LATE MIOCENE HIATUSES AND RELATED EVENTS IN THE CENTRAL EQUATORIAL PACIFIC: THEIR DEPOSITIONAL IMPRINT AND PALEOCEANOGRAPHIC IMPLICATIONS

A DISSERTATION SUBMITTED TO THE GRADUATE DIVISION OF THE UNIVERSITY OF HAWAII IN PARTIAL FULFILLMENT OF THE REQUIREMENTS FOR THE DEGREE OF DOCTOR OF PHILOSOPHY IN GEOLOGY AND GEOPHYSICS AUGUST 1987

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We certify that we have read this dissertation and that in our opinion it is satisfactory in scope and quality as a dissertation for the degree of Doctor of Philosophy in Geology and Geophysics.

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ACKNOWLEDGMENTS

I wish to express my sincere appreciation to my committee for their support, guidance, and interest throughout this dissertation project. Foremost, I am grateful to my advisor, Fritz Theyer. Without his support, inspiration, and encouragement, my research would not have been possible. I am thankful for his hospitality in Los Angeles and I also appreciate his release of yet unpublished rock-magnetic data. Jane Schoonmaker agreed to serve as an "Acting Advisor" at a time when she was certainly not aware of what she was getting into. I am very grateful to her, not only for all of her editing and her assistance in the XRD work, but also for her genuine interest in my work. My warmest thanks are extended to Fred Mackenzie for his unselfish support and interest in the progress of this study. His understanding and advice was especially welcomed at certain times when the going got really tough. Many thanks to Dave Epp, who unexpectedly came back, just in time to help me get things done. Furthermore, I am very grateful to Alex Malahoff, who enthusiastically joined my committee after Stan Margolis (who, nevertheless, kept his interest in my dissertation). Last but certainly not least, I am thankful to Peter Muller, with whom I have had many discussions about, and beyond, my dissertation topic.
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ABSTRACT

Most studies of Cenozoic hiatuses in deep-sea sediments have been restricted to stratigraphic problems in regions where sedimentation rates are relatively high, carbonate deposition prevails, or the erosional activity of bottom currents had been previously established. This dissertation presents a study of hiatuses recorded in predominantly siliceous, slowly deposited abyssal sequences and concentrates on changes in sedimentological properties across the hiatuses. In Middle-Upper Miocene piston cores K78-5-10 (7.4° N, 169.6° W) and M70-17 (7.5° S, 161.6° W) hiatuses have been identified which, on the basis of integrated radiolarian, diatom, and magnetostratigraphic data, correlate with hiatuses NH3(NH4?), NH5, and NH6 (about 12.6-10.4, 8.9-7.9, and 7.1-6.3 Ma). Both sites lie in the Central Basin of the Equatorial Pacific, below the carbonate compensation depth (5475 m and 4721 m) where sedimentation is dominated by siliceous brown clays that contain radiolarians, authigenic minerals, and volcanic debris. The hiatus intervals occur within the Diartus petterssoni, Didymocyrtis antepenultima, and Didymocyrtis penultima radiolarian zones and are associated with distinct textural, mineralogical, and sedimentological variations detected in X-radiographs, X-ray diffraction analyses, silt-size
particle analyses, and compositional analyses of the coarse fraction. Further evidence for the occurrence of distinct depositional changes related to the hiatuses is derived from anisotropy of magnetic susceptibility (AMS) measurements and magnetic grain-size analyses. Intensification of bottom currents seems to be the immediate cause for hiatuses in deep-sea sediments in the Central Basin of the Equatorial Pacific. A comparison of NH6 occurrence with graphic correlation plots of selected DSDP Sites, a recently revised global coastal onlap curve, and stable oxygen and carbon isotope curves, however, points to relative sea-level changes as a potential trigger mechanism. In particular, the data support a relationship between hiatus occurrence and relative sea-level highstands, followed by coastal offlaps, interpreted as indicating falling sea level, and correlating with the cessation of hiatus intervals. Based on data associated with the occurrence of hiatus NH6, a speculative, sea-level dominated depositional model is outlined, which explains the occurrence of marine hiatuses in various depositional environments, regardless of water depth.
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<th>Definition</th>
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<tbody>
<tr>
<td>AABW</td>
<td>Antarctic Bottom Water</td>
</tr>
<tr>
<td>AC</td>
<td>Alternating Current</td>
</tr>
<tr>
<td>AMS</td>
<td>Anisotropy of Magnetic Susceptibility</td>
</tr>
<tr>
<td>ARM</td>
<td>Anhysteretic Remanent Magnetism ((K_{ARM}))</td>
</tr>
<tr>
<td>B</td>
<td>Bottom</td>
</tr>
<tr>
<td>CCD</td>
<td>Carbonate Compensation Depth</td>
</tr>
<tr>
<td>DC</td>
<td>Direct Current</td>
</tr>
<tr>
<td>DSDP</td>
<td>Deep Sea Drilling Project</td>
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<tr>
<td>F</td>
<td>Magnetic Foliation</td>
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<tr>
<td>f</td>
<td>Foliation Difference</td>
</tr>
<tr>
<td>H</td>
<td>Total Anisotropy</td>
</tr>
<tr>
<td>HEBBLE</td>
<td>High Energy Benthic Boundary Layer Experiment</td>
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<tr>
<td>IRM</td>
<td>Isothermal Remanent Magnetism</td>
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<td>(l )</td>
<td>Lineation Difference</td>
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<tr>
<td>NH</td>
<td>Neogene Hiatus</td>
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<tr>
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<td>North Atlantic Deep Water</td>
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<td>NRM</td>
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<tr>
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<td>Mass Susceptibility</td>
</tr>
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<td>XRD</td>
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<td>(\bigcirc)</td>
<td>(1) Diatom Datum; (2) Minimum Susceptibility Direction</td>
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<td>(\triangle)</td>
<td>(1) Radiolarian Datum; (2) Intermediate Susceptibility Direction</td>
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<td>(\square)</td>
<td>(1) Polarity-Change Datum; (2) Maximum Susceptibility Direction</td>
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The intensive study of the ocean basins during the past fifteen years (mainly because of the outstandingly successful Deep Sea Drilling Project) disproved the idea that the deep sea is a final and uniform depositional reservoir (e.g., Johnson and Johnson, 1970; Johnson, 1972a; Watkins and Kennett, 1972; Kennett et al., 1972; Rona, 1973; Johnson, 1974; van Andel et al., 1975; Kennett, 1977; Moore et al., 1978; Ledbetter and Ellwood, 1980; Lonsdale and Smith, 1980; Schnitker, 1980; Lonsdale, 1981; Barron and Keller, 1982; Ledbetter and Ciesielski, 1982; Keller and Barron, 1983; Keller et al., 1987). Indeed, discontinuous oceanic sediment sections were found to be the rule, rather than the exception. Causes of hiatuses may be of local, regional, or global origin and they range, in broad terms, from changes in the productivity of surface waters to changes in the corrosiveness and dynamics of bottom waters. In turn, oceanographic oscillations, motions of plates (and associated openings or closing of passages), and climatic variations may trigger and control these immediate causes.

The term hiatus is used to define a gap or missing interval in the continuity of a stratigraphic sequence (= time not recorded by sedimentation), which is due to nondeposition, erosion, dissolution, extraordinarily slow
sedimentation rates, or a combination thereof. The term hiatus interval is used both in the sense of describing the duration of a hiatus (= time) and to indicate the core depth which contains an unconformity (= physical core length). The expression unconformity implies a surface of erosion, nondeposition, or dissolution, which separates younger from older strata and thus represents a definable hiatus. A barren zone in this study is understood to be a core interval devoid of diagnostic fossils and magnetostratigraphic data.

The distribution of deep-sea hiatuses seems to follow distinct patterns in time and space (e.g., Moore et al., 1978; Keller and Barron, 1983). Biostratigraphic analyses focussing on Miocene deep-sea hiatuses (Barron and Keller, 1982; Keller and Barron, 1983; Keller et al., 1987) resulted in the identification and cataloging of eight widespread hiatuses (Fig. 1) in the world ocean. The Neogene hiatuses were labeled "NH" and assigned numbers 1 through 7 upward from the Paleogene/Neogene boundary. Cool water faunal and floral assemblages and an enrichment of $\delta^{18}O$ in benthic foraminifers in corresponding intervals in complete sections suggest that a relationship exists between occurrence of hiatuses and global climatic cooling (Barron and Keller, 1982). A comparison with the 1979 version of the coastal onlap curve of Vail and Hardenbol led to a
<table>
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<th>ZONES</th>
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<tr>
<td>PLIOCENE</td>
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<tr>
<td>5</td>
<td>N18  A. tric</td>
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</tr>
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<td>6</td>
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<td>9</td>
<td>N15  D. hetero</td>
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</tr>
<tr>
<td>10</td>
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<td></td>
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<td>N12  D. kugler</td>
<td></td>
<td></td>
</tr>
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<td>13</td>
<td>N11  C. costata</td>
<td></td>
<td>NH3</td>
</tr>
<tr>
<td>14</td>
<td>N10  C. virgins</td>
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<tr>
<td>15</td>
<td>N9    D. alataa</td>
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<td>PH    D. alataa</td>
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correlation between hiatus occurrence and falling sea level (Barron and Keller, 1982).

All previous hiatus-oriented research has been conducted in high-productivity regions marked by high sedimentation rates and usually dominated by carbonate deposition. Similar studies in slowly-accumulating, predominantly siliceous, abyssal sequences are still outstanding. Any similarities or correlations between such different environments would indicate causative factors that transcend simple solution/erosion-type mechanisms. Most hiatus-oriented studies to date have concentrated on stratigraphic problems. To the author's knowledge, no mineralogical-sedimentological investigations (with the exception of grain-size analyses) have been conducted across known hiatus intervals in search of any depositional imprint. The use of rock-magnetic properties has usually been limited to depositional environments where the activity of bottom currents was well established, not just suspected (e.g., de Menocal et al., submitted; de Menocal and Laine, submitted).

In this dissertation, an attempt has been made to apply Barron and Keller's "hiatus stratigraphy" concept to siliceous brown clays of late Miocene age in the Central Basin of the Equatorial Pacific. The study of hiatuses is not only approached from a stratigraphic point of view, but
also by focussing on depositional imprints which might shed light on the origin of detected unconformities. Furthermore, comparisons with other data sets were conducted to investigate whether the occurrence of hiatuses in the Central Basin is a local phenomenon or if it can be linked to a global scenario. In particular, this study has addressed the following questions:

- What are the precise stratigraphic ages of Miocene unconformities in the Central Basin of the Pacific?

- Is there a stratigraphic and genetic relationship between unconformities and barren zones, as a tentative correlation suggested?

- What are the mineralogical, compositional, and textural changes of the sediments above and below the unconformities? That is, can they be linked to nondeposition or to erosion?

- Can analyses of the alignment of magnetic grains and other rock-magnetic properties elucidate the cause (causes?) of hiatuses?

- Are hiatuses in the deepest oceanic deposits (that accumulate at slow rates and are generally dominated by siliceous clayey sediments), such as those of the Central Basin, correlated with hiatuses or periods of
slow sedimentation in high-productivity areas (dominated by calcareous sediments), such as the east-central bulge of the Pacific?

- Is there a relationship between hiatus occurrence and relative sea-level changes?
II. CORES UNDER INVESTIGATION, THEIR PALEOCEANOGRAPHIC SETTING, AND THEIR STRATIGRAPHIES

1. Piston Cores K78-5-10 and M70-17

After preliminary inspections of magnetostratigraphies and radiolarian datums of more than 40 piston cores of the Hawaii Institute of Geophysics collection, cores K78-5-10 and M70-17 were chosen for study of late Miocene hiatuses. Selection of the two cores was based on the following reasons: (1) preliminary magnetostratigraphic and radiolarian datums were available; (2) initial inspection promised high stratigraphic resolution; (3) the cores covered approximately the same time interval; (4) both cores were from an isolated basin with well known bottom-water circulation; (5) hiatus-oriented studies in a clayey-siliceous environment, dominated by slow sedimentation rates, had not been conducted.

Both cores, K78-5-10 (7.4° N, 169.6° W; 5475 m water depth; 14.2 m recovery) and M70-17 (7.5° S, 161.6° W; 4721 m water depth; 11.5 m recovery), were collected in the Central Basin of the Equatorial Pacific (Fig. 2). Deposition at these sites, and in the Central Basin in general, occurs below the carbonate compensation depth (CCD), and is characterized by siliceous brown clays that contain
FIGURE 2. Location map showing positions of sites K78-5-10 and M70-17 in the Central Basin of the Equatorial Pacific. Arrows indicate the known bottom-current pattern. Contour lines indicate areas shoaler than 4.5 km. After Lonsdale (1981).
radiolarians, authigenic minerals (zeolites, manganese micronodules), and volcanic ash.

2. The Central Basin of the Equatorial Pacific

The Central Basin of the Pacific is a generally deep (>4500 m) and rather constricted feature (Fig. 2) bound by the Mid-Pacific Mountains to the north, the Line Island Ridge to the east, the Robbie Ridge and Manihiki Plateau to the south, and the Marshall-Gilbert-Ellice Ridge system to the west. These topographic highs serve as natural barriers to regional bottom-water flow. Owing to this general simplicity, and because of extensive hydrographic studies in this area (Stommel, 1958; Reid et al., 1968; Reid, 1969; Reed, 1969; Gordon and Gerard, 1970), as well as direct current measurements (Reid, 1969; Johnson and Johnson, 1970; Reid and Lonsdale, 1974), the region's abyssal circulation is well defined (Fig. 2).

Because intensification of bottom-current activity is one of the potential immediate causes for the occurrence of hiatuses, it is appropriate to briefly sketch the presently known current pattern of the Central Basin. The basin is supplied with bottom water by a western boundary current entering through the constricted Samoan Passage (Reid and Lonsdale, 1974; Hollister et al., 1974) between the Robbie Ridge and Manihiki Plateau. North of the exit of this
passage the flow of bottom water is less constrained by topography (Lonsdale, 1981) and the current diverges. Most of the bottom water apparently flows as a boundary current along the Marshall-Gilbert Ridge and leaves the basin through Wake Island Passage (Mantyla, 1975; Edmond et al., 1971; Lonsdale and Smith, 1980). An eastern branch flows along the southern flanks of the Mid-Pacific Seamounts, using a deep passage south of Horizon guyot, the so-called Horizon Passage, as an exit (Edmond et al., 1971). Some bottom water leaves the basin to the east through narrow breaks in the Line Island Ridge, such as the Clarion Passage (Mantyla, 1975) and the Kingman Passage (Normark and Spiess, 1976), and through the Clipperton Fracture Zone (Johnson, 1972b).

