10. CRETACEOUS VOLCANOGENIC SEDIMENTS OF THE NAURU BASIN, DEEP SEA DRILLING PROJECT LEG 61[1,2]

Ralph Moberly, Hawaii Institute of Geophysics, University of Hawaii, Honolulu, Hawaii
and
Hugh C. Jenkyns, Department of Geology and Mineralogy, University of Oxford, Oxford OX1 3PR, U.K.

ABSTRACT

Sedimentary rocks of Barremian through early Maestrichtian age recovered on Deep Sea Drilling Project Leg 61 had their principal source in the complex of igneous rocks with which they are interlayered in the Nauru Basin. Relict textures and primary sedimentary structures show these Cretaceous sediments to be of hyaloclastic origin, in part reworked and redeposited by slumps and currents. The dominant composition now is smectite, but locally iron, titanium, and manganese oxides, plagioclase, pyroxene, analcime, clinoptilolite, chalcedonic quartz, cristobalite, amphibole, nontronite, celadonite, and pyrite are also present. The mineral assemblages and the geochemistry reflect the original basaltic composition and its subsequent alteration by one or more processes of submarine weathering, authigenesis, hydrothermal circulation, and contact metamorphism.

Hyaloclastitic sandstone, siltstone, and breccia within the sheet flows below 729 meters sub-bottom depth have Barremian fossils, thus establishing the age of the lower, or extrusive, complex of post-ridge-crest volcanism. Similar hyaloclastites between 564 and 729 meters are invaded by hypabyssal sills of the upper igneous complex, and fossil ages of Albian or Cenomanian set an older limit to the age of that second post-ridge-crest episode. Cenomanian to early Campanian sedimentary rocks between 490 and 564 meters have a substantial contribution of clays of submarine-weathered-basalt origin, as well as hydrothermal and pelagic components. The interval of reworked hyaloclastitic siltstone, sandstone, and breccias between 450 and 490 meters is of late Campanian and early Maestrichtian age. These sediments probably formed from glassy basalt that fragmented upon eruption nearby, when sills were being emplaced. In addition to pelagic elements, these Upper Cretaceous volcanogenic sediments include redeposited material of shallow-water origin, apparently derived from the Marshall Islands.

INTRODUCTION

The principal scientific results of Leg 61 are the discovery and unexpected magnitude of two episodes of post-ridge-crest basaltic igneous activity in the Nauru Basin of the western equatorial Pacific. The flows and sills are interlayered with, and lie under, sedimentary rocks of Cretaceous age. One purpose of this paper is to demonstrate that the main components of these sedimentary rocks are of volcanic origin. Hyaloclastites are especially common, as in parts of the section with claystones derived from submarine weathering of basalt or hyaloclastite.

A second purpose is to assemble the stratigraphic evidence for the age of each igneous episode.

The general descriptions of the site, operations, cores, and main scientific results are in the Site Summary (this volume). Information from shore-based work is referred to the appropriate chapter in this volume.

HYALOCLASTITE WITHIN IGNEOUS COMPLEX

The lower 500 meters drilled in the Nauru Basin is mainly a section of dolerite sills and basalt sheet flows. Between igneous layers are hyaloclastite sediments, whose origin and meager fossil content help to constrain interpretations of the origin of the igneous complex. Sediment was recovered in 21 cores within the complex (Table 1). In general, sediment is rare in the section below 729 meters, which was initially thought to be a mixture of sills and pillowed flows, but more likely is a series of basalt sheet flows whose thicker, non-glassy portion represents massive flow interiors, each overlain by packets of thin, glassy sheets representing flow lobes covering the interior (Batiza and others, this volume). Sediment is common in the section of thick and thin sills between 564 and 729 meters, that is to say, the sills now above 729 meters depth intruded hyaloclastite which had its origin in the glassy rinds and slabs of the flows below 729 meters.[3]

Hyaloclastite of Barremian Age

The oldest fossils recovered from the Nauru Basin constitute a radiolarian assemblage assigned to the *Eucrytis tennis* Zone, from hyaloclastite in Section 80-1 of Hole 462A.

The hyaloclastite section lies between layers of fine-grained diabase or coarse-grained basalt, but the actual contacts were not recovered. A total of about 2.4 meters

[1] *Initial Reports of the Deep Sea Drilling Project*, Volume 61.
[2] Hawaii Institute of Geophysics Contribution No. 1122.

[3] Caution is necessary in matching cored intervals to logged intervals, because of incomplete recovery and because of suspected stretching of the logging cable. For example, Hole 462A information in Table 1 was obtained from the sonic-velocity logging of 26 July 1978, using a depth calibration of the first reading at 6234 meters (1048 m in-hole). The log matches the core fairly well in the middle of the igneous section, but near the base the log depths are 4 meters too low (the top of the sediment interval of Cores 79 and 80 almost certainly is at 992 meters, rather than as logged at 996 meters), whereas at the top of the sills the log depths may be 2 meters too high (the sediment over chilled margin in Core 22 probably is at 591 meters, rather than as logged at 589 meters).

Table 1. Hyaloclastite recovered within flows and sills of the Nauru Basin, DSDP Site 462.

Core	Top within Core, Section, and cm Depth	Recovery (m)	Probable Extent of Sediment between Sills and Flows (est. from 7 June and 26 July 1978 logs; see text) (m depth)	Comments[a]
462-63	1-0	0.6	578.2–581.5	Slightly zeolitic
64	1-30	3.0	585.8–590.5	Zeolitic (analcime)
65	1-0	0.2	Not logged	Zeolitic (analcime)
66	1-0	0.2	Not logged	
462A-15	1-0	0.1	Probably caved, as log shows high velocity and low gamma radiation	
21	Sections 1 and 2	Traces		Hardened red-brown sediment adhering to glassy margins of sills
22	2-0	1.1	587.0–589.0	Zeolitic
23	1-0	0.2	595.5–597.5	Zeolitic
23	2-75	0.2	601.5–603.5	Zeolitic
32	1-46	2.2	653.0–681.0	Zeolitic (analcime); porphyroblastic at contact
33	CC	0.2	Same	
35	CC	0.1	Same	
40	1-92	0.8	700.0–705.5	Zeolitic (abundant analcime). Contains Late Aptian nannofossils, scraps of plants, and soft-sediment deformation
41	7-100	1.0	719.0–728.5	Zeolitic; porphyroblastic at contact
42	1-0	2.6	Same	Zeolitic (abundant analcime); some deformed bedding
43	1-0	3.8	Same	Contains Barremian radiolarians, graded beds, and soft-sediment deformation
44	1-0	Trace	Base of same	Hematitic
46	1-0	Trace	Probably caved, as log shows high velocity and low gamma radiation.	Zeolitic; contains Barremian radiolarians.
70	1-0	0.2	921.5–924.5	Zeolitic (abundant analcime)
79	6-0	0.1	996.0–998.5	
80	1-0	2.3	Same	Contains Barremian radiolarians and intraformational breccias.

[a] Sediments are all smectitic claystones, altered from hyaloclastitic fine sand and silt.

of sediment is in these cores: 14 cm in 79-6, 116 cm in 80-1, and 110 cm in 80-2. The sonic-velocity log shows that the sedimentary interval is about 2.9 meters thick. Its top is at about 992 meters depth, according to the coring record.

The rocks are colored mainly in shades of dark gray and brownish black, but include some greenish-gray and grayish-red portions. Although the composition now is dominantly of clay minerals, these minerals have replaced volcanic fragments most commonly of silt and sand size. Parts of the interval are fine breccias. Much of the section is without sedimentary structure, but horizontal and cross-laminae and the breccias (Fig. 1) are evidence of former currents. Probably, eruptions in and near the area not only set up the currents but also provided the fresh glassy basalt that fractured to make the hyaloclastite. The radiolarians that establish the age are size-sorted with fish debris and agglutinated foraminifers within a well-laminated interval lying on an undulose surface. Thus they, like the bulk of the hyaloclastite itself, have been swept from adjacent parts of the sea floor and sorted and redeposited; they do not represent a direct pelagic rain onto the sea floor. Neither, however, are they likely to represent erosion and redeposition of an exposure of old radiolarian-bearing Lower Cretaceous sediment from the side of a seamount, as the carbonate content of the hyaloclastite is nil, and other fossils are deep-water benthic foraminifers and some fish debris, typical of deep-sea pelagic sediments.

In further detail, the original hyaloclastite of Cores 77 and 80 originally ranged in grain size from fine breccia to fine silt (Fig. 2), a range from 0.4 to 0.02 mm being most common. Many of the fine silty sands show a bimodal distribution, the darker grains and the rare plagioclase and pyroxene grains being generally about 0.08 mm in diameter. Some of the darker grains and various others are about 0.15 mm. Both light and dark grains are distinctly blocky and angular. They and their finer matrix have been extensively altered, but grain shapes are evident under crossed polarizers which commonly show microfibrous fringes of aligned clay platelets outlining the original shapes (Fig. 2A, B). The fringes are about 0.005 to 0.015 mm thick.

Some of the larger dark grains in condensed light show opaque dust and the incipient crystallization of laths; these grains originally were tachylite. The lighter grains are palagonitized sideromelane glass. Their present color is mainly golden brown, but ranges to pale olive green or reddish brown.

Elongate grains, strings of light and dark grains of about the same size, and alternation of light and dark grains provide a fair to poor fabric (Fig. 2C) as well as the lamination and bedding (Fig. 1). Of special note are two intervals of breccia (centered near 80-1-50 and 80-2-60), whose subangular to sub-rounded clasts are of

