh., v. 143, p. 223-245.
_______, 1974. Some aspects of cementation in chalk. *In* Hsü, K.J. and Jenkyns, H.C. (Eds.), Pelagic sediments: on land and under the sea: Int. Assoc. Sedimentol. Spec. Publ. 1, p. 149-176.
Nicholls, G.D. and Islam, M.R., 1971. Geochemical investigations of basalts and associated rocks from the ocean floor and their implications: Phil. Trans. Roy. Soc. London, ser. A, v. 268, p. 469-486.
Page, L.R., 1938. The deposition of native copper under hydrothermal conditions: Econ. Geol., v. 33, p. 522-541.
Park, C.F., Jr., 1931. Hydrothermal experiments with copper compounds: Econ. Geol., v. 26, p. 857-883.
Philpotts, J.A., Schnetzler, C.C., and Hart, S.R., 1969. Submarine basalts: some K, Rb, Sr, Ba, rare-earth, H_2O and CO_2 data bearing on their alteration, modification by

plagioclase, and possible source materials: Earth Planet. Sci. Lett., v. 7, p. 293-299.

Ramsay, A.T.S., 1973. A history of organic siliceous sediments in oceans. *In* Hughes, N.F. (Ed.), Organisms and continents through time: Spec. Pap. Palaeontol. 12, p. 199-234.

Renz, O., 1936. Stratigraphische und mikropalaeontologische Untersuchung. der Scaglia (Obere Kreide-Tertiär) im zentralen Apennin: Eclog. Geol. Helv., v. 29, p. 1-149.

Rex, R.W. and Goldberg, E.D., 1958. Quartz contents of pelagic sediments of the Pacific Ocean: Tellus, v. 10, p. 153-159.

Schellmann, W., 1969. Die Bildungsbedingungen sedimentärer Chamosit-und-Hämatit-Eisenerze am Beispiel der Lagerstätte Echte: Neues Jb. Miner., Abh., v. 111, p. 1-31.

Schlanger, S.O., Douglas, R.G., Lancelot, Y., Moore, T.C., Jr., and Roth, P.H., 1973. Fossil preservation and diagenesis of pelagic carbonates from the Magellan Rise, Central North Pacific Ocean. *In* Winterer, E.L. and Ewing, J.I., et al., Initial Reports of the Deep Sea Drilling Project, Volume 17: Washington (U.S. Government Printing Office), p. 407-427.

Schlanger, S.O. and Douglas, R.G., 1974. The pelagic ooze-chalk-limestone transition and its implications for marine stratigraphy. *In* Hsü, K.J. and Jenkyns, H.C. (Eds.), Pelagic sediments: on land and under the sea: Int. Assoc. Sedimentol. Spec. Publ. 1, p. 327-347.

Sclater, J.G., Anderson, R.N., and Bell, M.L., 1971. Elevation of ridges and evolution of the central Eastern Pacific: J. Geophys. Res., v. 76, p. 7888-7915.

Simpson, S., 1970. Notes on *Zoophycos* and *Spirophyton*. *In* Crimes, T.P. and Harper, J.C. (Eds.), Trace fossils, Geol. J. Spec. Issue 3: Liverpool (Seel House Press), p. 505-514.

Spooner, E.T.C. and Fyfe, W.S., 1973. Sub-sea metamorphism, heat and mass transfer: Contrib. Mineral Petrol., v. 42, p. 287-304.

Taylor, S.R., 1964. Abundance of chemical elements in the continental crust: a new table: Geochim. Cosmochim. Acta, v. 28, p. 1273-1285.

Tômasson, J. and Kristmannsdottir, H., 1972. High temperature alteration minerals and thermal brines, Reykjanes, Iceland: Contrib. Mineral Petrol., v. 36, p. 123-134.

van der Lingen, G.J., 1973. Ichnofossils in deep-sea cores from the southwest Pacific. *In* Burns, R.E. and Andrews, J.E., et al., Initial Reports of the Deep Sea Drilling Project, Volume 21: Washington (U.S. Government Printing Office), p. 693-700.

von Rad, U. and Rösch, H., 1972. Mineralogy and origin of clay minerals, silica and authigenic silicates in Leg 14 sediments. *In* Hayes, D.E., Pimm, A.C., et al., Initial Reports of the Deep Sea Drilling Project, Volume 14: Washington (U.S. Government Printing Office), p. 727-751.

———, 1974. Petrography and diagenesis of deep-sea cherts from the central Atlantic. *In* Hsü, K.J. and Jenkyns, H.C. (Eds.), Pelagic sediments: on land and under the sea: Int. Assoc. Sedimentol. Spec. Publ. 1, p. 327-347.

Warme, J.E., Kennedy, W.J., and Schneidermann, N., 1973. Biogenic sedimentary structures (Trace Fossils) in Leg 15 cores. *In* Edgar, N.T. and Saunders, J.B., et al., Initial Reports of the Deep Sea Drilling Project, Volume 15: Washington (U.S. Government Printing Office), p. 813-831.

Wells, R.C., 1925. Chemistry of deposition of native copper: U.S. Geol. Surv. Bull., No. 778.

Wendt, J., 1970. Stratigraphische Kondensation in triadischen und jurassischen Cephalopodenkalken der Tethys: Neues Jb. Geol. Paläont. Mh., p. 433-448.

Winterer, E.L., 1973. Sedimentary facies and plate tectonics of equatorial Pacific: Am. Assoc. Petrol. Geol. Bull., v. 57, p. 265-282.

Winterer, E.L., Lonsdale, P.F., Matthews, J.L. and Rosendahl, B.R., 1974. Structure and acoustic stratigraphy of the Manihiki Plateau: Deep-Sea Res., v. 21, p. 793-814.

Wise, S.W., Jr., and Kelts, K.R., 1972. Inferred diagenetic history of a weakly silicified deep sea chalk: Gulf Coast Assoc. Geol. Soc. Trans., v. 22, p. 172-203.

Wise, S.W., Jr., and Weaver, F.M., 1974. Chertification of oceanic sediments. *In* Hsü, K.J. and Jenkyns, H.C. (Eds.), Pelagic sediments: on land and under the sea: Int. Assoc. Sedimentol. Spec. Publ. 1, p. 301-326.