3. Integrated Stratigraphies of Cores K78-5-10 and M70-17

An integrated magneto-biostratigraphic approach, based on paleomagnetic, radiolarian, and diatom data (Tab. 1), was used to date cores K78-5-10 (Fig. 3) and M70-17 (Fig. 4). The magnetostratigraphy and radiolarian events of core M70-17 are discussed by Theyer and Hammond (1974) and Theyer et al. (1978). Polarity changes and radiolarian events of core K78-5-10 are from Theyer (unpubl. data). Critical core sections were resampled for radiolarians and diatoms at close intervals (10 cm) to enhance stratigraphic
TABLE 1. Paleomagnetic, radiolarian, and diatom datums used for stratigraphic analyses. Event numbers correspond with numbers assigned to symbols in the graphic correlation plots (Figs. 5, 31). References: (A) Kent and Gradstein (1987); (B) Barron et al. (1985); (C) Barron (pers. comm.); (D) Theyer et al. (1978), reinterpreted; (E) Mayer, Theyer, et al. (1985).
<table>
<thead>
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<th>Age (Ma)</th>
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<td>T</td>
<td>Sidufjall subchron (Gilbert chron)</td>
<td>4.1</td>
</tr>
<tr>
<td>2</td>
<td>B</td>
<td>Thvera subchron (Gilbert chron)</td>
<td>4.7</td>
</tr>
<tr>
<td>3</td>
<td>T</td>
<td>Thalassiosira miocenica</td>
<td>5.1</td>
</tr>
<tr>
<td>4</td>
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</tr>
<tr>
<td>5</td>
<td>T</td>
<td>Chron 5</td>
<td>5.3</td>
</tr>
<tr>
<td>6</td>
<td>T</td>
<td>Nitzschia miocenica</td>
<td>5.6</td>
</tr>
<tr>
<td>7</td>
<td>T</td>
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</tr>
<tr>
<td>8</td>
<td>T</td>
<td>Chron 6</td>
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</tr>
<tr>
<td>9</td>
<td>T</td>
<td>Stichocorys delmontensis</td>
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</tr>
<tr>
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<td>B</td>
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<td>12</td>
<td>B</td>
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<td>13</td>
<td>B</td>
<td>Stichocorys peregrina</td>
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<td>14</td>
<td>T</td>
<td>Chron 7</td>
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<td>T</td>
<td>Rossiella paleacea</td>
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<td>T</td>
<td>Coscinodiscus nodulifer var. cyclopus</td>
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<td>20</td>
<td>T</td>
<td>Actinocyclus ellipticus var. javanicus</td>
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<td>Long normal subchron (Chron 7)</td>
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<td>T</td>
<td>Chron 8</td>
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<td>Coscinodiscus plicatus</td>
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<td>26</td>
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<td>29</td>
<td>T</td>
<td>Coscinodiscus loeblichii</td>
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<tr>
<td>30</td>
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<td>Diartus Hughesi</td>
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<tr>
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<tr>
<td>33</td>
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<td>Chron 11</td>
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<td>Chron 15</td>
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<td>47</td>
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<td>49</td>
<td>B</td>
<td>Thalassiosira tappanae</td>
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B = bottom; T = top.
FIGURE 3. Magneto- and biostratigraphy of the studied interval of piston core K78-5-10. In the polarity log, black and white denote normal and reversed magnetization, respectively. Diatom datums are offset to the right of radiolarian datums. T and B indicate the top and bottom of the given microfossil zone.
K 78-5-10
7.4° N
169.6° W
WATER DEPTH: 5475 m

1000
T. miocenica

1050
T. miocenica

1100
T. praecoverxa

1150
T. delmontensis
B T. convexa var. aspinosa
B T. miocenica

1200
T. praecoverxa

1250
B N. miocenica var. elongata
B N. miocenica
T R. paleacea

1300
T C. plicatus
T C. yabei

1350
B A. tritubus
B D. antepenultima
B D. hughesi

1400
B T. burckliana

(cm)

T=top
B=bottom
FIGURE 4. Magneto- and biostratigraphy of the studied interval of piston core M70-17. In the polarity log, black and white denote normal and reversed magnetization, respectively; hachuring indicates an interval of disturbed paleomagnetic and biostratigraphic data which is referred to as "barren zone" in this study. Diatom datums are offset to the right of radiolarian datums. For details on core M70-17 and the paleomagnetic data see Theyer and Hammond (1974) and Theyer et al. (1978).
resolution. Diatoms were identified by Dr. John A. Barron from the United States Geological Survey.

Radiolarian and diatom stratigraphies are based on datum planes or events. An event or datum is defined by the first ("B"/bottom) or last appearance ("T"/top) of consistently occurring specimens in a continuous morphological range. The radiolarian events follow the tropical zonations of Riedel and Sanfilippo (1978). Recognition and definition of the hiatuses were initially based on missing magnetic polarity changes, missing radiolarian datums, and the concentration of discontemporaneous radiolarian events. This combined approach allowed the detection of hiatuses within the established radiolarian zonations (Riedel and Sanfilippo, 1978). Hiatuses NH3, NH5, and NH6 were thus determined to occur within the Diartus petterssoni, Didymocyrtis antepenultima, and Didymocyrtis penultima radiolarian zones, respectively.

Radiolarian, diatom, and magnetic polarity-change datums (Tab. 1) are based on a recently revised geochronology (Berggren et al., 1985; Barron et al., 1985). Stratigraphies in previous publications (Theyer and Hammond, 1974; Theyer et al., 1978) were based on a correlation of Chron 9 with Chron C5. The presently favored correlation of Chron 11 with Chron C5 (Berggren, 1985) results in an
approximately 1.5-2.0 Ma shift towards younger datum planes over part of the middle and late Miocene interval. Thus a recalibration of certain stratigraphic events was necessary. Datum planes affected by the recalibration of core M70-17 might not be consistent with estimates from previous publications (Theyer and Hammond, 1974; Theyer et al., 1978).

Graphic correlation plots (Shaw, 1964) for each core (Fig. 5), based on the integrated datums (Tab. 1), bracket the duration of NH6 between 7.1 and 6.3 Ma, NH5 between 8.9 and 7.9 Ma, and NH3 between 12.6 and 10.4 Ma. The extraordinarily long duration of NH3 in core M70-17 suggests a possible temporal overlap with NH4. The apparent duration of the hiatus intervals, however, does not necessarily reflect the duration of "events" which caused them. Age versus depth plots (Fig. 5) do not identify the occurrence of a "real" hiatus related to NH6 in core M70-17. Nevertheless, in this core, NH6 seems to correlate with a barren zone which is interpreted as a depositional disturbance caused by the same event which is responsible for the occurrence of NH6 in core K78-5-10.

Integration of multiple datums (Tab. 1) linked to high-resolution time scales (Barron et al., 1985; Berggren et al., 1985), and close sample spacing (averaging 10 cm) suggest a stratigraphic resolution of 0.1-0.3 Ma. The
FIGURE 5. Age versus depth plots of cores K78-5-10 and M70-17 using the graphic correlation technique of Shaw (1964). The plots are based on the datum events listed in Table 1, following the same number assignment. Squares identify paleomagnetic, triangles and circles radiolarian and diatom datums. Paleomagnetic time scale and chronology are from Berggren et al. (1985). Hiatus intervals are represented by shading and vertical wavy lines. Hiatuses are labeled after the classification of Barron and Keller (1982). Note that the hiatus event NH6 in core M70-17 is not recorded as a period of nondeposition.
importance of such a high resolving power, which allows the detection and correlation of relatively short-lived paleoceanographic events, cannot be overstated.
III. TEXTURAL, MINERALOGICAL, AND SEDIMENTOLOGICAL CHANGES ACROSS HIATUS INTERVALS

NH6, NH5, AND NH3

1. X-Radiography

1.1. Introduction

X-ray radiography, as applied in this study, uses the differential passage of X-radiation through sedimentary core halves recorded on a photographic film (Bouma, 1969). Density differences in the sediment control the amount of radiation that reaches the film and cause variances in photographic images. X-ray radiography is thus a simple and inexpensive method to detect subtle bedding features not generally found by surface inspection. It can also show the distribution of particles, such as manganese micronodules, which are generally associated with unconformities or low sedimentation rates (Glasby et al., 1982).

1.2. Methods

X-radiographs were taken of the archive halves (not previously sampled) of the cores under investigation. Because the sediment available was limited, the core-halves were not sliced, even though slices enhance X-ray image
sharpness and therefore resolution (Bouma, 1969). Nevertheless, because of the overall homogeneity of the present sediments, the results were satisfactory. A FAXITRON apparatus served as the X-ray source. The unit was operated in the manual exposure mode, with exposure times of 2-5 minutes at 60 kV, using ready-packed KODAK Industrex AA-2 film (each strip 0.4 m long). Overlapping X-radiographs were taken of the entire core sections (K78-5-10: 978-1350 cm; M70-17: 411-900 cm) under investigation. Each core interval was photographed at least twice to avoid possible misinterpretation owing to procedural artifacts (e.g., static marks or developer stains).

1.3. Results and Discussion

X-radiographs of cores M70-17 and K78-5-10 reveal noticeable textural changes only in association with hiatus intervals defined previously by stratigraphic analyses and the barren zone in core M70-17. The remainder of the sedimentary column is very homogeneous. This distinctive homogeneity is characteristic of Pacific deep-sea sediments (below the CCD) and was first reported by Calvert and Veevers (1962). Although the observed textural changes are subtle, and usually limited to intervals of <10 cm, the

- 24 -
overall homogeneity of the cores makes the changes clearly visible in the radiographs.

In core K78-5-10, the X-radiographs (Fig. 6) show textural changes only between 1249-1256 cm. This is the depth interval determined to contain hiatus NH6 (Fig. 5). Core K78-5-10 does not contain the older hiatuses NH5 and NH3. The top (495-515 cm) of the barren zone in core M70-17 (which is suspected to reflect the NH6 event) also reveals distinct textural changes. Within this barren zone, the X-radiographs show an increase in size and abundance of coarse particles at 530 cm. These features are discussed later. The M70-17 core interval containing hiatus NH5 shows subtle textural changes between 790-795 cm (Fig. 7). In contrast to the textural changes associated with the above hiatuses, hiatus NH3 in M70-17 is linked to a distinct and coarse-grained volcanic ash layer occurring at 872-876 cm. Despite this apparently clear-cut definition in macroscopic terms, X-radiographs of this ash layer reveal that it has been reworked and components of this layer can be traced in the radiographs up to the lower boundary of hiatus interval NH5 at 793 cm. If these coarse components are in fact reworked from the ash layer near NH3, then the apparently abrupt cessation of reworking at 793 cm provides further evidence that an interruption of sedimentation occurred during hiatus interval NH5 at the location of core M70-17.
FIGURE 6. X-radiograph taken across hiatus interval NH6 in core K78-5-10. Arrows mark the top and bottom of textural changes between approximately 1248 and 1257 cm.
FIGURE 7. X-radiograph taken across hiatus interval NH5 in core M70-17. Arrows mark the top and bottom of textural changes between approximately 789 and 797 cm.
The textural changes observed in the radiographs can be interpreted as unconformities and depositional disturbances. The images in the X-radiographs, however, do not furnish a causal explanation of their existence. Neither fine laminae, resembling ripple-mark structures, nor laminae resembling Bouma sequences (Bouma, 1969) were detected. Such structures were reported by Huang and Watkins (1977) for hiatus intervals in the South Pacific and interpreted as expressions of erosional processes. It is further impossible to detect evidence in the X-radiographs that these observed textural changes represent lag or residual deposits. In summary, although stratigraphically-defined hiatus intervals and barren zones are marked by textural changes, the cause for the occurrence of these hiatuses is not apparent from the X-radiographs.

2. Mineralogical Changes Associated with Hiatus Intervals

NH3, NH5, and NH6

2.1. Introduction

Changes in the mineral, especially clay-mineral, assemblages of deep-sea sediments are sensitive indicators of environmental changes (Heath, 1969; Jacobs and Hays, 1972). In particular, relative changes in the mineral content of deep-sea sediments reflect varying importance of
different sources and transport mechanisms. Thus, the mineralogical record across the hiatus intervals might contain information about the mechanism responsible for their origin.

2.2. Methods

Changes in the mineralogical composition across the defined hiatuses were determined by X-ray diffraction analysis after appropriate sample preparation (Fig. 8). The samples, taken with 6 cm³ plastic cubic boxes and previously used for magnetostratigraphic measurements, were dried at 40°C for 24 hours. To remove organic matter the samples were first treated with 5% NaOCl, buffered to pH 9.5 with HCl (Anderson, 1963). The samples were then washed with distilled water several times. Amorphous iron-oxyhydroxide-coatings were removed using the Na-citrate/Na-dithionite method of Mehra and Jackson (1960). Further rinsings were followed by a grain-size separation by repeated centrifugal decantations. This treatment produced two subfractions, one <2 micrometer, the other >2 micrometer.

Oriented slides were prepared from the stirred suspension of the fine fraction (<2 micrometer) and dried at room temperature. These slides were scanned from an angle of 2–32° 2θ at a rate of 1° 2θ/min on a Philips-Norelco
FIGURE 8. Summary diagram, showing stepwise sample preparation for X-ray diffraction analysis (XRD).
6 cm$^3$ sediment in NALGENE centrifuge tubes

ORGANIC DISSOLUTION:
Three treatments: 5% NaOCl (buffered to pH 9.5 with HCl) at 60°C for 20 min, with periodic stirring. Rinse three times with distilled H$_2$O.

AMORPHOUS IRON-OXYHYDROXIDE DISSOLUTION:
Three treatments: 20 ml 0.3 M Na-citrate and 2.5 ml of 1 M NaHCO$_3$ solution at 60°C with 0.5 g Na$_2$S$_2$O$_3$ for 15 min with constant stirring. Rinse three times with distilled H$_2$O.

SIZE SEPARATION:
Fill tubes to a height of 8 cm with distilled H$_2$O. Centrifuge at 1000 rpm for 45 sec, decant, and repeat if necessary.

<2 micrometer

XRD

Glycolate sample at 40°C for 24 hrs

XRD

>2 micrometer

Grind sample in mortar

XRD
X-ray diffractometer using Ni-filtered CuKα radiation. The slides were then placed in an ethylene-glycol atmosphere at 40°C for 24 hours to allow the detection of expandable clays. The scanning process was then repeated.

The coarse fraction (>2 micrometer) was ground with distilled water in a mortar, yielding a homogeneous suspension. From this suspension slides were prepared and analyzed as described above without glycolation.

Semiquantitative analysis of the mineral composition of the <2 micrometer fraction was based on weighted peak areas. Peak areas were calculated according to the "triangle method" of Mann and Fischer (1982), using the weighting factors of Mann and Muller (1979). The weighted areas of all clay-mineral peaks were summed and normalized to 100%. The combination peak near 12.4° 2θ for kaolinite and chlorite was used for the semiquantitative analysis. Because of the comparatively small amounts of kaolinite and chlorite, the relative contributions of these two minerals were not determined. Non clay-minerals, such as quartz and feldspars, present in all samples, and clinoptilolite, detected in a few samples, were omitted in these calculations.