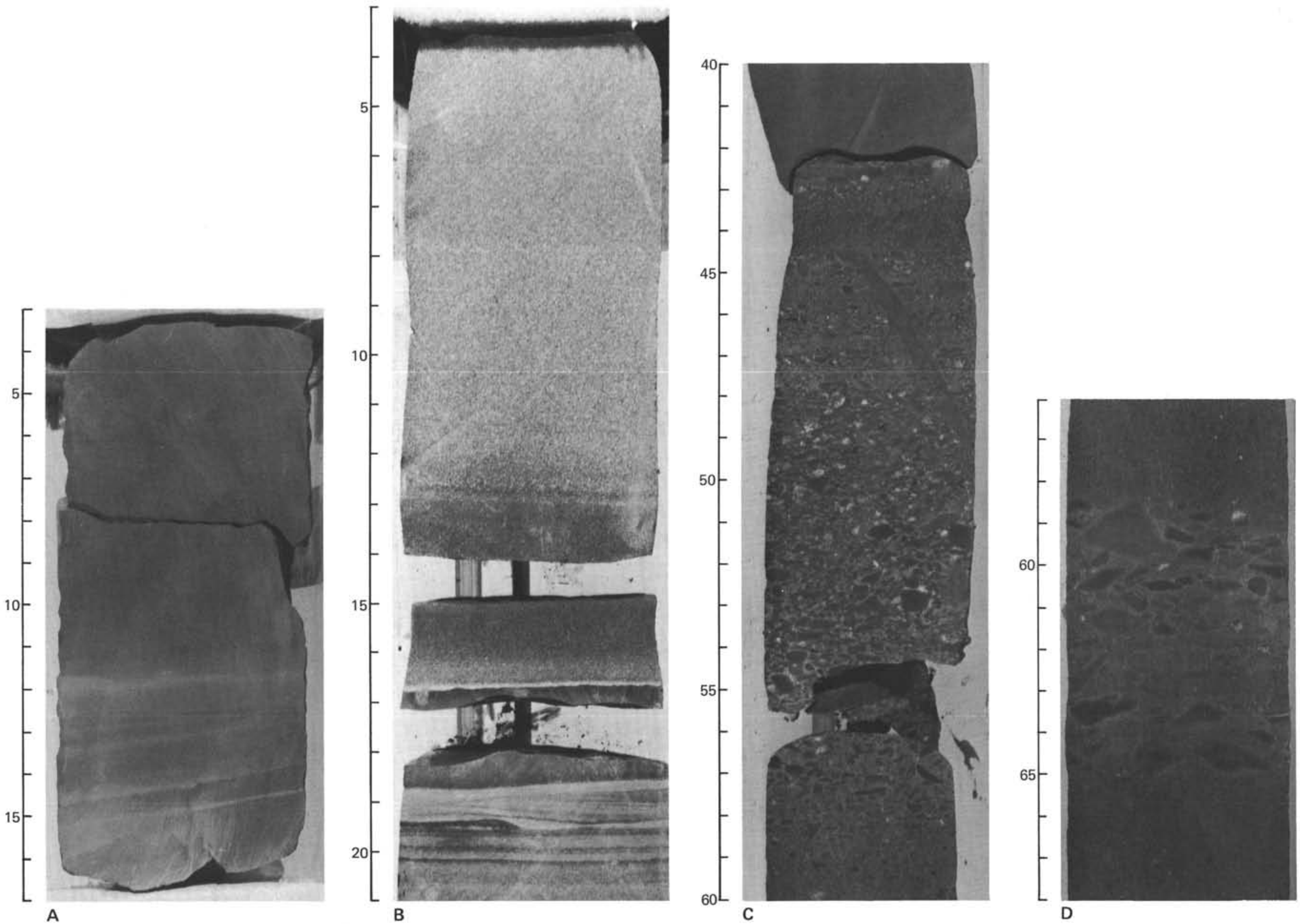

Figure 1. Photographs of split cores of hyalocastites of Barremian age. Section depth (cm) shown at left of each core. A. Thinly laminated sand- and silt-sized hyaclastite near top of sediment interval (462A-79-6). B. Graded, thinly laminated and cross-laminated, sand- and silt-sized hyaloclastite containing radiolarian and benthic-foraminifer fauna (462A-80-1). C. Graded and inversely graded hyaloclastic sandstone and fine breccia in middle of sediment interval (462A-80-1). D. Hyaloclastite breccia near base of interval (462A-80-2).

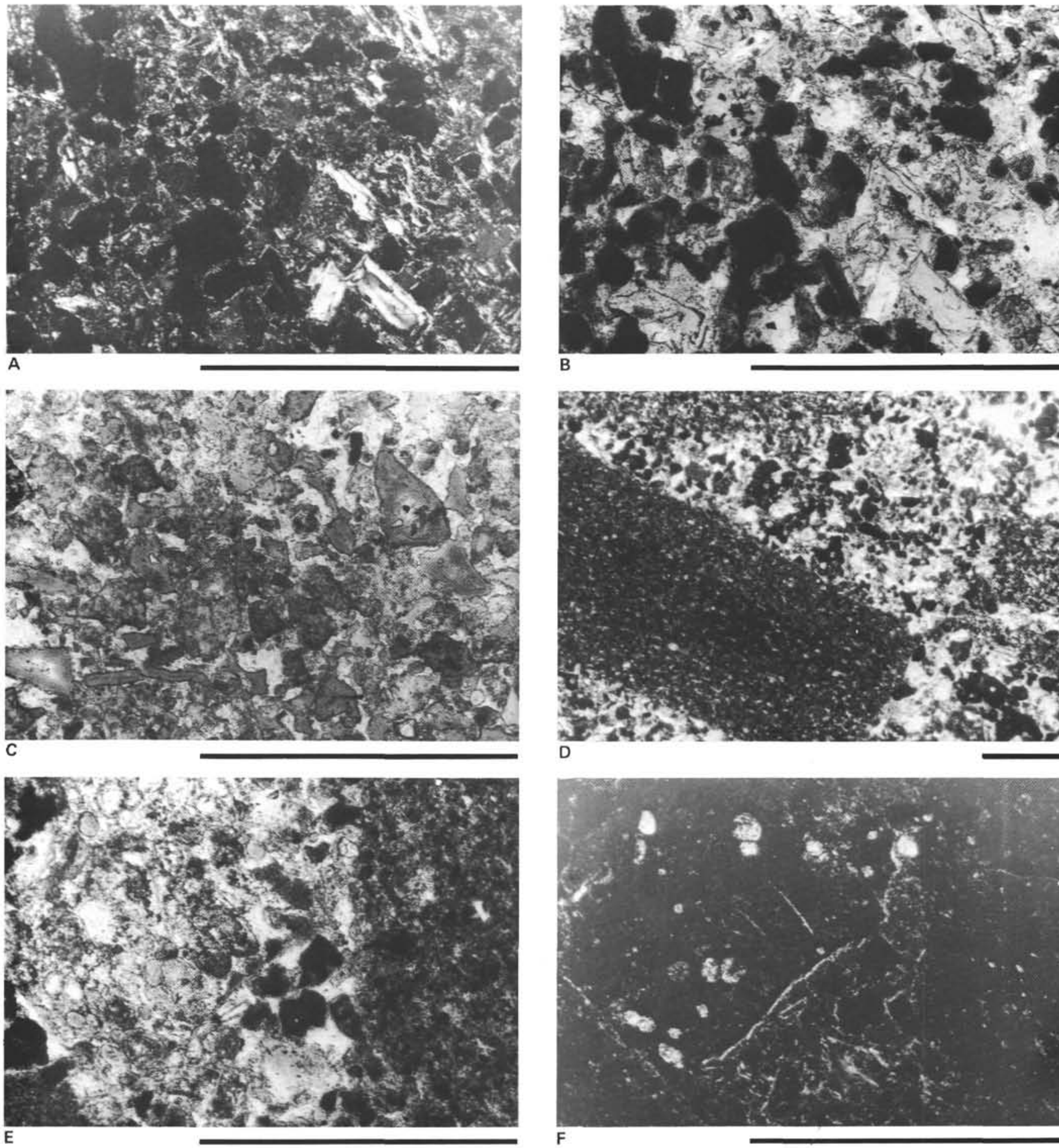

Figure 2. Photomicrographs of thin sections of hyaloclastites of Barremian age. Dark grains are interpreted as former tachylite; light ones as former sideromelane. Length of bar is 0.5 mm. A. Blocky, fine-sand-sized grains of altered hyaloclastite with microfibrous fringes of aligned clay flakes. Plagioclase grains at lower right (462A-80-1 at 102 cm; partly crossed polarizers). B. Same as Figure A; plane-polarized light. C. Angular to blocky fine hyaloclastite sand from graded bed shown in Figure 1B, a few cm from fossiliferous horizon (462A-80-1 at 11 cm; plane-polarized light). D. Breccia clast from hyaloclastite breccia shown in Figure 1D. Clast is of a fine-silt-sized hyaloclastite, now in a matrix of light and dark hyaloclastite sand and silt (462A-80-2 at 60 cm; plane-polarized light). E. Clasts (right side and lower left corner) of hyaloclastite of dark fine sand and coarse silt, and matrix of mainly light sand-sized grains. From breccia of Figure 1C (462A-80-1 at 45 cm; plane-polarized light). F. Same as 2E, crossed polarizers, showing birefringent clay as veinlets and as films around dark (former tachylite glass?) hyaloclastitic sand grains.

hyaloclastite siltstone (Figs. 1, 2D–F). The clasts float in a matrix of hyaloclastitic sand and silt. The coarse beds are rudely graded in both directions, i.e., fine to coarse to fine.

Similar Barremian radiolarians and benthic foraminifers are present at the top of Core 46 of Hole 462A, in hyaloclastite recovered from about 738 meters depth, near the top of the sheet-flow sequence. The sediment probably caved from higher in the hole, because the log shows no indication of sediment near that level.

The top of core 462A-70 has some fragments of hyaloclastite and hyaloclastic breccia that may have been in place, or may have caved from the side of the hole during a wire-line trip (Fig. 3A). The sonic-velocity log shows a 3-meter interval of decreased velocity there that may be caused by a hyaloclastite section like that of 992 to 995 meters, which also has a decreased velocity. Most of the low-velocity intervals below 729 meters, however, appear to be thinly sheeted glassy sections, such as in Cores 47, 48, 54, 56, 57, 59, 60, and 62, but massive doleritic flow interiors were recovered in Cores 67, 73, and 78 from depths showing low velocities on the log. Possibly some thin hyaloclastite layers were washed away in the coring process, but in any event they could not have been very thick.

Hyaloclastite of Middle Cretaceous Age

The upper sills of the volcanic complex intrude hyaloclastites that grade upward into zeolitic mudstones. Their abundance in place almost certainly is greater than their recovery, as is evident from the high-gamma-ray and low-sonic-velocity parts of the logs, and from inspection of Table 1, showing that many of the cores have their hyaloclastite sections at the tops of recovered portions. Of special interest in these sedimentary rocks are their fossil content, diagenetic and metamorphic alteration, and remains of organic carbon and plant fragments.

A late Aptian nannofossil assemblage provides a date for Core 40 of Hole 462A, the only fossil age between the sheet-flow complex (Barremian, as discussed above) and the top of the igneous rocks, which intrude late Albian or Cenomanian sediments.

The thickness of the sediments before intrusion can be determined by adding the parts interpreted as sediment from the well logs, assuming that during intrusion the injections of basaltic magma merely lifted sediment and did not shove it aside or ingest it. About 57 meters of sediment are in the present interval of sills between 564 and 729 meters. The accumulation rate was therefore about 1 to 2 m/m.y. for the 14 meters below dated Core 40, and about 6 to 10 m/m.y. for the approximately 53 meters of hyaloclastite between Core 40 and the late Albian or Cenomanian within Section 462-59-1, at about 550 meters depth. The total section of altered hyaloclastite from 462-A-80 through the lower Campanian zeolitic claystones in about 160 meters.