Semiquantitative analysis of the coarse fraction (>2 micrometer) was based on peak heights rather than peak
areas. Using peak heights and factors determined by Mann and Muller (1979), the sum of the components feldspar, quartz, clinoptilolite, and cristobalite was normalized to 100%. Although present in almost all of the samples, the kaolinite/chlorite and illite peaks were omitted from these calculations.

2.3. Results

Profiles of the relative abundances of the major mineralogical components, as determined by XRD analysis, are presented (Figs. 9, 10) separately for the fine and coarse size fractions (<2 and >2 micrometer, respectively). The major constituents of the fine fraction are smectite, illite, and kaolinite/chlorite, with smectite being the most abundant. The coarse fraction consists of quartz, feldspar, cristobalite, and clinoptilolite. It is emphasized that the profiles show only relative changes in the mineralogical composition. Therefore, a relative increase of a certain component might be due to either an absolute increase in the abundance of that particular mineral or to decreases in other components ("dilution effect").

2.3.1. Mineralogical Composition of the Fine Fraction

Smectite is by far the most abundant mineral of the fine size fraction (Figs. 9, 10). The samples of core
FIGURE 9. Relative abundances of major mineralogical components (of both the fine and coarse fractions), textural changes (based on X-radiographs), and ratio plots of selected minerals of core K78-5-10. Feldspars (F), quartz (Q), and a combination of illite (ILL), kaolinite (K), and chlorite (CH), multiplied by an arbitrary scaling factor of 50, were chosen to represent the supposedly terrigenous influx. The minerals smectite (SM) and clinoptilolite (CL) are considered to be of "oceanic" origin. Cristobalite (CR) was not used for ratio plots. Shading indicates hiatus-related core interval.
K78-5-10
MINERALOGIC X-RADIOGRAPHY/COMPOSITION
TEXTURAL CHANGES

<2 µm >2 µm

0 100% 0 100%

0 1000 1050 1100 1150 1200 1250 1300 1350
(cm)

SM ILL K/CH
Q F CR CL

○ F/CL
□ Q/CL
△ (ILL + K + CH)*50/SM

- 37 -
FIGURE 10. Relative abundances of major mineralogical components (of both the fine and coarse fractions), textural changes (based on X-radiographs), and ratio plots of selected minerals of core M70-17. Shadings indicate hiatus-related core intervals. For legend see Fig. 9.
M70-17
MINERALOGIC X-RADIOGRAPHY
COMPOSITION TEXTURAL
CHANGES

<2 μm >2 μm
0 4 8 12 16 20 24

○ F/CL
□ Q/CL
△ (ILL + K + CH)*50/SM

SM
ILL
K/CH
Q
F
CR
CL

100%
100%

(cm)

39
K78-5-10 contain on the average 92% smectite (minimum: 85%; maximum: 97%). Samples of core M70-17 consist of comparatively less smectite: the average per sample is 86%, with 68% being the lowest and 95% the highest value. The three other clay minerals detectable by XRD analysis in this size fraction are illite and a combination of kaolinite/chlorite. In core K78-5-10, on the average, illite is present (Fig. 9) in quantities twice as great as kaolinite/chlorite together. These findings correspond well with data of Jacobs and Hays (1972) for clays from the equatorial Pacific. In core M70-17 (Fig. 10) the difference in relative abundances of illite and kaolinite/chlorite is less pronounced than in K78-5-10.

Despite the overwhelming contribution of smectite to the mineralogical composition of the fine fraction (<2 micrometer), distinct variations across the defined hiatus intervals occur in both cores (Figs. 9, 10). In general terms, hiatus intervals NH6, NH5, and NH3 are marked by a relative decrease in the abundance of smectite and a relative increase in illite. More specifically, in core K78-5-10 this relative decrease in smectite occurs at 1250 cm, the depth at which hiatus NH6 was found. Compared to samples taken from above and below this level, the relative amount of smectite drops from 92 to 86% and the proportion of illite increases from 6 to 10%. Across the same hiatus
interval kaolinite/chlorite combined show a relative increase in abundance from approximately 2 to 4%. The profile of the mineralogical composition of this core (Fig. 9) indicates that a gradual increase of the proportions of illite and kaolinite/chlorite at the expense of smectite started at a depth of 1370 cm. This trend ends at 1230 cm and seems to have culminated at 1250 cm.

Similar mineralogical changes occur across the depositional disturbances (which seem to reflect the NH6 event) in core M70-17 (Fig. 10): at a depth of 500 cm the percentage of smectite drops from, on the average, 85 to 73%. The relative proportion of illite increases from approximately 9 to 14%, and that of kaolinite/chlorite from 6 to 13%. Although the decrease in smectite abundance at 500 cm is very pronounced, it seems to be part of a general trend, which begins at a depth of approximately 570 cm. From this level up to the last analyzed sample (430 cm), smectite shows a gradual decrease relative to illite and kaolinite/chlorite. Although less pronounced, these shifts in the mineralogical composition of the fine fraction were also detected across the older hiatus interval of NH5 in core M70-17. At 790 cm smectite decreases from, on the average, 94 to 89%, illite increases from 3 to 5%, and kaolinite/chlorite from 4 to 7%.

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Hiatus NH3 is related to a well defined coarse volcanic
ash layer between 872 and 876 cm in core M70-17 (Fig. 10).
Although no abrupt mineralogical changes occur across this
hiatus level in the fine fraction (<2 micrometer), the ash
layer can be linked to an interval of relative decrease in
the abundance of smectite. This interval seems to begin at
a depth of 890 cm and ends at 820 cm. This interpretation
is arguable, however, because the sample at 900 cm was the
last one measured in core M70-17. Between 890 and 820 cm the
portion of smectite decreases from approximately 89 to 81%.
In the same depth interval illite increases from 8 to about
13%, and kaolinite/chlorite from 3 to 7%.

2.3.2. Mineralogical Composition of the Coarse Fraction

In both cores quartz, feldspar, cristobalite, and
clinoptilolite are the major mineralogical components of the
coarse size fraction (>2 micrometer) (Figs. 9, 10). Quartz
and feldspar are relatively more abundant than cristobalite
and clinoptilolite. The two cores differ fundamentally,
however, in the relative abundances of these minerals. Core
K78-5-10 has, on the average, a much higher proportion of
quartz: 34%, compared to 11% in core M70-17. In contrast,
M70-17 contains proportionally more feldspar: 59%, compared
to 44%.
In the >2 micrometer fraction the mineralogical changes are more complex than those of the fine fraction. A profile across the hiatus interval NH6 in core K78-5-10 (Fig. 9) reveals a relative decrease in the abundance of clinoptilolite (and to a lesser extent of cristobalite) between 1240 and 1250 cm. Quartz and, especially, feldspar increase proportionally. An analysis of the barren zone in core M70-17 leads to opposite findings (Fig. 10): a relative increase in clinoptilolite between 490 and 500 cm, and again at 540 cm, is coupled with decreases in the quartz and feldspar abundances. In core M70-17, between 780 and 800 cm there is a relative decrease in the abundance of clinoptilolite which coincides in depth with hiatus interval NH5 (Fig. 10). This decrease in clinoptilolite is associated with an increase in feldspar and, again, a decrease in quartz. Similar changes were detected across the interval of NH3. Clinoptilolite and quartz decrease in relative abundance at 870 cm, and feldspar increases.

Core K78-5-10 has a rather uniform quartz content. The coarse fraction data for core M70-17, however, reveal a gradual decrease in quartz from 900 cm (the bottom of the sampled section) up to 430 cm. The mineral cristobalite follows this trend. The diagenetic relationship between quartz and cristobalite, generally assumed to be of a positive correlation (Riech and von Rad, 1979), might
explain the striking similarity between the changing abundances of both minerals. This does not necessarily mean that the observed downcore increase in quartz is primarily a diagenetic effect. Diagenetic transformations from biogenic silica and cristobalite to quartz in sediments of relatively young age and shallow burial depth are unlikely (Riech and von Rad, 1979). The occurrence of authigenic silicate phases, however, might be due to reworking processes. The relative decrease in quartz and cristobalite between 470 and 600 cm is more likely due to an unrelated and sudden increase in clinoptilolite across this interval.

2.4. Discussion of Results

Both cores consist of the same mineral assemblages, although their relative abundances differ (Figs. 9, 10). The origin of these minerals in deep-sea sediments is discussed by Peterson and Goldberg (1962), Heath (1969), Jacobs and Hays (1972), and Kennett (1982). The relatively higher quartz content in core K78-5-10 (Fig. 9) is a result of the core sites' different latitudes: core K78-5-10 was taken 7.4° north, and core M70-17 7.5° south, of the equator (Fig. 2). It has been well established that sediments south of the equator lack the high quartz concentrations of northern samples (Heath, 1969). It is also known that in samples collected north of the equator, quartz and
plagioclase contents covary (Heath, 1969). In contrast, the plagioclase content in cores taken south of the equator does not vary systematically with the quartz content (Heath, 1969). These findings are consistent with the observation that in core K78-5-10 the quartz content varies parallel to the feldspar content, but not in core M70-17 (south of the equator).

The cores' mineral assemblages can be classified in two categories (Heath, 1969): (1) a "continental" mineral association which is characterized by abundant quartz and illite and the presence of potash feldspar (2) an "oceanic" mineral association which is characterized by the abundance of smectite and plagioclase, with zeolites commonly present. The present study did not, however, distinguish between different feldspar varieties. The "continental" mineral association is derived from detrital matter as a result of weathering on the continents. The "oceanic" mineral assemblage apparently consists of alteration products from volcanic debris generated within the ocean basin (Heath, 1969).

Changes in the relative abundances of "continental" and "oceanic" components should reflect environmental changes on the continents and/or in the ocean basins. To evaluate these assumptions, ratios were calculated of "terrigenous" versus "oceanic" mineral assemblages (Figs. 9, 10). In the
sediment's coarse fraction, quartz and feldspar were selected to represent the terrigenous influx. A portion of these minerals might be of volcanic origin, however, even though oceanically-derived quartz apparently occurs only along the crest of the East Pacific Rise in the Equatorial Pacific (Peterson and Goldberg, 1962). The zeolite clinoptilolite was used to represent the oceanic influx of the coarse fraction. In the fine fraction the ratio of illite, kaolinite, and chlorite combined ("terrigenous") versus smectite ("oceanic") was used, multiplied by a factor of 50 (the latter was chosen arbitrarily to enable the presentation of all ratios in the same graph). The results (Figs. 9, 10) for each separate hiatus interval are discussed below.

2.4.1. NH6 - K78-5-10

Both grain-size fractions in core K78-5-10 reveal a relative increase in the abundance of terrigenous minerals across the hiatus interval of NH6 (Fig. 9). If this increase occurred only in one size fraction, it could be argued that the apparent increase in continental influx arises from a reduced supply of oceanically derived material in the same fraction. A parallel increase in the terrigenous portions of both size fractions, however, makes this interpretation less likely. The shift to a relatively
higher influx of continental minerals is well documented in the ratio plots (Fig. 9).

In core K78-5-10 mineralogical changes related to hiatus NH6 seem to have been anticipated in the fine size fraction. In this fraction the trend towards a relatively higher portion of terrigenous minerals begins at a depth of 1300 cm (Fig. 9). If this shift can be interpreted as the result of an increase in terrigenous influx owing to an increase in weathering on the continents (and, presumably, an intensification of the wind system), then the fine size fraction should be affected first. Therefore, an assumed increase in continental weathering resulting in greater supply of terrigenous debris to the deep sea seems to have intensified with time: after first affecting the mineralogical composition of the fine size fraction, coarser particles reached the Central Basin and led to changes in the composition of the coarse size fraction (>2 micrometer). This explanation agrees with the findings of Janecek (1985) who reports large increases in eolian grain size from mid to late Miocene in the Northwest Pacific Ocean. He interprets this coarsening as a response to steepening thermal gradients resulting from increasing polar isolation.

Another relative increase upwards in the amount of terrigenous minerals begins in the fine fraction (<2
micrometer) at 1070 cm in K78-5-10 (Fig. 9). It is followed by a comparable trend in the coarse fraction (>2 micrometer) of the last two samples analyzed above 1020 cm in core K78-5-10. The mineralogical changes in the fine fraction occur across the Chron 5/Gilbert boundary (Fig. 1) and might precede the occurrence of hiatus interval NH7 (Barron and Keller, 1982). This speculation is supported by the occurrence of hiatus NH7 in sediments of Gilbert Chron age (Barron and Keller, 1982), and by similar mineralogical changes found in this age interval in core RC12-65 from the Equatorial Pacific (Jacobs and Hays, 1972).

2.4.2. NH6(?) - M70-17 (Barren Zone)

The depositional disturbances across the barren zone (between approximately 490 and 600 cm) which seems to reflect the NH6 event in core M70-17 are related to different mineralogical changes (Fig. 10) than those found in core K78-5-10 (Fig. 9). The fine size fraction shows a relative increase in the abundance of terrigenous components above 680 cm similar to changes in core K78-5-10. The coarse size fraction, however, records a rapid increase above 600 cm in the relative amount of clinoptilolite, that is, of an "oceanic" component as discussed earlier. In contrast, hiatus NH6 of core K78-5-10 was marked by a pronounced relative decrease in the abundance of clinoptilolite.
Clinoptilolite is a high-silica member of the heulandite group of zeolites. This mineral is of authigenic origin and commonly associated with siliceous tephra (Hay, 1978). It can also be found in apparently non-volcanogenic pelagic clays which contain biogenic silica and are rich in opal-CT (von Rad and Rosch, 1972; Kastner and Stonecipher, 1978). Furthermore, clinoptilolite might be an alteration product of the zeolite phillipsite (Boles and Wise, 1978). Neither phillipsite, nor any other zeolites besides clinoptilolite were detected in the samples analyzed by XRD. This is surprising, because the cores were taken in depositional environments with low sedimentation rates and below the CCD, in which one would expect to find higher abundances of phillipsite (Kastner and Stonecipher, 1978). In the Pacific Ocean, clinoptilolite is known to be more abundant than phillipsite only in marginal areas, rich in carbonate, where sedimentation rates are much higher (Kastner and Stonecipher, 1978). The observation that phillipsite is seemingly limited to surface samples and becomes unstable after burial might explain the absence of this zeolite in the analyzed samples (Kastner, pers. comm.).

In contrast to phillipsite (Bernat and Church, 1978), clinoptilolite does not seem to form at the water/sediment interface (Kastner and Stonecipher, 1978). In addition,
regardless of its precursor, authigenic clinoptilolite is usually found in sediments of at least early Pliocene (Kastner and Stonecipher, 1978) or, more likely, middle Miocene (von Rad and Rosch, 1972) age. Clinoptilolite is most abundant in Eocene and Cretaceous samples (Boles and Wise, 1978). The formation of clinoptilolite apparently requires a minimum burial depth of approximately 120 m (von Rad and Rosch, 1972), although in carbonate-rich sediments, clinoptilolite has been found from 20-100 m below the sediment surface (Kastner and Stonecipher, 1978). Clinoptilolite reported in Pleistocene sediments of DSDP cores at depths of less than 20 m subbottom, is probably redeposited (Kastner and Stonecipher, 1978). Iijima (1978) also suggests that the above mentioned shallow occurrences of deep-sea clinoptilolite might be caused by submarine erosion and reworking. At DSDP Site 478 in the Guyamas Basin very small amounts of clinoptilolite appear at a subbottom depth of 2.5 m (Kastner, 1982). Sedimentation at Site 478, however, is influenced by hydrothermal activity. There is no indication of thermal alterations at site M70-17.

Although age appears to be more important than actual burial depth (von Rad and Rosch, 1972), it is very unlikely that the occurrence of clinoptilolite at subbottom depths of less than 15 m can be explained by autochthonous growth. It
is thus suggested, especially in the absence of any detectable phillipsite, that the clinoptilolite found in the coarse-size fraction of both cores is reworked and of allochthonous origin. An analysis of drift deposits west of the Manihiki Plateau and in the vicinity of site M70-17 revealed that they consist of slowly-accumulated radiolarian and zeolitic clays, which are rich in clay aggregates and manganese micronodules (Hollister et al., 1974; Lonsdale, 1981). Phillipsite was detected in these drift deposits, but only as rare and broken crystals, which are believed to be redeposited. Lonsdale's (1981) samples were collected north of the Samoan Passage and thus downstream of a known inlet of Antarctic Bottom Water (AABW) into the Central Basin (Fig. 2). The sedimentation of these drift deposits is apparently controlled by fluctuations in the strength of currents (Lonsdale, 1981).

The depth level at which clinoptilolite is most abundant in core M70-17 does not specifically coincide with textural changes detected by X-radiographic and stratigraphic analyses (Fig. 10). If, as argued above, the relative increase in the abundance of clinoptilolite reflects an intensification of bottom current activity, it can be speculated that after a period of deposition of reworked sediments a threshold may have been reached after which active erosion began. This shift from reworking to
erosion could have been the result of an increase in bottom-current velocity and/or lateral movements of the current axis. Lonsdale (1981) describes a late Miocene hiatus in the aforementioned drift deposits near the site of core M70-17 (Fig. 2). Off the main current axis, this late Miocene interval in some of Lonsdale's cores is represented by zeolitic and, presumably, reworked clays. Reworking and drift deposition do also explain why the barren zone in core M70-17, which seems to reflect the NH6 event, does not reveal a decrease in sedimentation rates (Fig. 5).

2.4.3. NH5 - M70-17

Hiatus NH5 in core M70-17 displays similar mineralogical changes (Fig. 9) as NH6 in core K78-5-10 (Fig. 10). Both hiatus intervals are marked by a relative increase in the abundance of apparently terrigenous material in both size fractions. Despite these similarities, the same interpretations cannot be easily used to explain the two hiatuses. Across hiatus interval NH6 in core K78-5-10, a relative increase in the abundance of quartz and feldspar was detected in the coarse size fraction. Conversely, mineralogical analyses across the interval of NH5 in core M70-17 reveal a relative increase in the abundance of feldspar only. These findings, however, are consistent with the latitudinal differences between the cores. As mentioned
earlier, south of the equator in the Pacific the concentrations of feldspar, especially plagioclase, do not show systematic variations with quartz content, but they do north of the equator (Heath, 1969). In addition, sediments south of the equator have lower quartz concentrations than northern samples. These latitudinal differences complicate a comparison of the XRD data of both cores. It is possible that the detected mineralogical changes (Fig. 10) over the interval of NH5 in core M70-17 are the result of increased weathering and a related increase in wind-blown debris, similar to hiatus interval NH6 in core K78-5-10. The apparent relative decrease in the abundance of quartz across hiatus NH5 might be explained by an overprint owing to a synchronous increase in volcanogenically derived feldspar. This interpretation is supported by the observation that mineralogical analyses across hiatus interval NH3 in this core (Fig. 10), which contains a coarse volcanic ash layer, reveal similar results. Thus, hiatus NH5 in M70-17 seems to correspond to an increase in terrigenous material in the fine size-fraction, probably associated with an increase in volcanogenically derived feldspar in the coarse size-fraction. One cannot, however, exclude the possibility that some of these feldspars are reworked and therefore of allochthonous origin.
Just as with hiatus interval NH5, mineralogical changes across NH3 in this core reveal a relative increase in the abundance of terrigenous components (Fig. 10). The coarse size fraction is characterized by a decrease in "oceanically" derived minerals, namely cristobalite and clinoptilolite. The behavior of quartz in the coarse size fraction across this hiatus is not very clear: the relative decrease in the two "oceanic" minerals is accompanied by a relative increase in the abundance of quartz and feldspar. The relative increase in quartz is very slight and is followed upcore by a relative decrease in this mineral. The fact that hiatus NH3, in contrast with the other hiatus intervals, is linked to a coarse volcanic ash layer complicates the interpretation of the mineralogical data.

3. Quantitative Particle Analyses

3.1. Compositional Analyses of the Sand Fractions

X-ray diffraction analyses detect only crystalline substances. In addition to minerals, pelagic brown clays consist of many other components, e.g. radiolarian tests, manganese micronodules, volcanic glass, broken tests and other biogenic detritus, like fish teeth. Thus, relative
abundances of minerals do not reflect variations in the total particle population of a sample. Compositional analyses are required to determine variations in the bulk particle content.

3.1.1. Methods

Samples used for radiolarian stratigraphy (>63 micrometer) were dry-sieved through a 212 micrometer mesh. The resulting >212 micrometer subfractions were analyzed for compositional variations. This fraction was chosen because it allowed quick identification and counting of major components. Strewn particles were counted, grid by grid, in a sample tray, using a microscope and reflected light. Because samples were not split, all major components were counted. The total particle count differed from sample to sample, partially because the samples sieved were not of uniform volume. Major components of this size fraction were radiolarian tests, manganese micronodules (together with fragments), and biogenic detritus (like broken tests, fish teeth). Other accessory components included volcanic ash, zeolites, and feldspar fragments. Calculations of relative abundances were solely based on the major components.
3.1.2. Results

Variations of the major components across the barren zone of core M70-17 were plotted with depth (Fig. 11). The plot reveals a distinct inverse relationship between the relative abundance of radiolarian tests and manganese micronodules: increases in manganese micronodule percentage correlate with decreases in radiolarian tests. A major increase in the abundance of manganese micronodules occurs between 495 and 570 cm, just below textural changes detected by X-radiography. The relative increase in manganese micronodules, seemingly gradual upcore and downcore, culminates between 515 and 545 cm. This core interval does not contain any radiolarian tests. The absence of radiolarians is preceded (both upcore and downcore) by a shift in the radiolarian population towards more robust and resistant species. A second increase in manganese micronodules occurs in a single, radiolarian-free, sample at 465 cm.

Macroscopic inspection of the size fraction >212 micrometer across hiatus interval NH6 of core K78-5-10 did not reveal any compositional changes. This observation is mainly due to a higher abundance of dissolution-resistant radiolarian tests across this hiatus interval, as compared to NH6 of core M70-17. The relative higher percentage of
FIGURE 11. Compositional analysis of major components of the coarse size fraction (>212 micrometer) across the NH6-related barren zone of core M70-17. Hachuring of the magnetic polarity log indicates an interval of disturbed paleomagnetic and biostratigraphic data.
M70-17
(>212 µm)

Mag. Polarity

Micromanganese nodules
Biogenic detritus
Radiolarians

[Diagram with vertical axis labeled in cm and horizontal axis labeled 0 to 100%]
radiolarian tests in core K78-5-10 masks changes in manganese micronodule abundance and did not allow a meaningful compositional analysis.

3.2. Particle-Size Analysis of the Sediments' Silt Fractions

3.2.1. Introduction

One of the most favored explanations for hiatuses in pelagic sediments is erosion by bottom currents (e.g., Johnson, 1972; Watkins and Kennett, 1972; Huang and Watkins, 1977; Johnson et al., 1977; Barron and Keller, 1982; Ledbetter and Ciesielski, 1982). The particle-size distribution reflects textural and compositional characteristics of the sediment. It is also a sensitive and important indicator of bottom-water velocity fluctuations (Ledbetter and Ellwood, 1980; Blaeser and Ledbetter, 1982). Particle-size analyses were therefore conducted on samples from cores K78-5-10 and M70-17 to determine if a link can be detected between hiatus occurrence and bottom-current activities in the Central Basin. Because relative changes in the silt fraction are most significant for paleocurrent studies (Ledbetter and Ellwood, 1980; Blaeser and Ledbetter, 1982), the particle-size analysis was restricted to this fraction. A variety of statistical parameters can be computed from a given particle-size distribution. One of
the parameters commonly used in current-related paleoceanographic studies is the mean grain size (e.g., Ledbetter and Ellwood, 1980; Blaeser and Ledbetter, 1982; Ledbetter and Balsam, 1985; Johnson et al., 1985). Despite wide application, this parameter has been criticized, even described as 'unrealistic' (Anderson and Kurtz, 1985), on the basis of laboratory flume experiments and theoretical considerations. Actual field observations in known current-controlled environments, however, do support a relationship between silt-size parameters and the bottom-water dynamics (Ledbetter and Ellwood, 1980; Bulfinch et al., 1982).

Many authors use the silica- and carbonate-free subfractions for particle-size analysis and sometimes also remove organic carbon and iron-manganese-oxy-hydroxides. This is done to eliminate the distorting effects of (1) postdepositional dissolution of silica and carbonate grains; (2) the growth of authigenic minerals; (3) the evolutionary size changes in organisms; or (4) productivity changes. All these factors can potentially alter the original particle-size distribution. Furthermore, most biogenic sediment components fall vertically through the water column; they are unrelated to components laterally-transported by and indicative of bottom currents. Thus, a silt-fraction particle-size analysis, based solely
on vertically-derived biogenic components, would lead to erroneous results if used to establish the presence or absence of laterally winnowing bottom currents (or changes in their velocity).

Because of their depths of deposition, the sediments of cores M70-17 and K78-5-10 are generally free of carbonate particles. Manganese micronodules are a common component in some sections of the cores and the possibility that they were transported laterally cannot be excluded. Therefore, digestion of the silt fraction was limited to the removal of biogenic silica.

3.2.2. Sample Preparation and Analytical Methods

The grain-size analysis of the silt fraction was conducted with an Elzone Model 80XY Particle Analyzer, manufactured by Particle Data, Inc. at Duke University Marine Lab, Beaufort, North Carolina. This instrument operates by the same principle as a Coulter Counter: a subsample of the silt fraction is suspended in an electrolyte solution and sizes of particles are analyzed as they pass through a small orifice in a capillary tube between two electrodes. Every pass causes a change in resistivity which is proportional to the size of the particle. The Elzone's silt-size range is divided into 128 discrete increments, allowing a resolution of size
difference to about 0.03 phi. The instrument measures the relative particle abundance in each of these channels. This results in a population distribution, which is converted by the Elzone's computer to a volume distribution, by assuming spherical particles. This volume distribution is comparable to a mass distribution, if the measured particles have approximately the same density. The Elzone's microprocessor smooths the volume distribution before computing the mean size, mode, and median.

The instrument was calibrated at the beginning of the measurements. A more detailed description of the operation and calibration of the Elzone electronic particle analyzer is given by Halfman (1982). The samples were analyzed following the procedures outlined in Fig. 12. All reagents were filtered through 45 micrometer filters to remove larger particles in the fluids which would alter the original silt-size distribution.

For comparison, most of the samples were measured both with and without biogenic silica and the clay fraction. All analyses were duplicated, and, at times, triplicated, if the mean grain sizes of the first two measurements were not within three channels (approximately a 0.9 phi increment) of the range of the instrument. The arithmetic average of the individual analyses was used to calculate the mean grain sizes. The series containing the clay fraction required the
FIGURE 12. Summary diagram, showing stepwise sample preparation for silt-size particle analysis.
0.3 g wet sediment, weighed in a glass beaker (taken from cubes used for AMS measurements)

REMOVAL OF SAND FRACTION:
- add 15% H₂O₂ to disperse sediment
- wet-sieve sample through 63 micrometer mesh into a 250 ml NALGENE bottle
- rinse with filtered deionized H₂O

REMOVAL OF BIOGENIC SILICA:
- add 200 ml filtered 5% Na₂CO₃
- let samples sit in hot water bath at 85⁰C for 2 (8) hrs
- stir sediment every 30 min
- decant supernatant
- rinse three times with filtered deionized H₂O

REMOVAL OF CLAY FRACTION:
- fill NALGENE bottles with a filtered CALGON solution (2.5 g/l)
- centrifuge at 500 rpm for 3.5 min
- decant supernatant and repeat until solution stays clear
- decant supernatant and add filtered 1% NaCl solution

ELZONE Particle Analyzer

ELZONE Particle Analyzer
repetition of the measurements more frequently than the clay-free one, owing to flocculation problems. In general, the results were reproducible to within the chosen channel width.

The removal of biogenic silica (Fig. 12) followed the procedures outlined by Follet et al. (1965 a, b) as modified by DeMaster (1979). To determine what concentration of Na$_2$CO$_3$ dissolved most of the biogenic silica, without significantly attacking the detrital portion of the sediment, several samples were treated with a variety of different Na$_2$CO$_3$ concentrations at a constant temperature of 85° C. At time intervals of 30 minutes, microscopic slides of the test samples were prepared and the dissolution effects studied under a microscope. A 5% Na$_2$CO$_3$ solution was determined to be the most effective. A digestion time of about 2 hrs was required for the M70-17 samples. Because of the relatively higher silica and lower clay content, 8 hrs of digestion time were necessary for dissolution of biogenic silica in the K78-5-10 samples. Radiolarian tests intact after the leaching process were usually covered by clay particles, and therefore protected from dissolution.

To be certain that most of the biogenic silica was actually dissolved after the Na$_2$CO$_3$ treatment, a silica extraction analysis was also conducted for some of the samples. This analysis is based on the fact that the
dissolution curve for the sediment's total silica content consists (DeMaster, 1979) of two separate intervals: an initial steep increase of dissolved silica, followed by a much slower linear increase (Fig. 13a). The first interval corresponds to the leaching of the more easily dissolved biogenic silica, whereas the following flat slope corresponds to the dissolution of clay minerals, which dissolve at a slower rate. To determine the percent silica extracted versus time, 1 ml of solution was extracted from the top of each sample bottle at 30, 60, 90, and 120 minute intervals during the leaching process. These 1 ml extractions were then diluted in 99 ml of silica-free deionized water and analyzed on a Technicon Auto Analyzer. The bend in the resulting dissolution curve at 60 minutes indicates, that at that point most of the biogenic silica was dissolved. The dissolution curve of sample 612.5 cm of core M70-17 is plotted as a representative example (Fig. 13b).