From about 700 to 729 meters, much of the section is hyaloclastite, similar in general to that described for Core 80. Primary sedimentary structures include horizontal and cross-laminae, graded beds, and soft-sediment deformation. Most of the sediment was originally angular fragments of basaltic glass of fine-sand to coarse-silt size, now mainly smectitic clay and analcime. Parts are palagonite that is red-brown from hematite (Fig. 3B). Metamorphic effects under some sills include the growth and coalescence of magnetic spheres in Section 41-7 (Figs. 3C, D). Analcime, smectite, fine needles of actinolite, patches of pyrite, magnetite, and radiating fibrous aggregates of an unidentified zeolite are present. The spheres are smaller but closer together near the sill; about 60 cm below the contact they did not develop, but actinolite continues, and some chalcedony is present. Clay flakes are especially coarse next to the sill in Section 40-2 (Fig. 3F). Some coarse sand and fine pebbles are present in Sections 40-2, 41-8 (Fig. 3E), 42-1, and 42-2. A discussion of the mineralogy and geochemistry of these beds is presented by Timofeev et al. (this volume).

Interest in Section 40-1 results mainly from the late Aptian nannofossil age it provides. Other items of interest are its content of pyrite-coated, carbonized scraps of plants, which probably accounts for the fractional percentage of organic carbon determined on one sample, and the soft-sediment deformation that has tilted some laminae 10° or more from the horizontal. These fossils and sediments are discussed by Jenkyns and Schlanger (this volume).

Above 700 meters, the hyaloclastites are known from both 462 and 462A (Table 1). Many of the features already described for the deeper rocks are present in these sediments as well. Magnetic spherulites are porphyoblasts in 462A-32, where the aureole extends about 0.3 meters below the sill contact. Some cores show laminations. The hyaloclastites are altered to clays and zeolites (or zeolite-like minerals), dominantly smectite and analcime.

The extent of alteration is also evident in the shipboard XRF analyses of samples from 63-1 and 64-2, and also from 66-1 (but that sample may have caved). Compared with sea-floor basalts in general, K_2O and loss on ignition are 5 to 20 times higher in the sediment. The relatively higher gamma radiation and lower velocities caused by these changes are evident in the logs. Also, MgO is about twice as high as in typical basalt, and total iron about the same or somewhat higher. Thus, the clays bear Fe, Mg, and K, and so are probably nontronite, Mg-rich saponite, and celadonite. CaO and MnO are distinctly lower, and TiO, Al_2O_3, and some other components generally lower. Thus, it appears that the moderate-temperature reactions between circulating sea water and basalt (Spooner and Fyfe, 1973; Bischoff and Dickson, 1975; Mottl and Holland, 1978) have acted on the basalt glass of these hyaloclastites. The zeolitic claystones above the sills apparently altered at lower temperatures, as will be discussed in detail in the next section. In Section 60-1, MnO is about 25 times greater, total iron 1.5 times higher, and MgO 0.5 times lower than in typical basalt, and both K_2O and loss on ignition are 40 to 50 times higher.

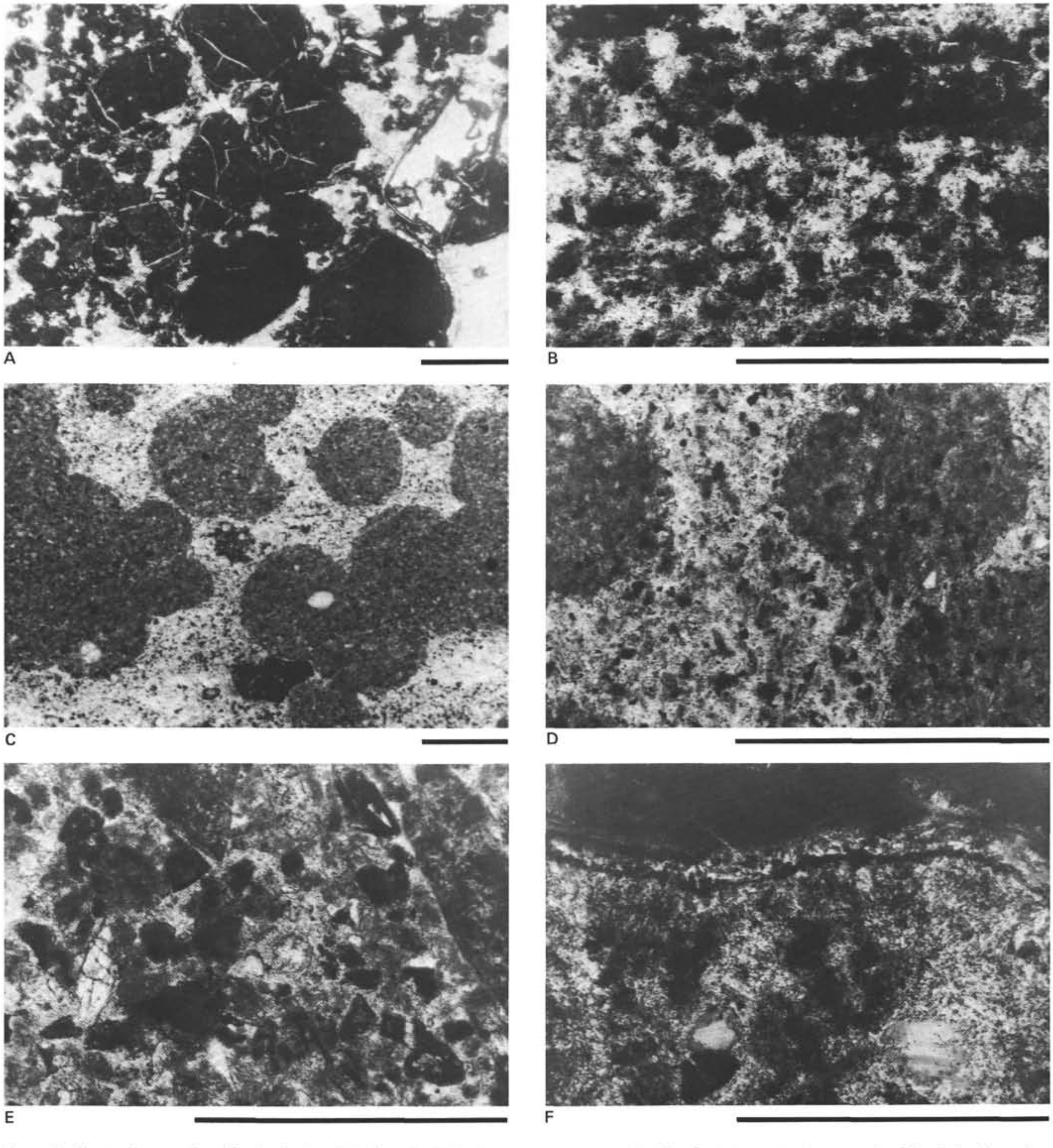

Figure 3. Photomicrographs of hyaloclastite of Early and middle Cretaceous age. Length of bar is 0.5 mm. A. Coarse-grained hyaloclastic sandstone of altered tachylite recovered from about 920 meters depth, probably in place within sheet flows and of Early Cretaceous age, but possibly cavings of middle Cretaceous age (462A-70-1 at 24 cm; plane-polarized light). B. Zeolitic hyaloclastite from about 729 meters depth. Radiolarians of the *Eucyrtis tenuis* Zone (Barremian) and benthic foraminifers were recovered from several horizons in the 6 meters of hyaloclastite that immediately overlie this core. Basalt section below 729 meters is composed of sheet flows, whereas the basalts above are sills (462A-44-1 at 1 cm; plane-polarized light). C. Spotted hyaloclastite about 10 cm below a massive basalt (dolerite) sill. Mineralogy of magnetic spheres listed in text (462A-41-7 at 113 cm; plane-polarized light). D. Another part of thin-section of contact-metamorphosed rock of Figure 3C, enlarged to show zeolitized hyaloclastic silt in which the spheres grew (plane-polarized light). E. Blocky, angular hyaloclastic sandstone below metamorphic spheres of Figures 3C and D (462A-41-8 at 14 cm; plane-polarized light). F. Contact between sill and hyaloclastite at about 704 meters depth. The sediment yielded a late Aptian nannofossil flora (462A-40-2 at 12 cm).

VARICOLORED CLAYSTONES AND RELATED SEDIMENTS ABOVE THE SILLS

Lithologic Description

Deposits of Santonian and early Campanian age in Holes 462 and 462A comprise grayish-red and reddish-brown claystones to siltstones containing occasional centimeter-thick zones colored in shades of bluish-green; these latter layers become less common down-section. Apart from the local presence of faint horizontal lamination and rare fucoid and *Chondrites* burrows, this argillaceous section is essentially structureless; rare intercalations of horizontally bedded, greenish-gray volcaniclastics provide the only punctuation. Of note is the presence of thin strands of native copper (462A-9-5) which are surrounded by haloes of pale-bluish-green sediment. Fossils, which are locally common, include radiolarians, sponge spicules, sparse nannofossils, and fish remains (Fig. 4A). The radiolarians typically are recrystallized, but are better preserved in the greenish-blue than in the reddish-brown claystones. These claystones pass downward through more-brownish-black (and manganiferous) claystones (Fig. 4B) that rest, with an intercalated black chert, on basalt of the highest sill.

Mineralogy

Six samples of these claystones were analyzed by conventional X-ray-diffraction techniques, using a Phillips PW 1050 diffractometer. Iron-filtered cobalt radiation (38kV, 24mA) was employed; smear-oriented specimens were run at a scanning speed of 1° of 2θ per minute. Results are set out in Table 2. The minerals are typical of pelagic clays: varying amounts of clay minerals, zeolites, and feldspars, plus quartz, phosphate, and calcite. Apart from the three last-named substances, the minerals betray a clear igneous parentage.

Most of the clay is mixed-layer illite–montmorillonite, with some smectite; the latter mineral strongly suggests derivation from degraded igneous material (e.g., Griffin et al., 1968) whereas the former's origin is perhaps more equivocal (Biscaye, 1965; Rateev et al., 1969). Rateev et al. (1969) commented that authigenic illite–montmorillonite clays either derive from diagenetic transformation of volcanic ash or are from continental soils; certainly for the more-smectite-rich mixed-layer minerals, an origin by submarine alteration of basalt seems well established (Eslinger and Savin, 1976; Perry et al., 1976). Furthermore, burial diagenesis can readily transform smectites to mixed-layer species, and much of the illite–montmorillonite may be a simple product of time and temperature (cf. Eberl and Hower, 1977). Palygorskite is also traditionally attributed to the alteration of smectite, in this case by Mg-rich solutions and excess silica (e.g., Bowles et al., 1971; Elderfield, 1976; Couture, 1977). That magnesium would be readily available in solution from the submarine alteration of basalt is apparent from the analysis of the Nauru Basin zeolitic claystones (average MgO = 3%), as opposed to the mean of ocean-floor basalt (average MgO = 7.84%; Cann, 1971). Soluble silica would be supplied by dissolution of opaline radiolarians and sponge spicules. Clinoptilolite similarly may form during the devitrification of volcanic glass, via smectite, and/or phillipsite, in silica-rich interstitial marine waters (e.g., von Rad and Rösch, 1972; Couture, 1977; Boles and Wise, 1978; Cosgrove and Papavassiliou, 1979). The feldspars are also of probable igneous origin (e.g., Peterson and Goldberg, 1962).