3.2.3. Results

Results of the silt-size analyses across the defined unconformities are not uniform for the two cores under investigation (Figs. 14, 15). Undigested, as well as opal-free, samples analyzed across hiatus NH6 of core K78-5-10 do not show a distinct coarsening of silt-size
FIGURE 13. (A) Silica dissolution model of DeMaster (1979); a steep slope of the dissolution curve (representing dissolution of biogenic silica) is followed by a lower, linear slope (representing dissolution of silica from clay minerals); the dissolution of silica from clay minerals is considered to be constant with time and therefore extrapolation of the linear portion of the curve indicates the percentage of biogenic silica dissolved; (B) Silica dissolution curve of sample 612.5 cm of core M70-17 as a representative example; intersection of the steep and shallow slopes indicates complete removal of biogenic silica after approximately 60 min.
A

SILICA EXTRACTED
(weight %)

DURATION OF EXTRACTION

% Silica from clay minerals

% Biogenic silica

B

M 70-17 (612.5 cm)

SILICA EXTRACTED (weight %)

TIME (min.)

- 68 -
FIGURE 14. Mean silt sizes of undigested (A) and opal-free (B) samples taken across the K78-5-10 core interval containing the NH6 event. Shading indicates the hiatus-related core interval.
FIGURE 15. Mean silt sizes of undigested (A) and opal-free (B) samples taken across the NH6-related barren zone and of undigested (C) and opal-free (D) samples taken across the core interval containing NH5, in core M70-17. Shadings indicate hiatus-related core intervals.
particles (Fig. 14). A single shift towards a coarser mean silt size in core K78-5-10 at a depth of 1265.5 cm, which occurs only in the undigested sample set, cannot be considered as sufficient evidence for silt-size changes related to hiatus NH6. A minor increase in the mean silt size in the opal-free samples between 1250.5 and 1258 cm also cannot be considered as sufficient evidence for particle coarsening, although this is the core interval in which textural changes were found by means of X-radiography (Fig. 6). The average mean silt size is 17 micrometers for the undigested sample set, and 19 micrometers for the opal-free samples.

Silt-size analysis data across the barren zone of core M70-17, however, reveal distinct increases in the mean silt size in both the undigested and opal-free samples (Fig. 15). This shift towards a coarser mean silt size occurs at different depth levels in the two sample sets: samples containing biogenic silica record a mean silt size increase between 494.5 and 504 cm. Thus, the coarsening of silt-size particles in the undigested samples coincides with textural changes detected by X-radiography. The coarsening of the mean silt size in the opal-free samples begins at the same depth level, 494.5 cm, but increases gradually downcore to a maximum between 526 and 530 cm. Below this peak the mean silt size decreases gradually down to 558.5 cm. From here
down to the last analyzed sample at 587 cm the mean silt size remains constant. The occurrence of coarsest silt-size particles in the undigested samples between 526 and 530 cm coincides with the maximum relative abundance of manganese micronodules (Fig. 11), suggesting a genetic relationship. The analyses of the undigested samples (Fig. 15) did not cover the same core interval as the opal-free samples and thus could not detect the coarsening of silt-size particles between 526 and 530 cm.

A similar coarsening of silt-size particles in the opal-free samples can be observed across hiatus interval NH5 of core M70-17 (Fig. 15). This increase in mean silt size occurs in a core interval between approximately 787.5 and 804 cm. This is the same core interval in which textural changes were detected by X-radiography (Fig. 7).

The older hiatus interval of NH3 in core M70-17 is not suitable for silt-size analyses, because of the occurrence of a very coarse volcanic ash layer at this depth.

3.3. Discussion of Results

Interpretation of the particle-size analysis results is very difficult and can, at this point, only be speculative. Grain-size data associated with NH6 in core K78-5-10 cannot be considered sufficient evidence for a silt-size increase
across this hiatus. Interpretation of data across the related barren zone of core M70-17 is hampered by the fact that it is unknown if manganese micronodules (and clinoptilolite) are of allochthonous or autochthonous origin. Thus, the silt-size increase around 530 cm (Fig. 15) in the opal-free samples might reflect either a coarsening because of a preferred growth of authigenic minerals or an intensification of reworking of allochthonous particles. With all the present knowledge of the zeolite clinoptilolite (as discussed earlier) it is intriguing to argue for reworking and the winnowing effect of bottom currents around 530 cm as a cause for silt-size coarsening.

Silt-size data across NH5 of core M70-17 (Fig. 15) suggest a coarsening of particles in the opal-free sample set but a shift towards finer mean grain size in the undigested samples. X-ray diffraction results (Fig. 10) indicated a decrease in the relative abundance of clinoptilolite across this hiatus interval. Thus, clinoptilolite in this case is an unlikely candidate responsible for the silt-size coarsening in the opal-free samples. Because of logistical problems compositional analyses across the NH5 interval do not exist to date. Therefore it is not known at this time if the increase in mean silt size is accompanied by a parallel increase in the relative abundance of manganese micronodules.
IV. CHANGES IN MAGNETIC PROPERTIES ACROSS THE UNCONFORMITIES AND RELATED CORE INTERVALS: ANISOTROPY OF MAGNETIC SUSCEPTIBILITY, MAGNETIC MINERALOGY, AND MAGNETIC GRANULOMETRY

1. Introduction

1.1. Previous Applications of Anisotropy of Magnetic Susceptibility (AMS) in the Study of Marine Sediments

Ever since the pioneering work of Ising (1942), Graham (1954, 1967) and Rees (1961, 1965), anisotropy of magnetic susceptibility (AMS) has been widely used to study sedimentary magnetic fabrics. In paleoceanographic research, AMS has been successfully employed to ascertain the influence of bottom currents and to determine relative flow magnitudes and directions (Ellwood and Ledbetter, 1977; Ellwood et al., 1979; Ellwood, 1980a, b; Ledbetter and Ellwood, 1980; Auffret et al., 1981; Bulfinch et al., 1982; Rees et al., 1982; Shor et al., 1984; Flood et al., 1985). Most of these studies, however, were conducted in well-studied areas of known high-velocity bottom currents, such as the Vema Channel in the South Atlantic, the
Southeast Indian Ridge, and the Nova Scotia continental rise (HEBBLE area). Sediments in the regions of these investigations are relatively young, usually of Quaternary age. In contrast to conditions in the present study area (the Central Basin of the Pacific), these earlier studies focused on sites where high sedimentation rates and coarse particle sizes prevail. Thus, to the author's knowledge, AMS studies have not previously been conducted in comparatively low-energy environments like the deep equatorial Pacific.

1.2. Theory of AMS

The ratio of induced magnetization to the strength of the magnetic field causing the magnetization is a symmetrical, second-rank tensor coefficient and is defined as susceptibility. The latter quantity can be expressed either as volume (K) or mass (X) susceptibility, and it is not necessarily distributed homogeneously or isotropically in a given sample. Factors which can cause anisotropy are described by Taira and Lienert (1979). The most important are: (1) shape of individual magnetic grains; (2) shape of the bulk sample; (3) distribution of magnetic grains within the sample; (4) magnetocrystalline anisotropy of individual magnetic grains. A combination of these effects is possible. The susceptibility anisotropy can be represented
by a triaxial ellipsoid, with dimensions defined by the magnitudes of the principal susceptibilities $K_1$, $K_2$, and $K_3$ (maximum, intermediate, and minimum directions, respectively). The mean or bulk susceptibility is then:

$$K = \frac{(K_1 + K_2 + K_3)}{3}.$$  

According to Taira and Lienert (1979) the shape of individual magnetic grains has the greatest effect on susceptibility anisotropy, provided the sediment contains low concentrations of magnetite. If hematite is the predominant magnetic mineral, however, the magnetocrystalline anisotropy of individual magnetic grains becomes important.

The use of AMS in paleoceanographic studies is based on the assumption that elongated magnetic grains become aligned uniformly during deposition by bottom current flow. Both parallel and normal alignment to the flow direction has been observed (e.g., Ellwood, 1980a; Flood et al., 1985). This variability in $K_1$ orientation may be due to different possible alignments of grains deposited from a flow, or to variations in bottom current strength and direction (Flood et al., 1985). In any case, the alignment of magnetic grains is thought to be representative of the alignment of other grains in the sample (Hamilton and Rees, 1970). Preferred
grain orientations in marine sediments are believed to reflect the aligning activity of bottom currents (Rees et al., 1968; Ellwood, 1980a). Other factors producing or influencing magnetic fabric are discussed by Rees et al. (1968).

Rees (1965) outlines the limitations of AMS: (1) sediments should contain a certain quantity of magnetite (0.01 to 1 %), enough to produce anisotropy, but not so much as to cause interactions between magnetic grains; and (2) the magnetic particle size should be coarser than approximately 10 micrometers, because the orientation of finer particles is in part controlled by the earth's magnetic field and not by depositional processes.

1.3. AMS Parameters used in this Study

Many different parameters and statistical factors are used in the literature to describe the magnetic fabric of rocks. Some of these parameters, not all of which have been universally accepted, are listed by Hrouda (1982). The nomenclature of the anisotropy parameters is often confusing and their present overabundance calls for a standardization of AMS data and their representation (Jelinek, 1981). With this aim in view, recommendations recently made by Ellwood et al. (submitted) were followed in the course of this study.
AMS parameters are usually based on the ratios of principal susceptibilities or on susceptibility differences. Sometimes a combination of both is used. Two important parameters are magnetic lineation and foliation. They are usually defined as ratio parameters:

\[
L = \frac{K_1}{K_2} \quad \text{(Balsley and Buddington, 1960)}
\]
\[
F = \frac{K_2}{K_3} \quad \text{(Stacey et al., 1960)},
\]

or difference parameters:

\[
l = \frac{(K_1 - K_2)}{K} \quad \text{(Khan, 1962)}
\]
\[
f = \frac{(K_2 - K_3)}{K} \quad \text{(Khan, 1962)}.
\]

The term magnetic lineation (either as L or l) describes the linear-parallel orientation of ferromagnetic components in a sample, while the term magnetic foliation (F or f) characterizes the relative intensities of planar-parallel oriented particles. The magnetic foliation plane is perpendicular to the axis of minimum susceptibility, \(K_3\). The magnetic lineation lies in this foliation plane and its direction is identical to that of the maximum susceptibility \(K_1\). Low bulk susceptibilities and few linear oriented components affect the reliability of the lineation direction. Flood et al. (1985) point out that the
lineation direction (and therefore the orientation of $K_1$) is one of the most poorly determined components of the magnetic fabric, especially in foliated sediments.

A parameter which describes the total anisotropy of a sample is $H$ (Owens, 1974), which is also called the degree of anisotropy:

$$H = \frac{(K_1 - K_3)}{K} \times 100.$$ 

All magnetic susceptibilities are quoted in dimensionless Systeme Internationale (SI) units.

2. Methods

Because AMS measurement required the use of truly cubic samples $8 \text{ cm}^3$ in volume, the cores were resampled across the previously defined hiatus intervals. The low sedimentation rates in the Central Basin of the Pacific made it mandatory to sample continuously, with overlapping samples taken at important levels. The maximum and minimum mid-point sample spacings were $5 \text{ cm}$ and $1 \text{ cm}$, respectively.

The AMS measurements were all made in the paleomagnetic laboratory of the University of Texas at Arlington, using two automated, low-field, torsion fiber magnetometers ("torque meter"; e.g., King and Rees, 1962; Stone, 1963).
They were calibrated using a machined, seamless copper ring (Noltimier, 1964). Initial bulk susceptibility of the samples was determined using a susceptibility bridge. The AMS ellipsoid was then calculated from torque meter and bridge data.

3. Results of Rock-Magnetic Analyses

3.1. Bulk Magnetic Susceptibility (K)

In both cores, intervals related to hiatus events, with the exception of the one linked to NH5, can be correlated with increases in the bulk magnetic susceptibility (K) of the sediment (Figs. 16, 17, 18). The susceptibility increases across the core intervals related to the NH6 event are not abrupt: both cores reveal a gradual upcore increase in K (Figs. 16, 17). In core K78-5-10, the susceptibility increase across the NH6 interval is correlated with textural changes seen in the X-radiographs (Fig. 6). Although no textural changes were detected in X-radiographs of the NH6-related barren zone in core M70-17, the susceptibility increase in this core culminates at a depth interval (523-530 cm) which also revealed the coarsest silt-size particles of the opal-free samples (Fig. 15) and a maximum in the abundance of manganese micronodules (Fig. 11).
FIGURE 16. Volume susceptibility (K), total anisotropy (H), lineation (l) and foliation (f) differences of samples taken across the K78-5-10 core interval containing the NH6 event. Shading indicates the hiatus-related core interval.
K78-5-10

(E-05/volume SI units)
FIGURE 17. Volume susceptibility (K), total anisotropy (H), lineation (l) and foliation (f) differences of samples taken across the NH6-related barren zone of core M70-17. Shading indicates the hiatus-related core interval.
M70-17

(E-04 /volume SI units)
FIGURE 18. Volume susceptibility (K), total anisotropy (H), lineation (l) and foliation (f) differences of samples taken across the M70-17 core interval containing the NH5 and NH3 events. Shadings indicate the hiatus-related core intervals.
M70-17

(E-04 /volume SI units)
A similar increase in $K$ occurs across the older hiatus interval of NH3 (between approximately 868.5 and 875 cm) of core M70-17 (Fig. 18). Compared to the susceptibility increases over the NH6 intervals in the two cores, the increase corresponding to NH3 is less pronounced, but also gradual. As in core K78-5-10, the increase in magnetic susceptibility correlates with textural changes detected by X-radiographs (Fig. 7).

3.2. AMS Parameters ($H$, $I$, and $f$)

In core K78-5-10, hiatus NH6 is characterized by distinct increases in total anisotropy ($H$), magnetic lineation ($I$), and foliation ($f$) at a depth of 1259.5 cm, as well as the gradual increase in bulk susceptibility (Fig. 16). The increase in $H$ is extraordinarily high and therefore may not reflect an improved alignment of elongated magnetic grains. Equal area, lower hemisphere plots of the principal susceptibility axes of samples between 1253 and 1263 cm (Fig. 19) reveal surprising details. The principal susceptibility directions are not randomly oriented but follow a distinct pattern: (1) all minimum directions are oriented parallel to the sedimentary bedding plane and lie on the same great circle; (2) directions of intermediate and maximum susceptibilities also all lie on a great circle; (3) the maximum susceptibility direction of the sample
FIGURE 19. Lower hemisphere plot of principal susceptibility directions across the K78-5-10 core interval containing the NH6 event. Maximum ($K_1$), intermediate ($K_2$), and minimum ($K_3$) susceptibility directions are indicated by squares, triangles, and circles, respectively. Across this core interval all axes of maximum and intermediate susceptibility are aligned along a great circle and generally oriented perpendicular or near-perpendicular to the bedding plane. Axes of minimum susceptibility are oriented parallel to the bedding plane.
K78-5-10

1253–1263 cm
lower hemisphere plot

□ $K_{\text{max}}$
△ $\Delta K_{\text{int}}$
○ $K_{\text{min}}$
containing greatest changes in AMS parameters (1259.5 cm) is oriented perpendicular to the bedding plane. As discussed earlier, theoretical considerations call for a horizontal alignment of maximum susceptibility directions under the influence of bottom currents. This peculiar magnetic fabric across hiatus interval NH6 is not the result of a sudden reorientation of the principal susceptibility axes (Fig. 20). Beginning downcore, samples between 1292 and 1325 cm exhibit a random orientation of axes. The next sample set, 1265.5-1289.5 cm, shows increased uniformity in the orientation of susceptibility directions, with the intermediate and maximum directions being along a great circle. This increase in alignment then culminates across hiatus interval NH6, between 1253 and 1263 cm. Across the next sample interval upcore, 1234.5-1250.5 cm, the uniformity of orientation of the susceptibility axes is less pronounced, but then increases again between 1200-1232 cm (Fig. 20).

In core M70-17, correlations between AMS parameters, bulk susceptibility, and textural changes are not as straightforward as in core K78-5-10. In core M70-17, textural changes are present between 495 and 515 cm (Fig. 10), and the peak in bulk susceptibility falls between 523 and 530 cm. There are no significant changes in AMS parameters in these intervals (Fig. 17). An increase in
FIGURE 20. Lower hemisphere plots of principal susceptibility axes of various successive sample sets in relation to depth and the polarity log of core K78-5-10. Note the improved, although unusual, alignment of susceptibility axes across the NH6-containing core interval between 1253 and 1263 cm. Although the magnetic fabric cannot be considered to be of primary origin, the results of the individual sample sets suggest gradual changes in axis alignment across the hiatus interval. For legend see Fig. 19.
total anisotropy and magnetic lineation, however, occurs at 490 cm, above the barren zone. Plots of the principal susceptibility directions (Fig. 21) do not show a preferred orientation pattern. Perpendicular (or near perpendicular) orientation of the axes of maximum susceptibility, however, is most pronounced across sample intervals 505.5-524.5 and 526-542.5 cm (Fig. 21).

In contrast to the barren zone related to NH6, hiatus NH5 of core M70-17 is linked to a shift towards a higher total anisotropy and an increase in lineation at a depth of 796 cm (Fig. 18). This shift also correlates with textural changes detected by X-radiography (Figs. 7, 10) and a coarsening of the mean silt size (Fig. 15). The magnetic foliation remains unchanged across this core interval. Lower hemisphere plots of the main susceptibility axes (Fig. 22) across the NH5 interval do not show a preferred alignment.

Hiatus NH3 in core M70-17 cannot be linked to changes in AMS parameters, although it correlates with an increase in bulk susceptibility (Fig. 18). Thus, directions of principal susceptibility axes were not plotted.
FIGURE 21. Lower hemisphere plots of principal susceptibility axes of various successive sample sets across the barren zone of core M70-17. Results of individual sample sets are plotted against depth along the polarity log. Hachuring in the polarity log indicates an interval of disturbed paleomagnetic data. A comparison of the results of the various sample sets does not indicate a preferred alignment of principal susceptibility axes at a certain depth. Note, however, that most of the axes of minimum susceptibility are oriented parallel to the bedding plane. The sample set 475-504 cm, which covers the core interval that contains textural changes detected by X-radiographs, displays a distinct orientation of both axes of maximum and minimum susceptibility parallel to the bedding plane. For legend see Fig. 19.
M70-17

475-504

505 - 524

526-545

546-587

592-617

480

500

520

540

560

580

600

620

(cm)
FIGURE 22. Lower hemisphere plots of principal susceptibility axes of various successive sample sets across the M70-17 core interval containing NH5.
3.3. Magnetic Mineralogy

The peculiar magnetic fabric (Fig. 19) across the core interval containing NH6 in core K78-5-10 suggests the possibility of postdepositional growth of an authigenic magnetic mineral perpendicular to the bedding plane (Harrison and Peterson, 1965; Ellwood et al., 1986). In order to identify the dominant magnetic mineral, isothermal remanent magnetization (IRM) acquisition and demagnetization curves (Fig. 23) were plotted for selected samples of core K78-5-10. To reach saturation IRM (SIRM), samples were exposed to a stepwise increasing DC magnetic field from 25 to 12,500 Oe at room temperature (for details see Theyer and Dorn, in prep.). The saturated samples were demagnetized in an alternating magnetic field from 0 to 1,000 Oe. All samples of core K78-5-10 attained saturation by approximately 5,000 Oe, indicating that magnetite is the dominant magnetic mineral (King et al., 1983). This interpretation is further supported by the intersection of IRM acquisition and demagnetization curves at approximately 50% of the maximum intensity of each sample. The uniform behavior of all analyzed samples and the fact that magnetite is considered to be detrital or of biogenic origin make it very unlikely that the perpendicular or near-perpendicular
FIGURE 23. Isothermal remanent magnetization (IRM) acquisition (black dots) and demagnetization (open circles) curves of selected samples of core K78-5-10 (1205, 1210, 1259.5, 1260.5, 1306, and 1308.5 cm). All samples attain saturation by approximately 5,000 Oe, indicating that magnetite is the dominant magnetic mineral. This interpretation is further supported by the intersection of IRM acquisition and demagnetization curves at approximately 50% of the maximum intensity of each sample.
orientations of the $K_1$ axes are caused by authigenic growth of a magnetic phase.

3.4. Magnetic Grain-Size Analyses

Theoretical considerations (Banerjee et al., 1981), verified by experimental studies (King et al., 1982), have established a relationship between the ratio of anhysteretic susceptibility ($K_{ARM}$) to magnetic susceptibility ($K$) and relative magnetic grain size (Fig. 24). $K_{ARM}$ and $K$ values are also an indicator of relative changes in the concentration of magnetic carriers (King et al., 1982). In order to acquire a remanent magnetization, the samples were subjected to a decreasing alternating magnetic field of 1,000 Oe, with a superimposed small steady field of 0.5 Oe (Theyer and Dorn, in prep.). The samples acquire a remanent magnetization that is approximately proportional to the superimposed DC field. $K_{ARM}$ values were plotted against $K$ for samples of both cores (Figs. 25, 26, 27). Higher values of $K_{ARM}/K$ (slopes of greater magnitude) correspond to relatively finer magnetic grain sizes. Greater distance to the origin of the plot (intersection of $K_{ARM}$ and $K$ axes) indicates higher concentrations of magnetic grains.

In both cores, samples associated with the hiatuses or related core intervals are marked by relatively coarser magnetic grain size and/or relatively higher concentrations.
FIGURE 24. Model of King et al. relating variations between the ratio of anhysteretic susceptibility and magnetic susceptibility to changes in relative concentrations and magnetic grain sizes of magnetite (after King et al., 1982).
Equidimensional
$\text{Fe}_3\text{O}_4 \approx 1\%$ by volume

\[ \chi_{\text{ARM}} \times 10^{-3} \text{ SI} \]

\[ \chi_{\text{volume}} \times 10^{-3} \text{ SI} \]

1.0 gram
0.1 $\mu$m

0.5 gram
0.2 $\mu$m

$\approx 5.0 \mu$m

20-25 $\mu$m

200 $\mu$m
FIGURE 25. Anhysteretic susceptibility (ARM) versus magnetic susceptibility (K) plots, indicating relative concentration and grain-size differences of magnetic particles in samples of core K78-5-10 (1200-1325 cm). NH6-related sample at 1260.5 cm is marked by relatively coarse magnetic grain size and high concentration of magnetic particles.
K78-5-10 (1200-1325 cm)
FIGURE 26. Anhysteretic susceptibility (ARM) versus magnetic susceptibility (K) plots, indicating relative concentration and grain-size differences of magnetic particles in samples of core M70-17 (475-617 cm). Samples associated with the NH6-related barren zone are marked by relatively coarse magnetic grain sizes and high concentrations of magnetic particles.
M70-17 (475-617 cm)
FIGURE 27. Anhysteretic susceptibility (ARM) versus magnetic susceptibility (K) plots, indicating relative concentration and grain-size differences of magnetic particles in samples of core M70-17 (763-899 cm). Samples associated with hiatus NH5 are marked by relatively coarse magnetic grain sizes and high concentrations of magnetic particles. No changes can be observed in samples related to hiatus NH3(NH4?).
M70-17 (763-899 cm)
of magnetic particles compared to the majority of samples in a set (Figs. 25, 26, 27). The remaining samples of each set have comparatively uniform magnetic grain size (all data points aligned along the same slope) and vary only in the concentration of magnetic particles (shifts of data points along the slope).

In core K78-5-10 (Fig. 25) coarsest magnetic grain sizes were found at core depths of 1306 and 1308.5 cm. Graphic correlation plots (Fig. 5) do not indicate a hiatus in this core interval. A comparison with the age-depth plot of core M70-17 (Figs. 5, 31), however, suggests a relationship between the NH5 event and this magnetic grain size coarsening. At a depth of 1260.5 cm in core K78-5-10 a sample associated with hiatus NH6 is marked by coarse magnetic grain size and high concentration of magnetic particles (compared to the bulk of the sample set between 1200 and 1325 cm).

Within the sample set (475-617 cm) representing the NH6 event in core M70-17 (Fig. 26), samples from the top and middle of the barren zone are characterized by a distinct coarsening of magnetic carriers. Another interval of coarsening of magnetic particles is present between 602 and 617 cm. The majority of the NH6 sample set is of uniform grain size with individual samples differing only in the concentration of magnetic particles. Samples associated
with the NH5 event (Fig. 27) are also marked by a coarsening of magnetic grains. No changes in magnetic grain size can be observed across the NH3 interval. There seems to be a correlation for M70-17 samples between magnetic grain size and the results of the silt-size analyses. Samples containing coarse magnetic particles across the NH6 interval also have coarse mean silt sizes as measured on the undigested samples. A similar correlation exists across NH5 between samples of coarse magnetic grain size and those with coarse mean silt sizes for the opal-free samples (Fig. 15).

In addition to the $K_{\text{ARM}}$ (= ARM) versus $K$ graphs (Figs. 25, 26, 27), ARM/$K$ ratio curves were plotted together with ARM, IRM, and natural remanent magnetism (NRM) intensities (Figs. 28, 29, 30). The significance of the various intensity curves is discussed in detail by Theyer and Dorn (in prep.). ARM and IRM curves both reflect variations in the concentration of magnetic particles. The ARM curves, however, are more susceptible to concentration variations of relatively finer magnetic grains, whereas the IRM curves are more sensitive to relatively coarse magnetic grains. ARM/$K$ curves are a particularly sensitive granulometric parameter (King et al., 1982). High values reflect the dominance of relatively finer magnetic grains, lower values those of relatively coarser ones. ARM/$K$, ARM, and IRM curves (Figs. 28, 29, 30) support previous findings: (1) hiatus NH6 of
FIGURE 28. NRM, ARM/K, ARM, and IRM curves based on samples of core K78-5-10 (1200-1325 cm).
K78-5-10 (1200-1325 cm)

Depth (m)
FIGURE 29. NRM, ARM/K, ARM, and IRM curves based on samples of core M70-17 (475-617 cm).
M70-17 (475-617 cm)

Depth (m)

- 117 -
FIGURE 30. NRM, ARM/K, ARM, and IRM curves based on samples of core M70-17 (763-899 cm).
core K78-5-10 is marked by an increase in the concentration and size of magnetic carriers (Fig. 28); (2) the NH6-related barren zone of core M70-17 is marked by an increase in the concentration of magnetic grains; samples near the top and in the middle of this zone also show a relative magnetic grain-size coarsening (Fig. 29); (3) hiatus NH5 is marked by both an increase in concentration and the coarsening of magnetic grains; (4) hiatus NH3 of core M70-17 seems to correlate with an increase in the concentration of magnetic particles. In contrast to the ARM versus K plots (Figs. 25, 26, 27), however, the ARM, ARM/K, and IRM curves (Figs. 28, 29, 30) reveal more detailed rock-magnetic changes associated with the various hiatus events. Hiatus NH6 of core K78-5-10, for example, can be correlated with distinct oscillations of size and concentration of magnetic particles, suggesting a succession of several "events". Hiatus NH6 of core K78-5-10, the related barren zone of core M70-17, and hiatus NH3 in M70-17 are also marked by distinct increases in the NRM intensity (Figs. 28, 29, 30). Hiatus NH5 of core M70-17, however, is associated with a relative decrease in NRM intensity (Fig. 30).
4. Discussion

The striking similarity between the susceptibility increase across hiatus NH6 of core M70-17 (Fig. 17) and an increase in the relative abundance of manganese micronodules (Fig. 11) across the same core interval suggests an intimate causal relationship: increase in the abundance of manganese micronodules apparently results in a susceptibility increase. It is known that manganese nodules can carry a significant remanent magnetization (Carpenter et al., 1972; Crecelius et al., 1973; Henshaw and Merril, 1980), although little is known about the nature of the individual ferromanganese phases. This is partially due to the fact that many of the ferromanganese phases in the marine environment are microcrystalline or amorphous (Henshaw and Merril, 1980) and thus escape detection when using standard X-ray techniques (Crecelius et al., 1973; Johnston and Glasby, 1978). The interval in which the increases in bulk susceptibility and manganese-micronodule abundance occur in core M70-17 is also marked by a loss of the coherent magnetic stratigraphy (Fig. 4). Increases in the concentration of manganese micronodules which result in magnetic instability have been described by other authors (Opdyke and Foster, 1970; Haggerty, 1970). Previous analyses of similar cores (Kent and Lowrie, 1974; Johnson et al.,
1975) linked the deterioration of the magnetic signature to an increase in maghemization, owing to low-temperature oxidation of detrital magnetite. Kent and Lowrie (1974) do not dismiss the possibility that some maghemite may have formed authigenically and is perhaps related to the presence of ferromanganese oxides, instead of solely being a result of oxidation of detrital magnetite. This interpretation is consistent with results from Henshaw and Merril (1980), who observed that the authigenesis of ferromanganese phases and maghemization occur together in non-fossiliferous cores from the Pacific containing manganese micronodules. Results of leaching experiments provide strong evidence (Henshaw and Merril, 1980) that authigenically produced ferromanganese particles, most likely todorokite, overprint the sediment's original remanence and cause magnetic instability. Thus, associated maghemization seems to play only a minor role in altering the magnetic signature.