The distribution of quartz, apatite and calcite is largely biologically controlled, reflecting the presence or absence of radiolarians, fish skeletal remains, nannofossils, and structureless fine-grained carbonate of uncertain nature. A small eolian contribution to the quartz fraction is possible (cf. Rex and Goldberg, 1958).

Overall, the mineralogy of these sediments seems typical of the Pacific Cretaceous (cf. Couture, 1977).

Chemistry

The claystones were subjected to X-ray-fluorescence analysis, using a Phillips PW 1410 semi-automated spectrometer. Elemental-line intensities were measured with respect to igneous and sedimentary rocks of established composition, including several international standards.

The chemistry of the deposits is set out in Table 3. The distribution of elements is unremarkable and shows patterns typical of the halmyrolitic alteration of basalt. Relative to unaltered tholeiite (Cann, 1971), these sediments are clearly enriched in K_2O, and possibly in P_2O_5; depletions are apparent in the case of CaO and perhaps MgO. This exactly parallels the relations found by Hart (1973) in his study of altered lavas and shows comparisons with the data of Matthews (1971) and Melson and Thompson (1973) on Atlantic sea-floor basalts, except in the case of the more-independent magnesium. The high phosphorus content clearly relates to the presence of the fish debris. Magnesium apparently has been lost from the primary basalts of the Nauru Basin; such transfer into sea water seems the most usual situation, at least in low-temperature environments (e.g., Drever, 1974; cf., Perry, 1976).

As would be expected, the grayish-blue-green sample is considerably richer in divalent Fe than the reddish-brown sediments, which overwhelmingly contain iron in its oxidized state. Interestingly, the turquoise-colored layer is relatively poor in total iron, perhaps suggesting that reduction was secondary, and that some of the more-soluble Fe^{2+} was removed into adjacent zones.

Some trace-element data are given in Table 4; no startling geochemical trends are exhibited. The levels, although varied, do not depart significantly from those of ocean-floor basalts (cf., Nicholls and Islam, 1971), and there is no strong suggestion of mobility in these metals.

These multicolored claystones, compared with recent Pacific pelagic clays, show roughly comparable values of SiO_2, MgO, and K_2O; slightly lower values of Al_2O_3 and CaO (excluding the carbonate-rich sample); lower NaO; distinctly lower MnO; comparable or perhaps slightly higher values of Fe_2O_3; higher P_2O_5; and higher TiO_2 (cf. Goldberg and Arrhenius, 1958; Cronan, 1969). The trace-element concentrations in the Nauru

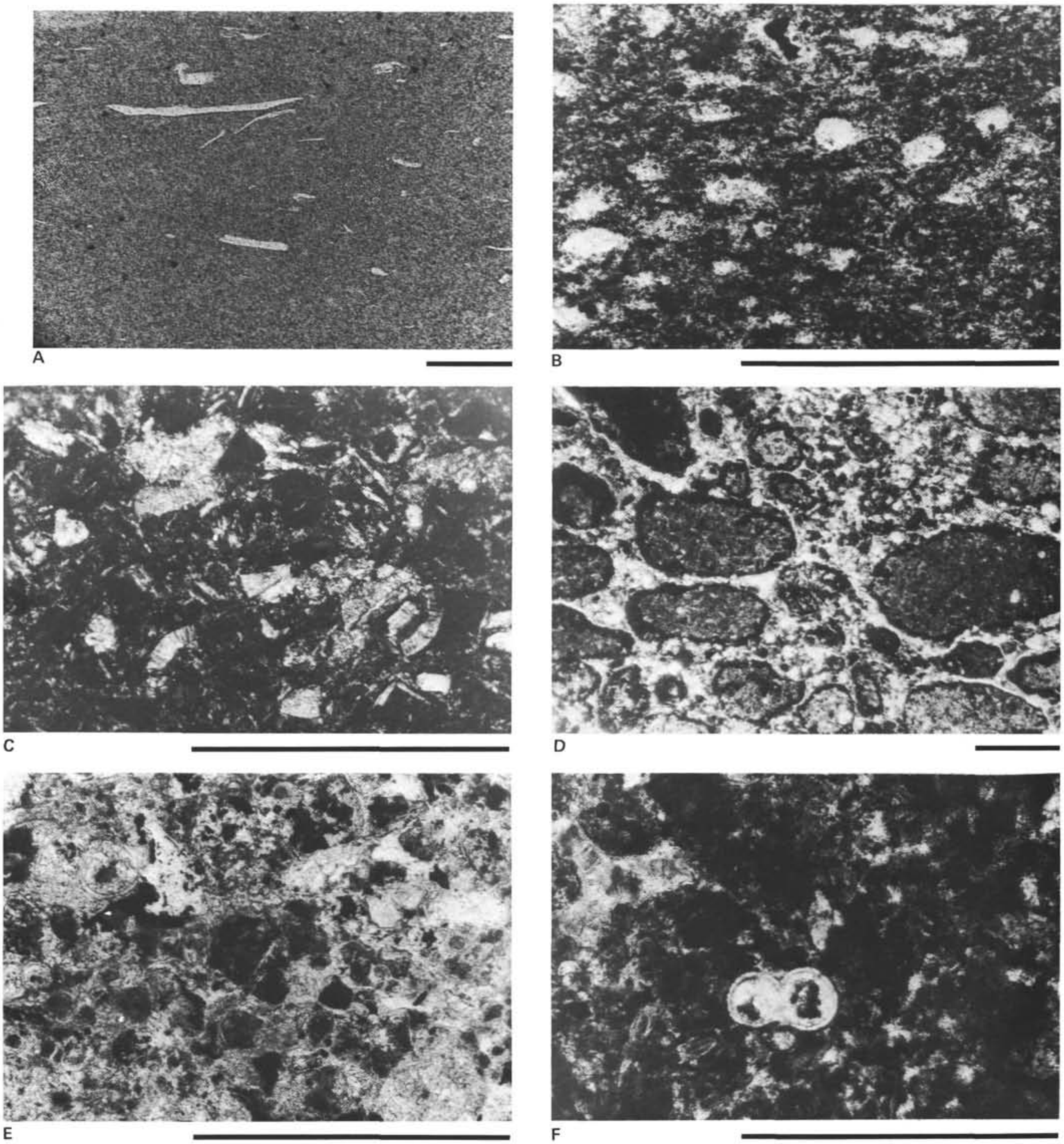

Figure 4. Photomicrographs of volcanic-rich sediments of middle to Late Cretaceous age. Length of bar is 0.5 mm. A. Zeolitic claystone, showing abundant fish remains parallel to bedding, of ?Early Santonian age (462-58-3 at 129 cm; plane-polarized light). B. Dusky-brown claystone showing relict hyaloclastite texture, of probable late Albian age, recovered within 1 meter of highest sill encountered in Nauru Basin (462A-14-1 at 68 cm; plane-polarized light). C. Vermicular authigenic clays within hyaloclastic sandstone of possible late Santonian age (462A-11-1 at 139 cm; partly crossed polarizers). D. Hyaloclastic siltstone redeposited as coarse sand-sized grains, with hyaloclastite matrix, all altered to clay, and cemented with calcite, early Maestrichtian in age (462-50-1 at 132 cm; plane-polarized light). E. Redeposited foraminifers and both epiclastic and hyaloclastic volcanic grains, from a sandstone layer within late Campanian limestone (462A-8-2 at 40 cm; plane-polarized light). F. Redeposited, cross-bedded, early Maestrichtian volcanic sandstone, from lower part of Fig. 6D; patch of celadonite cement at upper left (462-50-3 at 137 cm; plane-polarized light).

Table 2. X-ray-diffraction data from zeolitic claystones of Holes 462 and 462A.

Sample (interval in cm)	Mixed Layer Illite-Montmorillonite and Smectite	Illite	Palygorskite	Clinoptilolite	Feldspar	Calcite	Quartz	Apatite
462A-9-6, 0-5	Present	Trace	Present	Present	—	Trace	Present	Present
462-56,CC	Present	Trace	Present	—	—	Trace	Present	—
462A-10-2, 61-64	Present	Trace	Present	—	Present	—	Present	—
462A-10-2, 90-94	Present	Trace	Present	—	Present	Trace	Present	Present
462A-10-3, 138-142	Present	—	—	Present	Present	—	—	Present
462A-11,CC	Present	—	—	Present	—	Present	Present	Present

Table 3. Major-element X-ray-fluorescence data (%) from zeolitic claystones of Holes 462 and 462A.[a]

Sample (interval in cm)	Color	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	MgO	CaO	K_2O	MnO	P_2O_5	Na_2O	FeO	Volatiles	Total
462A-9-6, 0-5	Pale reddish-brown	55.51	12.44	1.10	8.15	3.81	1.08	3.14	0.17	0.26	1.99	0.19	12.91	100.75
462-56,CC	Pale reddish-brown	53.60	14.00	1.07	8.20	3.62	1.14	3.19	0.15	0.32	1.65	0.15	13.15	100.25
462A-10-2, 61-64	Pale reddish-brown	52.20	14.33	1.09	7.77	3.69	1.29	3.37	0.15	0.42	1.61	0.18	13.60	99.67
462A-10-2, 90-94	Greyish blue-green	56.01	14.57	1.12	2.92	3.78	1.09	3.70	0.08	0.34	1.45	2.59	12.77	100.41
462A-10-3, 138-142	Moderate brown	53.91	13.24	1.64	10.31	2.78	1.99	1.85	0.08	0.23	2.94	0.27	12.73	101.97
462A-11,CC	Pale yellowish-brown	46.15	9.90	1.55	8.12	3.36	10.31	1.6	0.24	0.31	2.40	0.28	17.18	101.48

[a] The chemistry of such deposits departs from that of average Pacific clays and active-ridge sediments by manifesting enhanced values of titanium.

Table 4. Minor-element X-ray-fluorescence data (ppm) from zeolitic claystones of Holes 462 and 462A.

Sample (interval in cm)	Ni	Cu	Zn	Ba	Pb
462A-9-6, 0-5	100	85	146	195	23
462-56,CC	92	119	123	253	14
462A-10-2, 61-64	98	114	133	234	15
462A-10-2, 90-94	114	36	163	190	8
462A-10-3, 138-142	114	34	127	111	6
462A-11,CC	103	24	96	109	3

Basin claystones are distinctly lower than those of recent Pacific pelagic clays, except in the case of zinc.

When compared with Pacific spreading-ridge sediments, the multicolored claystones are markedly enriched in SiO_2, Al_2O_3 and TiO_2; locally enriched in K_2O; and depleted in P_2O_5, Fe_2O_3, and particularly MnO; trace metals are also lower in the claystones (cf., Boström and Peterson, 1966, 1969; Boström, 1973; Dymond et al., 1973; Cronan, 1976). Although the quantities of Mn are low, as the contact with basalt is approached the colors change to shades of brown through to black, and the amounts of manganese rise (462A-13-1, 27-31 cm; 13-2, 138-142 cm; 13-3, 1-3 cm: MnO = 0.23, 1.56, and 2.05% respectively; shipboard XRF data). Other elements (Cr, Ni, Co, Sc, Pb, and Cu) also increase in abundance towards the contact (Kurnosov, this volume). The basalt/sediment contact is at 15-1, 12 cm in Hole 462A.