In view of these previous studies and the aforementioned similarity between increases in bulk susceptibility and manganese micronodule concentrations across NH6 in core M70-17 (Figs. 11, 17), it seems apparent that yet undefined ferromanganese phases, associated with the nodules, are indeed responsible for both susceptibility increase and magnetic instability. If this interpretation is correct, it might also explain susceptibility increases
across hiatus NH6 of core K78-5-10 (Fig. 16), although analyses of the manganese micronodule concentrations across this interval were inconclusive owing to an apparent lack of nodules in the >212 micrometer size fraction.

Magnetic grain size analyses (Figs. 25, 26, 27) show that samples taken closest to the hiatus intervals in both cores contain relatively coarse magnetic particles. This coarsening can be interpreted as a result of winnowing by bottom currents which resulted in the concentration of coarser particles and/or particles of relatively higher specific gravity (Bloemendal, 1983). It has been argued that magnetotactic bacteria play a major role in the magnetization of sediments in detritus-poor environments (Kirschvink and Lowenstam, 1979; Petersen et al., 1986). If this assumption is correct, then one also has to consider the possibility that periods of nondeposition might coincide with increases in growth of magnetotactic bacteria, and thus in the size of magnetosomes. Our present knowledge of magnetotactic bacteria is too limited to discuss the likelihood of such a scenario.

The orientation of the principal susceptibility directions across NH6 of core K78-5-10 (Figs. 19, 20) is difficult to explain. The distinct arrangement of maximum and intermediate directions along the same great circle, as well as the uniform horizontal orientation of the minimum
directions, imply the influence of some aligning force, although the magnetic fabric cannot be considered to be of primary origin. Bottom-current activity would easily explain such an alignment if the axes of maximum susceptibility were parallel to the bedding plane and if the magnetic fabric was original. As the lower hemisphere plots show (Figs. 19, 20), some of the $K_1$ directions are oriented perpendicular or near-perpendicular to the bedding plane. Vertical orientations of $K_1$ have also been reported by other authors (Harrison and Peterson, 1965; Kent and Lowrie, 1975; Flood et al., 1985) and were described to be common in deep-sea sediments (Lovlie et al., 1971). Harrison and Peterson (1965) attribute this peculiar alignment to some authigenic magnetic mineral. This interpretation has been questioned by Lovlie et al. (1971) who ascribe it to a secondary magnetic fabric caused by the coring process. Kent and Lowrie (1975) follow this explanation and discuss the possibility of undetected "flow-in" occurring within a core. Recent studies (Rees et al., 1982; Flood et al., 1985) interpret vertical orientations of $K_1$ as secondary magnetic fabrics caused by bioturbation. Other investigations (Ellwood et al., 1986) attribute a vertical orientation of axes of maximum susceptibility to the presence of the mineral siderite, either due to a postdepositional overprint or to authigenesis during core storage (Ellwood et al., 1986).
None of the aforementioned speculations can sufficiently explain the observed vertical orientation of $K_1$ across hiatus interval NH6 of core K78-5-10 in a siliceous brown-clay environment. IRM acquisition and demagnetization curves (Fig. 23) do not indicate the presence of an unusual magnetic mineral. The change in principal axes orientations across NH6, although most likely of secondary origin, appears to be too gradual and systematic (Fig. 20) to be ascribed to coring disturbances or other sampling artifacts. Furthermore, it is not apparent why mechanical deformation should result in such distinct uniform alignment of the principal axes of susceptibility as observed across NH6 (Figs. 19, 20). For the same reasons, bioturbation is an unlikely explanation for a vertically prolate fabric (Ellwood, 1984), although it seems to be of secondary origin. Lovlie et al. (1971) demonstrated that vertical orientation of $K_1$ does occur to the same extent in both mottled and burrowed and seemingly unbioturbated sediments. Their analysis of 120 specimens (from various sites) shows also that a vertical orientation of $K_1$ is much more common than the expected horizontal one. Although being far from uncommon, the vertical or near-vertical orientation of $K_1$ in relation to the bedding plane cannot be explained.
V. LATE MIocene HIATUSES AND THEIR RELATIONSHIP TO
PALEOCEANOGRAPHIC PARAMETERS: THE SEARCH FOR
A GLOBAL TRIGGER MECHANISM

1. Immediate Causes for Late Miocene Hiatuses in the
Central Basin of the Equatorial Pacific:
Nondeposition, Dissolution or Erosion?

The results presented in Chapters III and IV are
inconclusive as to the specific immediate cause(s?) of the
hiatuses under investigation. Nondeposition, dissolution,
and erosion are potential explanations. A combination of
these mechanisms, however, is also possible.

Because NH6 is the best documented hiatus recorded in
both cores (either as a "real" hiatus or as a depositional
disturbance), and because data related to this event are the
most reliable in terms of stratigraphic resolution, the NH6
hiatus interval was chosen for further study.

Intensification of bottom-current activity as the immediate
cause for the occurrence of NH6 in the Central Basin is the
least complicated explanation. This interpretation is based
mainly on the following observations: (1) in both cores no
evidence for dissolution of radiolarian tests was found; (2)
distinct orientations of the principal susceptibility
directions, although unusual and not of primary origin,
correlate with the NH6 occurrence in core K78-5-10; (3) reworked drift deposits of late Miocene age are reported in the close vicinity of site M70-17 (Hollister et al., 1974; Lonsdale, 1981); (4) distinct coarsening of magnetic grains in NH6-related core intervals of both cores was observed; (5) mean silt size coarsens across the NH6-related M70-17 core interval; (6) there is a relative increase in the allochthonous zeolite clinoptilolite across the NH6-related M70-17 core interval; (7) oscillations of IRM and ARM data across hiatus NH6 in core K78-5-10 were observed; and (8) there are high sedimentation rates across the NH6-related barren zone in core M70-17 despite the documented occurrence of a corresponding hiatus in core K78-5-10.

If intensified bottom-current activity is indeed the immediate cause for the occurrence of hiatuses in the Central Basin, it is not surprising that current-related events are recorded as "real" hiatuses at one location and as depositional disturbances at another. The hydrodynamic setting at each core location determines if intensification of bottom currents will cause erosion, drift deposition or only winnowing. Therefore, it is not surprising that the NH6 event is recorded as a "real" hiatus in core K78-5-10 and as an apparent drift deposit in core M70-17. If bottom-current accelerations cause erosion and result in "real" hiatuses, then the duration of the same hiatuses in
individual cores, although being caused by the same "event", may differ from location to location. Apparent "durations" of hiatuses (the time not represented by sediment) may have been caused by erosional episodes of much shorter lengths of time (and spatially different intensities). Thus, the ages of the cessations of the hiatus intervals or related depositional disturbances are more reliable than ages linked to their onsets. As a consequence, the following comparisons between hiatus occurrences and paleoceanographic parameters are solely based on datums associated with the cessation of hiatus intervals and related depositional disturbances.

The locally differing depositional record of erosional events can explain why hiatus NH5 in core K78-5-10 was not detected in age vs. depth plots. The graphic correlation plots (Fig. 5) indicate continuous sedimentation in core K78-5-10 during the time interval corresponding to the NH5 hiatus. Magnetic grain-size analyses (Fig. 23), however, show a distinct coarsening of magnetic grains at a core depth of 1306-1308.5 cm, correlating approximately in age with the cessation of NH5 in core M70-17 (Fig. 5). X-radiographs taken across this core interval in core K78-5-10 reveal some faint textural changes. These sedimentological changes, apparently related to NH5 further indicate that an erosional event can be recorded as a hiatus.
sensu stricto (that is by a gap in the stratigraphic succession) at one site, and by depositional disturbances at another.

2. Redefinition of "Hiatus Stratigraphy"

Based on historical reasons (Barron and Keller, 1982), the terms "Hiatus Stratigraphy", "NH6", "NH5", etc., have been used throughout this study. The presented data suggest that the same paleoceanographic event can be recorded in different ways at different sites. A paleoceanographic change thus can cause reworking/erosion at one site and winnowing, magnetic grain-size coarsening, or carbonate dissolution at another. Some of these sedimentary expressions do not represent a "real" hiatus, that is, there is no gap in the stratigraphic succession. Although hiatuses could be correlated with seemingly synchronous depositional changes, the application of the aforementioned "hiatus" terminology for hiatus-related depositional signals would be misleading. It is thus suggested to use the term "Event Stratigraphy" and rename Neogene hiatuses "Neogene Events" ("NE") or apply the original "NH" terminology only to "real" hiatuses (where time is not recorded and unconformities exist) and label related depositional changes of shorter duration, for example, "NH Markers". An alternative would be to redefine the term "Hiatus Interval"
(as it has been done for this study; Chapter I) and use it to describe the impact of a paleoceanographic event in both the sense of time (to characterize the duration of a hiatus), and in the sense of physical core depth (to mark a depth level that contains a depositional signal which can be linked to a "real" hiatus).

3. Depositional Changes in the Late Miocene and Global Paleoceanographic Parameters: Do Sea-Level Changes Control Hiatus Occurrences?

If intensified bottom-current activity is responsible for the occurrence of NH6 in abyssal siliceous clay provinces of the Central Basin, it is possible that presumably contemporaneous hiatuses and seismic reflectors in carbonate-rich sections of the Central Pacific (Mayer et al., 1985; Mayer et al., 1986) are genetically related. Furthermore, apparently synchronous depositional changes have been recorded in sediments from the continental slope and rise (Barron, 1986; Poag and Ward, 1987). If the hiatuses and depositional changes in these different oceanographic settings can indeed be interrelated, then hiatus occurrence is not only sediment-independent, but also depth-independent. Thus, a global trigger mechanism must be invoked, which apparently results in erosion or depositional disturbances in siliceous sections of the deep sea,
dissolution in shallower, carbonate-rich sections, and nondeposition or erosion on the shelves. Can a comparison with depositional data from other sites and global parameters such as global eustatic sea-level curves, carbonate-dissolution curves, and oxygen and carbon isotope values shed light on the nature of such a mechanism?

To address the question of apparent synchronicity of hiatuses and related depositional disturbances in different environmental settings, the age-depth plots of cores K78-5-10 and M70-17 were compared with those for DSDP Sites from the East-Central Equatorial Pacific and Central Indian Ocean (Fig. 31). DSDP Sites 572 (01°26.09'N, 113°50.52'W, 3893 m) and 573 (0°29.91'N, 133°18.57'W, 4301 m) are located in the Eastern Pacific, beneath the equatorial high-productivity belt, and Site 238 is on the Central Indian Ocean Ridge (11°09'S, 70°31'E, 2844 m). These DSDP Sites actually do not contain hiatus NH6, but the timing of its cessation (or the cessation of a related depositional disturbance) in cores K78-5-10 and M70-17 correlates with a distinct acceleration of sedimentation rates at all sites under investigation, with the exception of M70-17. At the location of core M70-17 the sedimentary record is severely influenced by intensive reworking and drift deposition. It is important to mention that the acceleration of sedimentation rates at DSDP Site 238 is preceded by a core
FIGURE 31. Comparison between the late Neogene time scale (A), graphic correlation plots of cores K78-5-10 and M70-17, DSDP Sites 238, 572, and 573 (B), relative coastal onlap (C) and eustatic sea-level (D) curves, a percent carbonate curve (E), and benthonic δ¹⁸O (F) and δ¹³C (G) isotope curves. Paleomagnetic time scale and geochronology after Berggren et al. (1985). Datums used for graphic correlation plots are listed in Tab. 1. Diatom datums are represented by circles, radiolarian datums by triangles, and paleomagnetic datums by squares, respectively. Symbol number corresponds to event number in Tab. 1. Shading indicates hiatus intervals. Data for DSDP Site 238 from Vincent et al. (1980) and for Sites 572 and 573 from Mayer, Theyer, et al. (1985). Relative coastal onlap and eustatic sea-level curves after Haq et al. (1987). In the coastal onlap curve L indicates landward ("onlap") and B basinward ("offlap") direction. In the eustatic sea-level curve H indicates sea-level highstands and L sea-level lowstands. The percent carbonate curve is a composite of data from DSDP Sites 71 (Tracey, Sutton, et al., 1971), 158 (Bode and Cronan, 1973), and 572 (Pisias and Prell, 1985). Seismic reflector "LM-B" after Mayer et al. (1986). Benthonic δ¹⁸O and δ¹³C isotope curves are based on data from DSDP Site 588 (Kennett, 1986).
interval void of radiolarians (Vincent et al., 1980). This interval, although not described as a hiatus, seems to reflect the NH6 event. NH5 has been recorded at DSDP Site 573 (Mayer, Theyer, et al., 1985). Its duration appears to be shorter here as compared to NH5 of core M70-17. The relatively shorter duration of NH5 at Site 573 may be due to both a different hydrodynamic setting and a lack of sufficient datum events, limiting stratigraphic resolution.

The age vs. depth plots at the different sites were also compared with newly revised coastal onlap and eustatic sea-level curves (Haq et al., 1987), and the global paleoceanographic parameters of carbonate-dissolution curves and benthic oxygen and carbon isotopic curves (Fig. 31). The cessation of hiatus NH6 (and of related depositional disturbances) correlates (Fig. 31) with a coastal offlap event around 6.3 Ma (Loutit and Keigwin, 1982; Haq et al., 1987). This offlap event is interpreted as indicating fall in sea level. Cessation of NH6 also correlates with a global acceleration of sedimentation rates (Fig. 31) and with the well known "Magnetic Chron-6 carbon shift" (Keigwin, 1979; Bender and Keigwin, 1979; Loutit and Kennett, 1979; Haq et al., 1980; Keigwin and Shackleton, 1980; Vincent et al., 1980; Bender and Graham, 1981; Savin et al., 1981; Vincent et al., 1985; Berger and Vincent, 1986) towards lighter $\delta^{13}$C values in both benthonic and
planktonic foraminiferal calcite. The time period preceding the carbon shift, and spanning the duration of NH6, is marked (Fig. 31) by dissolution of carbonate (Mayer, Theyer, et al., 1985). This carbonate dissolution event resulted in the development of a distinct, regionally traceable seismic reflector ("1M-B") in the Equatorial Pacific (Mayer et al., 1985; Mayer et al., 1986) which correlates with NH6. Reflector 1M-B, as well as several similar seismic horizons in Neogene sediments, are believed to coincide with times of sea-level highstands (Mayer et al., 1986). The cessation of NH6 is further marked by a shift towards heavier $\delta^{18}O$ values (Fig. 31) in benthonic foraminifera suggesting an increase in ice volume (Keigwin, 1979; Miller and Fairbanks, 1985). Such a relationship lends additional support to the correlation of hiatus cessation and lowering of sea level.

In light of the data presented for NH6 (Fig. 31), it appears plausible that hiatuses and related depositional disturbances in deep-sea sections occur during relative sea-level highstands and cease during sea-level falls. A relationship between eustatic sea-level changes and stratigraphic interruptions was first proposed by Barrell (1917). Rona (1973) postulated a link between deep-sea erosion and sea-level lowstands because of oceanic circulation changes. Later findings (e.g., Loutit and
Kennett, 1981; Mayer et al., 1986; Keller et al., 1987) led to a hypothesis relating the occurrence of deep-sea hiatuses with relative sea-level highstands. Discussions concerning the relationship between sea-level changes and hiatus occurrences have been revived recently (Poag and Ward, 1987; Barron and Keller, submitted; Keller and Barron, submitted), because of the revised geochronology for the Cenozoic (Berggren et al., 1985; Barron et al., 1985) and the refined eustatic sea-level curve (Haq et al., 1987). Some hiatuses seem also to coincide with major tectonic events (Barron, in press).

Correlations between the cessations of hiatuses NH5 and NH3(NH4?) and the coastal onlap curve (Fig. 31) support the proposed link between sea-level falls and the termination of hiatus events. The cessation of both NH5 and NH3(NH4?) is, within the limits of stratigraphic resolution, associated with a coastal offlap event, which suggests hiatus generation during sea-level highstands.

4. Speculative Sea-Level Dominated Sedimentation Model

The correlation of the termination of hiatus event NH6 with a coastal offlap event (Fig. 31) suggests a genetic relationship between hiatus occurrence and sea-level highstands. Therefore, a sea-level dominated sedimentation model is illustrated (Fig. 32), which tries to explain the
FIGURE 32. Illustration of a sea-level dominated depositional model, which ascribes the generation of hiatuses in various environments and at different water depths to relative sea-level highstands (A), and their cessation to lowered sea level (B).
A

Terrigenous sediments trapped on shelves

Deposition of organic material, nutrients, and carbonate on flooded shelves

Condensed sections, hiatuses

Less nutrients cause decrease in surface-water productivity

Disturbance of carbonate balance, shoaling of CCD, hiatuses

Decrease in biogenous sediment input

Sedimentary starvation of deep basins, hiatuses

B

Weathering and erosion of organic material, carbonate and nutrients

Terrigenous sediments bypass shelf

Subaerial hiatuses

Additional nutrients cause increase in surface-water productivity

Increase in biogenous sediment input

Deepening of CCD

Bulk sedimentation-rate increase
occurrence of marine hiatuses in different depositional environments during sea-level highstands. Basically, the model follows the idea of basin-shelf fractionation of Berger (1970) and a similar scenario outlined by Loutit and Kennett (1981). Admittedly, this simplified model does not take into account climate-related feedback mechanisms (Berger, 1982) and other possible complications. Nevertheless, it can explain the apparently synchronous occurrence of hiatuses in different depositional settings, regardless of water depth.

It has been widely accepted that short-term variations in sea level during the late Neogene are caused by changes in continental ice volume (e.g., Miller and Fairbanks, 1985). Falling sea levels (reflecting increase in ice volume) thus coincide with a shift of $\delta^{18}O$ towards heavier values in benthonic foraminifera (Keigwin, 1979), as observed at the cessation of hiatus event NH6 (Fig. 31). High sea-level stands trap terrigeneous sediments on the shelf areas (Hay and Southam, 1977; Davies and Worsley, 1981), causing depositional starvation of deep basins and apparently result in slowed-down sedimentation rates or the formation of hiatuses. During times of transgression, sediment depocenters move landward and cause condensed sections and hiatuses on basinward shelf areas (Haq et al., 1987). Deposition of carbonates on the available shelf space
(Hay and Southam, 1977; Davies and Worsley, 1981) draws on the ocean's carbonate budget, because it exceeds the input of calcium from rivers (Hay and Southam, 1977) and thus raises the carbonate compensation depth (Milliman, 1974; Hay and Southam, 1977). Shoaling of the CCD then results in the dissolution of carbonate-rich sediments in the deep sea and explains the occurrence of seismic reflectors associated with hiatus events (Mayer et al., 1985; Mayer et al., 1986). The flooded shelf areas also act as a temporary sink for organic carbon (Vincent et al., 1980; Loutit et al., 1983; Berger and Vincent, 1986) and nutrients (Broecker, 1982). Storage of $^{12}$C-enriched organic carbon on the shelves leads to heavier $\delta^{13}$C values in the carbonate of foraminifera shells. Such a relationship between $\delta^{13}$C values and transgressions has been documented by Woodruff and Savin (1985), based on the isotopic record of Miocene benthic foraminifera from the Pacific. The decrease in available nutrients, especially phosphorus, would furthermore limit biogenic productivity in pelagic surface waters (Broecker, 1982), provoking less flux of biogenic components to the deep sea. Such a change in surface waters would very significantly influence deposition in a siliceous brown-clay environment with its naturally low sedimentation rates. The reduced overall sediment influx to the deep sea during sea-level highstands (Worsley and Davies, 1979) makes oceanic deposits more susceptible to erosion. Erosion
during sea-level highstands might not be expected, because of the widely accepted assumption (e.g., Barron and Keller, 1982; Keller and Barron, 1983) that production and intensification of Antarctic Bottom Water (AABW) occurs during periods of polar cooling, that is when ice volume is large and thus sea level low. A recent overview of modern deep-water circulation (Corliss et al., 1986), however, could not support such a correlation and argues for AABW production during both glacial and interglacial times.

If one accepts the scenario outlined above, then it is obvious that, depending on the depositional setting, episodes of transgression will result in different individual immediate causes for hiatus occurrence. Although believed to be genetically interrelated and contemporaneous (within the limits of stratigraphic resolution) global hiatuses during sea-level highstands seem to be caused by the following: (1) nondeposition (or condensed sedimentation) on basinward shelf areas; (2) dissolution of carbonate-rich sections affected by the shoaling of the CCD; and (3) erosion in siliceous-clayey environments.

Following an episode of transgression, a drop in sea level would reverse the sedimentation pattern. The exposed shelves no longer act as sediment traps and the continentally derived sediment load will bypass directly to the deep basins. Consequentially, bulk sedimentation will
increase (Worsley and Davies, 1979), owing to both higher terrigenous influx and increased surface-water productivity because of the addition of nutrients (Broecker, 1982). As can be seen in Fig. 31, the cessation of hiatus event NH6 is followed by accelerated sedimentation rates in the Indo-Pacific realm (see also Vincent, 1977; Vincent et al., 1980). Removal of carbonates from the shelves, as well as the non-availability of shelf space for further carbonate sedimentation (Hay and Southam, 1977), is assumed to deepen the CCD and increase carbonate preservation. Such a deepening of the CCD associated with carbonate input from the shelves has been reported for the late Miocene (Berger and Winterer, 1974; van Andel, 1975). A positive correlation between decreases in flooded shelf areas during sea-level lowstands and increases in pelagic carbonate sedimentation has also been established by Davies and Worsley (1981).

The cessation of hiatus NH6 correlates (Fig. 31) with the "Magnetic Chron-6 carbon shift" (Vincent et al., 1980), which is believed to be the result of some complex interaction between a global regression, oceanic circulation changes, and biological processes (Vincent et al., 1980; Loutit et al., 1983; Berger and Vincent, 1986). This shift towards lighter δ^{13}C values, which is also known as the "Messinian carbon shift" (Berger and Vincent, 1986), is recorded in both planktonic and benthonic foraminifera
(Loutit et al., 1983). The carbon shift is discussed in detail by Berger and Vincent (1986). In particular, the carbon shift has been attributed to the following factors (Vincent et al., 1980; Berger and Vincent, 1986), in declining order of relative importance: (1) exchange of the ocean with an external organic carbon reservoir leading to increased input of $^{12}$C-enriched organic matter (in the form of eroded soil and oxidized organic-rich marine sediments); (2) basin-basin fractionation between the Indo-Pacific and Atlantic, owing to increased production of North Atlantic Deep Water (NADW); and (3) "vital effects" causing separation of values of individual benthic foraminifera species with time because increases in fertility drive disequilibrium. As already outlined, increases in productivity would be indeed expected after lowering of sea level, because of the introduction of nutrients to the oceanic system (Broecker, 1982). After an increase in productivity, however, a shift towards heavier $^{13}$C values, not lighter ones, would be expected. It has been shown (Froelich et al., 1981) that hemipelagic continental-rise sediments have a carbon-to-phosphorus ratio approximately three times higher than deep-sea organic sediments. Therefore, during times of regressions the ($^{12}$C-enriched) carbon content of such organic matter seems to override the simultaneous phosphorus input (which increases fertility), resulting in the observed lighter $^{13}$C values.
The apparent increase of presumably terrigenous minerals associated with the occurrence of hiatus NH6 in cores K78-5-10 (Fig. 9) and M70-17 (Fig. 10), seems to contradict the proposed model. The sea-level rise considered to be responsible for hiatus NH6 should result in the entrapment of terrigenous sediments and thus less influx to the deep sea. Furthermore, if the sea-level rise is driven by a decrease in continental ice volume, a relatively humid continental climate is expected (Janecek and Rea, 1983). During times of humid climates the flux of eolian terrigenous material is also reduced (Janecek and Rea, 1983; Janecek, 1985). The apparent increase in terrigenous components is based on the assumption that quartz, feldspars, and a combination of illite and kaolinite/chlorite are of terrigenous origin. This assumption may be oversimplified, because quartz and feldspars could be derived from volcanic sources (Peterson and Goldberg, 1962). It is also possible that the observed mineralogical signal correlates with falling sea level and that the mineralogical record indicating sea-level highstand (that is decrease in terrigenous components) has been eroded. Furthermore, all interpretations are based on changes in the relative abundances of minerals, not absolute ones. The interpretation of the clay-size fraction, for example, is particularly complicated by this factor. Changes in the amount of smectite derived from local
volcanic sources may override variations in the eolian input of all the clay minerals, yielding apparent increases or decreases of terrigenous components.

The outlined depositional model cannot only explain hiatus NH6 occurrence and changes in related paleoceanographic parameters. The cessation of hiatus NH5 also correlates with a coastal offlap event and its apparent duration seems to be linked to a period of carbonate dissolution (Fig. 31), although it is allegedly not associated with a seismic reflector in the Equatorial Pacific (Mayer et al., 1986). Cessation of NH5 furthermore correlates with relatively heavier $\delta^{18}O$ and relatively lighter $\delta^{13}C$ values, as observed across the NH6 event. A comparison with the NH3(NH4?) hiatus interval leads to similar findings. At DSDP Sites 572 and 71, however, NH3(NH4?) cannot be linked to a major carbonate dissolution event (Fig.31). Nevertheless, both hiatuses NH3 and NH4 are known to be associated with a distinct seismic reflector in the central equatorial Pacific (Mayer et al., 1986).
CONCLUSIONS

Integrated magnetobiostratigraphic analyses of two piston cores from the Central Basin of the Pacific revealed the occurrence of three middle-late Miocene hiatuses in siliceous brown clays. Mineralogical, sedimentological, and studies concerned with rock-magnetic properties of the sediments were conducted to determine depositional changes which might shed light on the origin of the hiatuses. These investigations were complemented by comparisons between hiatus occurrence in the Central Basin and graphic correlation plots of selected DSDP Sites in the Indo-Pacific realm, sea-level curves, carbonate-content curves, and stable oxygen and carbon isotope curves. The main conclusions reached in this study are listed below.

(1) The hiatuses detected in piston cores K78-5-10 and M70-17 from the Central Basin of the Pacific correlate with global hiatuses NH3(NH4?), NH5, and NH6 of Barron and Keller (1982). Based on data from the two piston cores, the durations of these hiatuses in the Central Basin are determined to be 12.6-10.4, 8.9-7.9, and 7.1-6.3 Ma, respectively.

(2) X-radiographs link hiatus intervals NH5 and NH6 to faint but distinct textural changes. In core M70-17 hiatus
NH₃(NH₄?) is associated with a coarse volcanic ash layer.

(3) X-ray diffraction analysis (XRD) across the hiatus intervals reveals distinct changes in relative mineral abundances. Ratio plots of selected minerals across the core interval of NH₆ of K78-5-10 suggest a relative increase in the abundance of presumably terrigenous components. An increase in the percentage of the zeolite clinoptilolite across the NH₆-related barren zone of core M70-17 is interpreted as an indication for reworking and drift deposition. The barren zone of core M70-17 is also marked by a relative increase in the abundance of manganese micronodules and a lack of radiolarian tests.

(4) Coarsening of mean silt sizes across hiatus-related core intervals seems to indicate intensification of reworking and winnowing processes. The relationship between results based on silica-free and undigested samples, however, is not quite clear.

(5) Core intervals related to NH₃(NH₄?) and NH₆ are marked by an increase in bulk magnetic susceptibility. Across the NH₆-related barren zone of core M70-17 this susceptibility increase seems to have been caused by a parallel increase in the relative abundance of manganese micronodules. The interpretation of anisotropy of magnetic susceptibility (AMS) parameters is less conclusive. The
observed perpendicular or near-perpendicular orientations of axes of maximum susceptibility relative to the bedding plane in samples across the NH6-interval of core K78-5-10 cannot be explained. Isothermal remanent magnetization (IRM) acquisition and demagnetization curves identify magnetite as dominant magnetic carrier. Growth of authigenic magnetic phases, coring disturbances, and bioturbation are considered to be unlikely explanations for the distinct and systematic susceptibility axis alignments.

(6) All hiatus-related core intervals are marked by an increase in coarseness and concentration of magnetic grains. This increase in size and concentration of magnetic carriers is interpreted as additional evidence for intensified erosion, reworking, and winnowing during hiatus intervals. The cessation of hiatus interval NH5 in core M70-17 apparently correlates with a coarsening of magnetic grains in a corresponding core interval of K78-5-10. Thus, although graphic correlation plots did not reveal the presence of NH5 in core K78-5-10, a related depositional disturbance seems to have been recorded.

(7) NRM, ARM/K, ARM, and IRM curves show distinct variations across the hiatus-related core intervals.

(8) Intensified erosion and reworking seem to be the immediate causes for hiatuses in deep-sea sediments in the
Central Basin. A comparison between the cessation of hiatus NH6 and the global coastal onlap curve, however, points at relative sea-level changes as a potential trigger mechanism.

(9) In addition to a coastal offlap event the cessation of the NH6 hiatus correlates with the cessation of a carbonate-dissolution interval (which resulted in a distinct seismic reflector), an increase in global sedimentation rates, a distinct carbon shift towards lighter isotope values, and an oxygen isotope shift towards heavier values. Based on these correlations a sea-level dominated depositional model is proposed which explains the occurrence of presumably contemporaneous hiatuses in different depositional environments and at various water depths. The model ties the occurrence of hiatuses in marine sediments to relative sea-level highstands and their cessations to sea-level falls.
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