From the above brief survey, it is clear that the basal or near-basal Nauru Basin sediments show chemical features both in common with and distinct from Pacific pelagic clays and spreading-ridge deposits. The only element uniquely concentrated is titanium, and some explanation must be sought for this enrichment in the context of the genesis of these multicolored claystones.

Discussion

The geochemical behavior of titanium, which is readily hydrolyzed, is largely governed by the immobility of the dioxide above a *p*H of 2.5; thus, the element is concentrated in virtually all *subaerial* weathering environments (e.g., Goldschmidt, 1954; Rankama and Sahama, 1950; Loughnan, 1969). Indeed, titanium is known to be concentrated in many highly weathered soils, such as bauxites (e.g., Allen, 1952), and is also enriched in Carboniferous seat earths (e.g., Huddle and Patterson, 1961; Spears and Kanaris-Sotiriou, 1976). In different soil profiles, titanium may be present as rutile or anatase or, to a much lesser degree, as a substituted ion in various clay minerals (Dolcater et al., 1970).

Most relevant, perhaps, in any comparison with lavas altering to sediment on submarine plateaus and seamounts are studies of soils produced above oceanic-island basalts. The soils developed on Hawaii, for example, are highly titaniferous and contain a variety of minerals, including rutile, anatase, and a variety of Ti-rich ferruginous oxides (Katsura et al., 1962; Walker et al., 1969). In summary, high titanium contents in soils formed on basaltic and other substrates are indicative of intense weathering. Similarly, the enhanced Ti levels of the Nauru Basin claystones must be a token of *submarine* degradation of basalt. These sediments, therefore, represent a distinct class of basal deposit separate from those found on spreading ridges, and distinct again from typical pelagic clays, to which they nonetheless contribute (cf., Bonatti et al., 1973). Thus, as would be predicted, surficial sediments from the present Pacific Ocean show maximal Ti/Al ratios close to oceanic islands and seamount chains (Boström et al., 1973).

Bertine (1974) has suggested that the sediments on the slow-spreading Lau Basin Rise may be largely attributed to submarine weathering of basalt. These deposits are not, however, particularly rich in titanium—nor would one expect this with a spreading-ridge sediment—although there is one revealing analysis of a smectite-filled vesicle containing 2.8% TiO_2. The overall lack of any characteristic titaniferous signature must therefore militate against his interpretation.

Sediments very similar to the Nauru Basin claystones were cored off the Line Islands during Leg 33. These deposits, colored in various shades of red and blue-green, are similarly titaniferous, and were attributed by Jenkyns and Hardy (1976) to submarine weathering of basalt. Comparative analyses are presented in Table 5; clearly there is a general chemical similarity, although the samples from the Line Island site are generally more richly endowed with Fe_2O_3, MnO, and MgO and contain some levels particularly rich in TiO_2. Traces of anatase were in fact recorded from Hole 315A; this mineral was not identified in the Nauru Basin.

As mentioned above, as the basalt sill is approached down-hole the content of manganese and other metals rises, perhaps suggestive of post-depositional injection of these elements from weak hydrothermal solutions from the hypabyssal-sill complex. Similar geochemical trends were reported by Natland (1973) in his study of red claystones—apparently also produced by weathering of basalt—lying atop Meiji Guyot in the North Pacific. Persuasive, perhaps, of the reality of minor hydrothermal injection is the presence of native copper in the Nauru Basin claystones some 47 meters above the igneous/sedimentary-rock contact; certainly, the several occurrences of native copper in sediments of DSDP cores are conventionally attributed to this or a related mechanism (e.g., Hole 105, Lancelot et al., 1972; Hole 317A, Jenkyns, 1976; Hole 364, Siesser, 1978).

In conclusion, it is herewith suggested that the multicolored claystones of Holes 462 and 462A were derived largely from *in situ* weathering of glassy basalt fragments (Fig. 4C), and titanium provides a geochemical fingerprint of this process. Given that Cretaceous intraplate volcanism now has been recognized as a phenomenon of regional significance across the Pacific, influencing or creating the Line Islands, the Marshall Islands, the Mid-Pacific Mountains, other edifices, and the Central Basin itself (Watts et al., 1980; Schlanger and Premoli Silva, this volume), it follows that basal titaniferous sediments should also characterize this area.

Presumably, the intrusions of the sills into the hyaloclastite and weathered hyaloclastite took place because the igneous/sedimentary contact was a plane of weakness. The density of the deeply weathered and partly pelagic Cenomanian through Santonian clays probably was not greatly different from the density of water, yet denser than the more rapidly deposited Lower Cretaceous sandy hyaloclastite, and so some sills were emplaced at the base of the deeply weathered section. Other masses of magma almost certainly broke through to the sea floor, as will be discussed in the next section.

VOLCANICLASTIC SEDIMENT OF LATE CRETACEOUS AGE

Between about 450 and 490 meters depth is a series of greenish-black to gray sandstones, siltstones, and fine breccias of volcanic composition, displaying graded bedding, scours, and soft-sediment deformation. Planktonic fossils within them are of late Campanian and early Maestrichtian age.

Sedimentary Structures and Textures

The texture, composition, structures, and probable volume of the volcaniclastic assemblage can be used to deduce its probable origin. The primary sedimentary structures and textures indicate that part of the material had a nearby source, whereas the composition indicates that part had traveled from the edge of the Nauru Basin. The volume suggests local sources.

Table 5. Major-element composition (%) of Nauru Basin zeolitic claystones compared with that of similar sediments (Santonian?) from Hole 315A, Line Islands.[a]

Sample (interval in cm)	Color	SiO_2	Al_2O_3	TiO_2	Fe_2O_3	MgO	CaO	K_2O	MnO	P_2O_5	Na_2O
462A-9-6, 0–5	Pale reddish-brown	55.51	12.44	1.10	8.36	3.81	1.08	3.14	0.17	0.26	1.99
462-56,CC	Pale reddish-brown	53.60	14.00	1.07	8.37	3.62	1.14	3.19	0.15	0.32	1.65
462A-10-2, 61–64	Pale reddish-brown	52.20	14.33	1.09	7.97	3.69	1.29	3.37	0.15	0.42	1.61
462A-10-2, 90–94	Grayish blue-green	56.01	14.57	1.12	5.80	3.78	1.09	3.70	0.08	0.34	1.45
462A-10-3, 138–142	Moderate brown	53.91	13.24	1.64	10.61	2.78	1.99	1.85	0.08	0.23	2.94
462A-11,CC	Pale yellowish-brown	46.15	9.90	1.55	8.43	3.36	10.31	1.6	0.24	0.31	2.40
315A-28-3, 104–106	Blue-green	66.69	9.00	1.43	10.87	5.51	1.04	3.09	0.53	0.08	0.82
315A-29-1, 31–33	Grayish blue-green to dusky blue-green	71.89	7.84	0.53	9.15	5.34	0.56	3.08	0.38	0.05	0.77
315A-29-1, 105–108	Grayish blue-green to dusky blue-green	57.70	11.07	2.30	12.47	6.14	2.50	2.75	0.44	0.38	1.34
315A-30-1, 65–66	Grayish-red to very dusky red	55.88	13.30	1.50	12.11	5.86	1.40	4.33	0.36	0.28	0.98
315A-30-1, 103–104	Grayish red to very dusky red	55.23	13.45	2.57	12.68	4.64	1.09	5.42	0.40	0.29	0.89
315A-30-2, 132–134	Moderate reddish-brown to dark reddish-brown	54.32	8.03	1.41	19.56	3.84	1.54	1.49	0.66	0.30	1.12

[a] FeO recalculated as Fe_2O_3 in analyses from Holes 462 and 462A. Note the strong chemical similarity between the two sets of claystones.

A single graded bed of 2.5 meters thickness in Core 51 is the thickest of several volcanic-rich beds discovered. Its top shows escape burrows, and its base had inclined bedding and lies on contorted marlstones having phacoidal structure (Fig. 5A). If the entire bed has much lateral extent, its volume must be immense. To us it seems probable that most if not all of its source was a nearby slope that was gravitationally unstable.

The slump fold in Figure 5B, like that in 5A, suggests that the mass of sediment overlying the deformed position slipped *en masse*. Another slump is near the top of 462-51-4. These features record gravitational instability on at least three occasions. The sea floor close by was elevated or depressed fairly quickly relative to the DSDP site, and that may have been a consequence of nearby volcanism or faulting.

Scoured surfaces indicate strong local currents, which in themselves do not indicate closeness of source. The overhanging cut of Figure 5C is at the base of a section of centimeter- to millimeter-scale graded beds, at the top of which are small scours (Fig. 5D) filled with cross laminae that display lateral gradation. A similar but thinner sequence of overhanging cut, thin turbidites, and cross-bedded top is in 462-49-4.

Load casts from cross-bedded volcanic sands deposited on burrowed calcareous ooze, now sandstone on limestone, are in 462-50-5 and 51-1. They and the escape burrows in other cores (Fig. 6A) resulted from the sudden, episodic introduction of volcanic sand into this part of the Nauru Basin.

Composition

Small but important parts of these rocks are beds of redeposited shallow-water sands, reef limestones and fossils, phosphorite, and probable subaerial basalt. These sandy limestones are evidence of the growth and erosion of the volcanoes that today are the Marshall or Gilbert Islands and guyots, as discussed elsewhere in this volume (Schlanger and Premoli Silva). Some beds are pelagic limestones.

A great part of the sand and silt is hyaloclastite, now smectitic clay. The coarser sand grains of some beds are rounded intraformational clasts of hyaloclastic siltstone (Fig. 4D), with a sparse matrix of fine hyaloclastic sand, cemented by calcite spar. Pebbles and sand of basalt glass altered to clay are at the base of some graded beds (Figs. 5A, 6B). Even if they were not clay at the time of transportation, their size indicates that they did not travel far.

In the breccia bed of 462-51-1, the pebbles are mainly of volcanogenic siltstone, but also include limestone and basalt fragments (Fig. 6C). This intraformational breccia almost certainly had a closely adjacent source.

Some beds contain angular hyaloclastite with planktonic foraminifers and nannofossils, along with reworked grains of probable shallow-water origin (Fig. 4E), suggesting mixing and sorting of three components.

Fluids that moved in the sediment also provide some information about its origin and history, but some of the evidence is equivocal. Upon heating, organic matter forms hydrocarbons, and thermally mature hydrocarbons are preserved as high as the lower Maestrichtian rock in Core 462-49 (Thomson et al., this volume). Hydrocarbons may be generated and migrate one or more times, and so the unanswered question remains: Was there a Late Cretaceous episode of heating prior to the one that can be correlated with the late Santonian–early Maestrichtian volcaniclastic sediments?

In Section 462-50-3, the cross-beds between 136 and 141 cm (Fig. 6D) are partly replaced by iron-rich brown clays in shapeless masses that surround veinlets of bright-green, moderately birefringent phyllosilicate, probably celadonite (Fig. 4F). Apparently the bedding planes were conduits for hydrothermal solutions during (or after) the early Maestrichtian.

Possible Volume

Site 462 is within the Nauru Basin, which covers an area of about 1×10^6 km^2. The upper seismic reflectors and level topography are more or less continuous across the basin. Extrapolation of a third dimension from the record of a single drill site is extremely risky, but there is perhaps a 10-meter thickness of volcanic addition to the basin within the Campanian and Maestrichtian strata—the 30 meters of sediment of that age, reduced by the pelagic intervals and by the pores in the volcanic sands. To an order of magnitude, there may have been about $10 \pm 5 \times 10^4$ km^3 of lava that provided the sediment. How might that volume have been attained?

About 20 of the Marshall and Gilbert volcanic pedestals rim the north and east edges of the Nauru Basin. Each has a present volume of 1 to 8×10^4 km^3, but it is impossible to measure what volume they could have contributed to the basin. If we assume that all former rock above the present regional guyot depth of about 1100 meters was eroded as volcanic sand and silt (none leached or weathered to clay), and that all was delivered southwest into the Nauru Basin, that volume would have been about 4×10^3 km^3. Submarine erosion of fragmented glass from the volcano slopes may have doubled or tripled this amount, but it remains small compared with the estimate of silt and sand in the basin.

These very rough calculations suggest that—although far-traveled turbidites from the Marshall or Gilbert Islands were known to have contributed shallow-water detritus to the Nauru Basin—it is unlikely that turbidites were the sole source or even the dominant source of the volcanic grains. Our suggestion is that multiple nearby sources were available, wherever the feeders of the hypabyssal basalt sills broke through to the sea floor.

CENOZOIC VOLCANIC SEDIMENTS

Epiclastic volcanic grains are present in several of the beds of shallow-water debris that were carried into the Nauru Basin episodically, and they provide indirectly a history of the Marshall Islands. Pyroclastic contributions are fairly common in the Neogene, and an ashy layer in Core 462-5, logged at about 45 to 50 meters, probably indicates the late Pliocene age of nearby Kusaie Island. The pyroclastic and epiclastic (turbidite) components and their significance are discussed in the

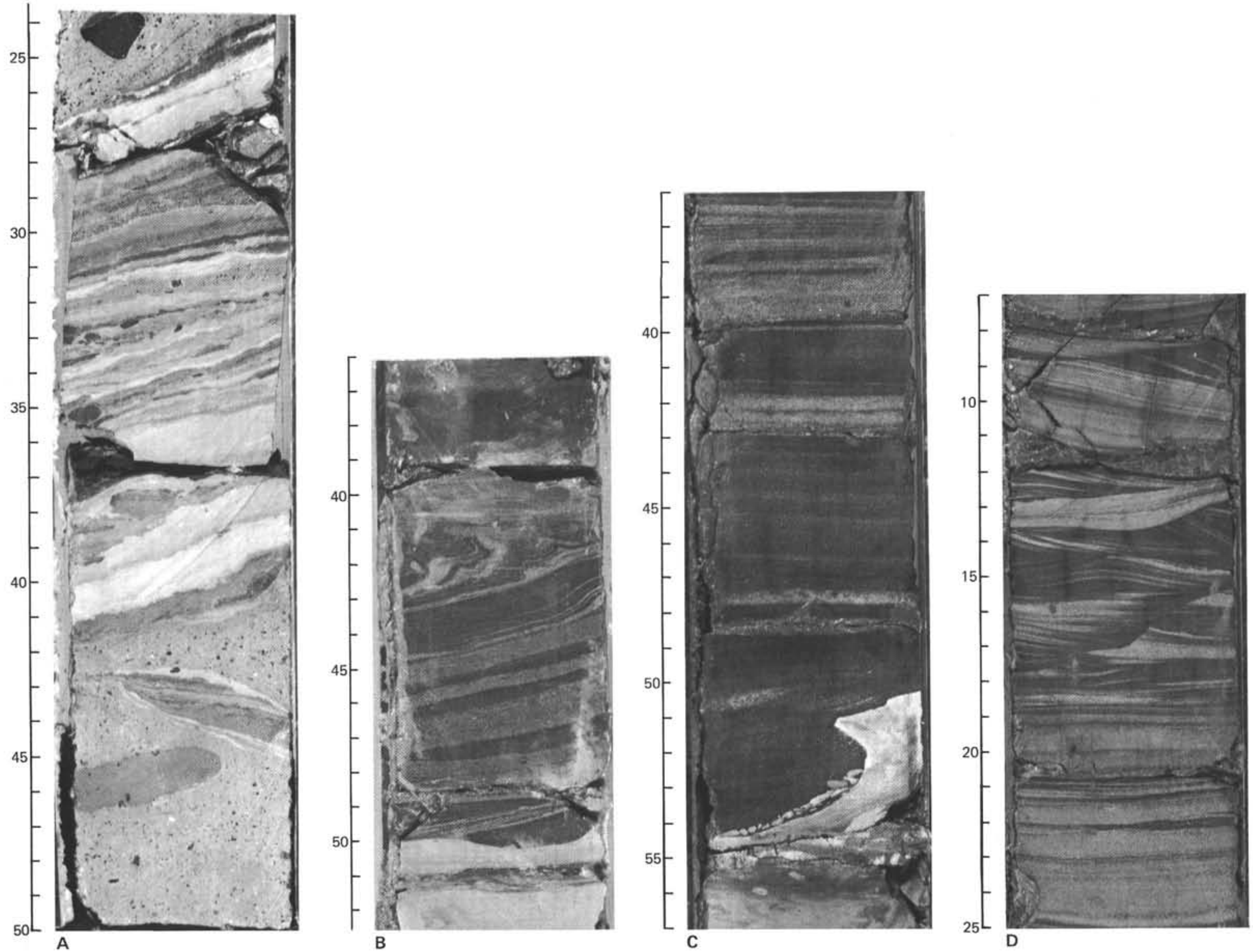

Figure 5. Photographs of split cores showing structures in volcanic-rich sedimentary rocks of Late Cretaceous age. Section depth (cm) shown at left of each core. A. Base of 2.5-meter-thick graded and inclined bed on contorted marlstone with phacoidal structure, on another tilted bed with centimer-thick marlstone clast, late Campanian (462-51-3). B. Slump fold in dark-gray volcaniclastic siltstone that has scoured the underlying light-gray nannofossil chalk (462-49-2). C. Scoured marlstone (overhanging cut, and small clasts within cut) overlain by thin, graded beds of volcaniclastic sandstone and siltstone (462-50-1). D. Scoured and cross-laminated volcanic siltstone at top of section that has 5C at base, early Maestrichtian (462-50-1).

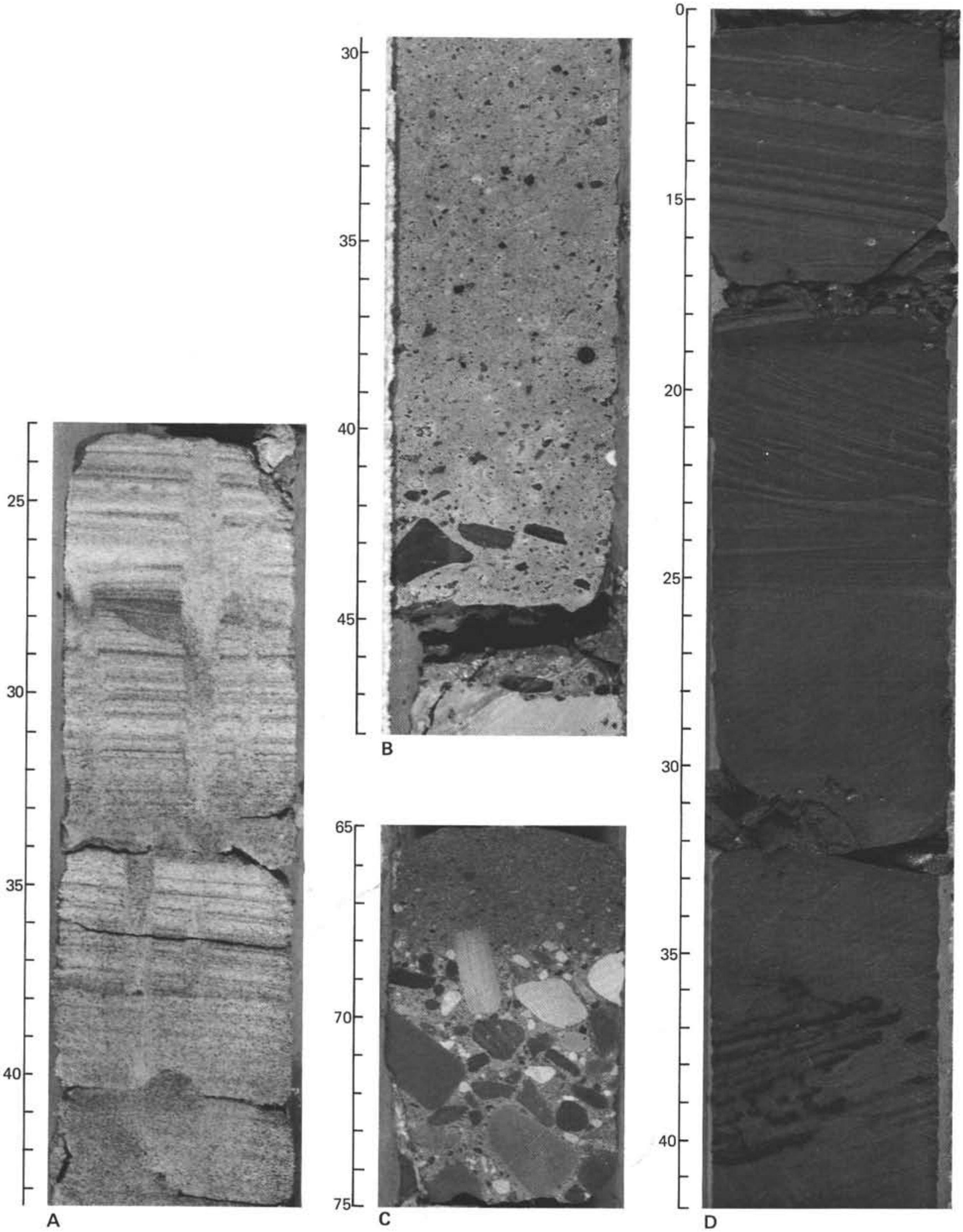

Figure 6. Photographs of split cores, showing structures and textures in volcanic-rich sedimentary rocks of Late Cretaceous age. A. Escape burrows in calcareous volcanic sandstones of middle Maestrichtian age (462-48-1). B Conglomeratic base of a graded bed of late Campanian age, probably representing initiation of nearby volcanic activity in Nauru Basin (462-52-1). C. Intraformational breccia of hyaloclastite and limestone, with some basalt clasts, indicating disruption of sediments within the immediate vicinity of the deposit, probably related to volcanism that broke through the sea floor (462-51-1). D. Cross-bedded hyaloclastite sand, now altered to clays and cemented by celadonite, within early Maestrichtian sediments (462-50-3).

Site Summary and by Schlanger and Premoli Silva (this volume).

AGES OF THE IGNEOUS COMPLEX

Two Cretaceous episodes of igneous activity in the Nauru Basin are shown by the evidence at Site 462. Both followed formation of the crust in Jurassic time. In Barremian time, voluminous basaltic flows erupted. Some of the hyaloclastite they produced was covered by flows, its meager radiolarian fauna recording the age, and its bathyal benthic foraminifers recording the depth of water. Many of the flows are thin, glassy sheets. The basalts have a TiO_2 content of about 0.9 to 1.1%, and assemblages of phenocrysts include olivine.

For several million years, a major component of the sediment lying on the flows was hyaloclastic glass of fine-sand and silt sizes, weathered on the sea floor and reworked into the Aptian, Albian, and Cenomanian from exposures of the decrepitated flows. The fine-clay component of the Cenomanian, Turonian, Santonian, and lower Campanian sediments came partly from submarine weathering and reworking of the altered hyaloclastites, and partly from pelagic sources.

This hyaloclastite and clay section was then intruded by numerous thin and thick sills of dolerite. They are mainly without olivine phenocrysts, and their TiO_2 contents are 1.1 to 2.2%. The age of this second episode of Cretaceous intraplate igneous activity is probably late Campanian, based on the following evidence. Sills intrude late Albian or Cenomanian sediments, and so clearly are younger. The Turonian through early Maestrichtian sediments have had their organic component heated, driving off some lighter hydrocarbons and maturing others, presumably a result of the sills. The sills are of normal magnetic polarity. In the well-studied Gubbio section, there is a moderately long epoch of normal polarity in the upper Campanian rocks (Guddio B+; Alvarez et al., 1977).

Of most importance, however, is the record of upper Campanian volcanic sands and fine breccias. Their principal source almost certainly was adjacent parts of the sea floor where the sills broke through to the surface, glassy basalt chilling and breaking apart, providing hyaloclastite for gravity and currents to spread nearby. This deposition continued into Maestrichtian time. The celadonite veinlets at 462-50 suggest that the early Maestrichtian was still a time of moderate hydrothermal circulation.

At Site 462, these Campanian and early Maestrichtian beds are the oldest which contain an admixture of shallow-water fossils and volcanic grains of probable epiclastic origin; they thus may record the origin and initial erosion of the Marshall Islands (Schlanger and Premoli Silva, this volume). Individual episodes of Cretaceous intraplate volcanism were not coeval by any means (the Mid-Pacific Mountains clearly were earlier), so there is no requirement to correlate the Nauru sills with the Marshall edifaces, but the evidence from the Marshall Islands debris is that voluminous igneous activity was occurring in this region in the Campanian.

Thus, the majority of the sedimentological, petrological, paleontological, and geophysical evidence favors the interpretation summarized above (three episodes in the igneous history of the Nauru Basin): (1) about early Oxfordian, 155 m.y. ago, crust; (2) Barremian, 115 m.y. ago, flows; and (3) late Campanian, 74 m.y. ago, sills. There have been, however, some other interpretations of this history, and it is appropriate to evaluate them:

1) *The flows (below 729 m) are oceanic basement.* The point is that mid-plate volcanic events are uncommon, according to DSDP drill sites in several oceans, and the flows at 462A are similar in structures, petrography, and geochemistry to basalts formed at spreading centers. If that were the case, there is either a misidentification of the *M*-anomaly sequence of the Nauru Basin, or a misidentification of the Early Cretaceous radiolarian faunas of Core 462A-80. Moreover, intraplate volcanism actually is common in the western and central Pacific record of the Cretaceous (Schlanger and Premoli Silva, this volume).

2) *Radiometric ages provide the proper framework for interpreting the igneous episodes.* The radiometric ages obtained are by Ozima et al. (this volume) by the $^{40}Ar/^{39}Ar$ method, giving 120 m.y. (462-60-1, 65–69 cm), 110 m.y. (462A-32-1, 46–49 cm), and 131 m.y. (462A-50-3, 130–134 cm), and by J. J. Naughton (pers. comm.) by the K/Ar method, giving 127 ± 7 m.y. (462A-32-1, 31–32 cm), 144 ± 5 m.y. (462A-50-3, 44–52 cm), and 134 ± 12 m.y. (462-69-10, 50–53 cm).

Considering the methods used, the low potassium content of these particular rocks, and the comments of the analysts, probably the best of these ages is 110 m.y. at 462A-32: Aptian, in the normal magnetic epoch between M0 and M1. If that were the case, with the Aptian sediments in Core 40 below the 110-m.y.-old intrusive sill there is either a misidentification of the Albian nannofossil flora, or the radiometric age is too young.

Similar arguments can be made for the other radiometric ages. The ages 120, 127, and 134 m.y. also are from the sills overlying the Albian sediments. The ages 131 and 144 m.y. at Core 462A-50 are in a flow interior that lies above the Barremian (maximum range about 112 to 118 m.y.) sediments of Core 80. Apparently, either some primary argon from the mantle was retained when these igneous rocks cooled, or some potassium was leached from them—or both. The deeper flows are dated 10 to 20% older than their fossil age, and the higher sills are dated 50 to 80% older than their probable fossil age.

3) *The sills were emplaced in Albian or Cenomanian time.* In this case they would be more or less contemporaneous with the fossiliferous sediments they intrude. Thus, there would be only a 5 to 10% discrepancy in the best of the radiometric ages (110 m.y.). The rocks having formed distinctly within the Cretaceous long normal, there would be no problem of explaining how more than 20 sills could have been emplaced within a single magnetic epoch.

On the other hand, the history of hydrocarbon maturation and the volcaniclastic record of the Upper Cretaceous require at least one Late Cretaceous volcanic episode. The coarse soft clasts and thick beds of coarse sand, and slump and scour structures, must have had a nearby source; most of the Campanian volcaniclastics did not travel great distances as turbidites, as did so much of the Cenozoic section. If the sills are Albian or Cenomanian, then there would have been four major igneous events: Jurassic crust, Early Cretaceous flows, mid-Cretaceous sills, and Late Cretaceous volcanism. Of course, it is only a small additional step from a postulate of three events to one of four events, but by the rule of parsimony we prefer the simpler explanation.

4) *All igneous rocks encountered at Site 462 were emplaced in a single episode*. The argument is that a section of Cretaceous sediment was intruded by a single complex of sills, more or less emplaced at successively higher levels over a fairly brief span of Campanian time. That would cut the number of igneous events (further parsimony), as well as provide an explanation for the general kindred of the lavas (but with increasing differentiation, reflected by Ti and trace elements, at higher and higher injections). The glassy sheets or pillows would be explained as earlier sills emplaced into cold, wet sediment, and the stratigraphic positions of the fossiliferous sediments would be correct.

Upon further examination, it seems very unlikely that so many tens of meters of pillows or sheets were all chilled by entering cold sediment. The formation of pillows has only been well studied in shallow water off Hawaii (Moore, 1975), but the geological evidence is that they have formed by intrusion into lakes, swampy muds, and under glaciers, as well as under oceans. They can become intimately mixed with sediment, so that the sequence of events is difficult to interpret (Garrison, 1973). Yet, apparently there is no geologic record yet studied of tens of meters of glassy volcanic units emplaced below tens of meters of sediment. Actually, the reverse of the stratigraphic arrangement at Site 462 might be expected, if at all; the deeper, undifferentiated intrusions should have been well-blanketed and formed sills, whereas only the latter, shallower, differentiated basalts might have been under a sediment cover sufficiently thin to allow rapid cooling and pillow formation.

The petrological information summarized in this volume by Batiza et al. indicates that, although all the recovered basalts are superficially similar to one another, in detail they are in two suites that are distinct from each other and from mid-ocean-ridge basalts.

There is also the difficulty of providing the hyaloclastites, 150 meters or more in thickness, into which the pillows were supposed to have intruded. Hyaloclastites are thought to be relatively scarce at oceanic depths (Williams and McBirney, 1979), even though much of the basal Fe- and Mn-rich clays at the base of many deep-sea sections show a relict hyaloclastite texture in thin section (Moberly and Heath, 1971; Moberly, unpublished data). According to Bonatti (1967), hyaloclastite can be common near seamounts. It may form most of the archipelagic aprons of seamounts that did not reach the sea surface (Moberly and Keene, 1975). If an immense, adjacent pile of hyaloclastite had formed at the time of Jurassic crustal age, and were periodically eroded, transported, and redeposited throughout the Cretaceous along with the appropriate fossils into the region of Site 462, that circumstance would be distinctly unusual compared to deep-sea evidence elsewhere. It seems more logical to consider the unusual event to have been an Early Cretaceous episode of post-ridge-crest volcanism, rather than an adjacent mass of Jurassic hyaloclastite which slumped and was redistributed throughout the Cretaceous.

CONCLUSIONS

Modifications and combinations of the interpretations of volcanic and sedimentary history in the Nauru Basin given above have been considered, but none seems to fit the evidence better. We conclude, therefore: (1) oceanic crust formed in the Nauru Basin in the Jurassic, (2) the basin experienced post-ridge-crest igneous activity in the Early Cretaceous in the form of flows; within and on the flows are hyaloclastic sediments with Barremian fossils; (3) the flows are overlain by hyaloclastic sediments from the Early Cretaceous volcanism, strongly altered by submarine weathering and moderately reworked, incorporating middle Cretaceous pelagic, fossiliferous sediment into the section; (4) an additional episode of post-ridge-crest volcanism in the Late Cretaceous resulted in intrusion of sills into the mid-Cretaceous section and contributions of volcanic sediment to the Upper Cretaceous sediments; apparently, the Marshall Islands or some other nearby seamount chain was formed at that time.

ACKNOWLEDGMENTS

This work was supported by resources of the Hawaii Institute of Geophysics, University of Hawaii, and by Department of Geology and Mineralogy, University of Oxford, and reviewed by John M. Sinton, Michael O. Garcia, and S. O. Schlanger. We thank S. O. Schlanger and Rodey Batiza for discussions about the regional setting of Site 462, the petrology of the igneous rocks that were recovered, and help in the routine shipboard processing of the sedimentary section. Assistance by the shipboard scientific and technical personnel of Leg 61 and 62 is appreciated.

REFERENCES

Allen, V. T., 1952. Petrographic relations in some typical bauxite and diaspore deposits. *Geol. Soc. Am. Bull.*, 63:649–688.

Alvarez, W., Arthur, M. A., Fischer, A. G., et al., 1977. Upper Cretaceous–Paleocene magnetic stratigraphy at Gubbio, Italy. V. Type section for the late Cretaceous–Paleocene geomagnetic reversal time scale. *Geol. Soc. Am. Bull.*, 88:367–389.

Bertine, K. K., 1974. Origin of Lau Basin Rise sediment. *Geochim. Cosmochim. Acta*, 38:629–640.

Biscaye, P. E., 1965. Mineralogy and sedimentation of Recent deep-sea clay in the Atlantic Ocean and adjacent seas and oceans. *Geol. Soc. Am. Bull.*, 76:803–832.

Bischoff, J. L., and Dickson, F. W., 1975. Seawater–basalt interaction at 200°C and 500 bars: implications for origin of sea-floor heavy metal deposits and regulation of seawater chemistry. *Earth Planet. Sci. Lett.*, 25:385–397.

Boles, J. R., and Wise, W. S., 1978. Nature and origin of deep-sea clinoptilolite. *In* Sand, L. B., and Mumpton, F. A. (Eds.), *Natural Zeolites: Occurrence, Properties, Use:* Elmsford (Pergamon Press), pp. 235–243.

Bonatti, E., 1967. Mechanisms of deep-sea volcanism in the South Pacific. *In* Abelson, P. H. (Ed.), *Researches in Geochemistry* (Vol. 2): New York (Wiley), 543–491.

Bonatti, E., Kraemer, T., and Rydell, H., 1973. Classification and genesis of submarine iron-manganese deposits. *In* Horn, D. (Ed.), *Ferromanganese Deposits on the Ocean Floor:* Washington (National Sci. Found.), pp. 146–166.

Boström, K., 1973. The origin and fate of ferromanganoan active-ridge sediments. *Stockholm Contr. Geol.*, 27:149–243.

Boström, K., Kraemer, T., and Gartner, S., 1973. Provenance and accumulation rates of opaline silica, Al, Ti, Fe, Mn, Cu, Ni and Co in Pacific pelagic sediments. *Chem. Geol.*, 11:123–148.

Boström, K., and Peterson, M. N. A., 1966. Precipitates from hydrothermal exhalations on the East Pacific Rise. *Econ. Geol.*, 61: 1258–1265.

_______, 1969. The origin of aluminum-poor ferromanganoan sediments in areas of high heat flow on the East Pacific Rise. *Mar. Geol.*, 7:427–447.

Bowles, F. A., Angino, E. E., Hosterman, J. W., and Galle, O. K., 1971. Precipitation of deep sea palygorskite and sepiolite. *Earth Planet. Sci. Lett.*, 11:324–332.

Cann, J. R., 1971. Major element variations in ocean-floor basalts. *Phil. Trans. Roy. Soc. London, Ser. A.*, 268:495–505.

Cosgrove, M. E., and Papavassiliou, C. Th., 1979. Clinoptilolite in DSDP sediments of the Indian Ocean (Site 223, leg 23): its stability conditions and estimation of its free energy. *Mar. Geol.*, 33: M77–84.

Couture, R. A., 1977. Composition and origin of palygorskite-rich and montmorillonite-rich zeolite-containing sediments from the Pacific Ocean. *Chem. Geol.*, 19:113–130.

Cronan, D. S., 1969. Average abundances of Mn, Fe, Ni, Co, Cu, Pb, Mo,V, Cr, Ti and P in Pacific pelagic clays. *Geochim. Cosmochim. Acta*, 33:1562–1565.

_______, 1976. Basal metalliferous sediments from the eastern Pacific. *Geol. Soc. Am. Bull.*, 87:928–934.

Dolcater, D. L., Syers, J. K., and Jackson, M. L., 1970. Titanium as free oxide and substituted forms in kaolinites and other clay minerals. *Clays Clay Minerals*, 18:71–79.

Drever, J. I., 1974. The magnesium problem. *In* Goldberg, E. D. (Ed.), *The Sea* (Vol. 5): New York (Wiley), 337–357.

Dymond, J., Corliss, J. B., Heath, G. R., et al., 1973. Origin of metalliferous sediments from the Pacific Ocean. *Geol. Soc. Am. Bull.*, 84:3355–3372.

Eberl, D., and Hower, J., 1977. The hydrothermal transformation of sodium and potassium smectite into mixed layer clay. *Clays Clay Minerals*, 25:215–227.

Elderfield, H., 1976. Hydrogenous material in marine sediments excluding manganese nodules. *In* Riley, J. P., and Chester, P. (Eds.), *Chemical Oceanography* (2nd ed., vol. 5): London (Academic Press), 137–215.

Eslinger, E. V., and Savin, S. M., 1976. Mineralogy and O^{18}/O^{16} ratios of fine-grained quartz and clay from Site 323. *In* Hollister, C. D., Craddock, C., et al., *Init. Repts. DSDP*, 35: Washington (U.S. Govt. Printing Office), 489–496.

Garrison, R. E., 1973. Space-time relations of pelagic limestones and volcanic rocks, Olympic Peninsula, Washington. *Geol. Soc. Am. Bull.*, 84:583–594.

Goldberg, E. D., and Arrhenius, G. O. S., 1958. Chemistry of Pacific pelagic sediments. *Geochim. Cosmochim. Acta*, 13:153–212.

Goldschmidt, V. M., 1954. *Geochemistry:* Oxford (Clarendon Press),

Griffin, J. J., Windom, H., and Goldberg, E. D., 1968. The distribution of clay minerals in the world ocean. *Deep-Sea Res.*, 15: 433–459.

Hart, R. A., 1973. A model for chemical exchange in the basalt-seawater system of oceanic layer II. *Can. J. Earth Sci.*, 10:799–816.

Huddle, J. W., and Patterson, S. H., 1961. Origin of Pennsylvanian underclay and related seat rocks. *Geol. Soc. Am. Bull.*, 72:1643–1660.

Jenkyns, H. C., 1976. Sediments and sedimentary history of the Manihiki Plateau, South Pacific Ocean. *In* Schlanger, S. O., Jackson, E. D., et al., *Init. Repts. DSDP*, 33: Washington (U.S. Govt. Printing Office), 873–890.

Jenkyns, H. C., and Hardy, R. G., 1976. Basal iron-titanium-rich sediments from Hole 315A (Line Islands, Central Pacific). *In* Schlanger, S. O., Jackson, E. D., et al., *Init. Repts. DSDP*, 33: Washington (U.S. Govt. Printing Office), 833–836.

Katsura, T., Kushiro, I., Akimoto, S., et al., 1962. Titanomagnetite and titanomaghemite in a Hawaiian soil. *J. Sediment. Petrol.*, 32:299–308.

Lancelot, Y., Hathaway, J. C., and Hollister, C. D., 1972. Lithology of sediments from the western North Atlantic Leg 11, Deep Sea Drilling Project. *In* Hollister, C. D., Ewing, J. I., et al., *Init. Repts. DSDP*, Π: Washington (U.S. Govt. Printing Office), 901–949.

Loughnan, F. C., 1969. *Chemical Weathering of the Silicate Minerals:* New York (Elsevier).

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.*, 268:551–571.

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.*, 84: 553–565.

Moberly, R., and Heath, G. R., 1971. Volcanic rocks from the Western and Central Pacific: Leg 7, Deep Sea Drilling Project. *In* Winterer, E. L., Riedel, W. R., et al., *Init. Repts. DSDP*, 7, Pt. 2: Washington (U.S. Govt. Printing Office), 1011–1025.

Moberly, R., and Keene, J. B., 1975. Origin and diagenesis of volcanic-rich sediments for North Pacific seamounts, DSDP Leg 32. *In* Larson, R. L., Moberly, R., et al., *Init. Repts. DSDP*, 32: Washington (U.S. Govt. Printing Office), 537–546.

Moore, J. G., 1975. Mechanism of formation of pillow lava. *Amer. Sci.*, 63:269–277.

Mottl, M. J., and Holland, H. D., 1978. Chemical exchange during hydrothermal alteration of basalt by seawater: I, Experimental results for minor elements. *Geochim. Cosmochim. Acta*, 42:1103–1115.

Natland, J. H., 1973. Basal ferromanganoan sediments at DSDP Site 183, Aleutian abyssal plain, and Site 192, Meiji Guyot, northwest Pacific, Leg 119. *In* Creager, J. S., Scholl, D. W., et al., *Init. Repts. DSDP*, 19: Washington (U.S. Govt. Printing Office), 629–640.

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*, 268:469–486.

Perry, E. A., Jr., 1976. Submarine smectite genesis as an oceanic magnesium sink. *In* Bailey, S. W. (Ed.), *Proceedings of the International Clay Conference, 1975, Mexico City:* Wilmette, Ill. (Applied Publishing), pp. 287–295.

Perry, E. A., Jr., Gieskes, J. M., and Lawrence, J. R., 1976. Mg, Ca and O^{18}/O^{16} exchange in the sediment-pore water system, Hole 149, DSDP. *Geochim. Cosmochim. Acta*, 40:413–423.

Peterson, M. N. A., and Goldberg, E. D., 1962. Feldspar distribution to South Pacific pelagic sediments. *J. Geophys. Res.*, 67:3477–3492.

Rankama, K., and Sahama, T. G., 1950. *Geochemistry:* Chicago Univ. Chicago Press).

Rateev, M. A., Gorbunova, Z. N., Lisitzyn, A. P., et al., 1969. The distribution of clay minerals in the oceans. *Sedimentol.*, 13:21–43.

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

Siesser, W. G., 1978. Native copper in DSDP Leg 40 sediments. *In* Bolli, H. M., Ryan, W. B. F., et al., *Init. Repts. DSDP*, Suppl. to Vols. 38, 39, 40, and 41: Washington (U.S. Govt. Printing Office), 761–765.

Spears, D. A., and Kanaris-Sotiriou, R., 1976. Titanium in some Carboniferous sediments from Great Britain. *Geochim. Cosmochim. Acta*, 40:345–351.

Spooner, E. T. C., and Fyfe, W. W., 1973. Sub-sea floor metamorphism, heat, and mass transfer. *Contr. Mineral. Petrol.*, 42:287–304.

von Rad, U., and Rösch, H., 1972. Mineralogy and origin of the clay minerals, silica and authigenic silicates in Leg 14 sediments. *In* Hayes, D. E., Pimm, A. C., et al., *Init. Repts. DSDP*, 14: Washington (U.S. Govt. Printing Office), 727–751.

Walker, J. L., Sherman, G. D., and Katsura, T., 1969. The iron and titanium minerals in the titaniferous ferruginous latosols of Hawaii. *Pacific Sci.*, 23:291–304.

Watts, A. B., Bodine, J. H., and Ribe, N. M., 1980. Observations of flexure and the geological evolution of the Pacific Ocean Basin. *Nature*, 283:532–537.

Williams, H., and McBirney, A. R., 1979. *Volcanology:* San Francisco (Freeman, Cooper).