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Coarse fraction fluctuations in pelagic carbonate sediments from the tropical Indian Ocean: A 1500-kyr record of carbonate dissolution

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Abstract. We examined coarse fraction contents of pelagic carbonates deposited between 2000- and 3700-m water depth in the tropical Indian Ocean using Ocean Drilling Program (ODP) sites 722 (Owen Ridge, Arabian Sea) and 758 (Ninetyeast Ridge, eastern equatorial Indian Ocean), and four giant piston cores collected by the French R/V *Marion Dufresne* during the SEYMAMA expedition. Over the last 1500 kyr, coarse fraction records display high-amplitude oscillations with an irregular wavelength on the order of ~500 kyr. These oscillations can be correlated throughout the entire equatorial Indian Ocean, from the Seychelles area eastward to the Ninetyeast Ridge, and into the Arabian Sea. Changes in grain size mainly result from changes in carbonate dissolution as evidenced by the positive relationship between coarse fraction content and a foraminiferal preservation index based on test fragmentation. The well-known "mid-Brunhes dissolution cycle" represents the last part of this irregular long-term dissolution oscillation. The origin of this long-term oscillation is still poorly understood. Our observations suggest that it is not a true cycle (it has an irregular wavelength) and we propose that it may result from long-term changes in Ca^{++} flux to the ocean. Sites 722 and 758 $\delta^{18}O$ records provide a high-resolution stratigraphy that allows a detailed intersite comparison of the two coarse fraction records over the last 1500 kyr. Site 722 (2030 m) lies above the present and late Pleistocene lysocline. The lysocline shoaled to the position of site 758 (2925 m) only during the interglacial intervals that occurred between about 300 and 500 ka (Peterson and Prell, 1985a). Despite these supralysoclinal positions of the two sites, short-term changes in coarse fraction contents are correlatable from one site to another and probably result from regional (or global) dissolution pulses. By stacking the normalized coarse fraction records from sites 722 and 758, we constructed a Composite Coarse Fraction Index (CCFI) curve in which most of the local signals cancelled out. The last 800 kyr of this curve appear to compare extremely well with the Composite Dissolution Index curve from core V34-53 (Ninetyeast Ridge), which unambiguously records past variations of carbonate dissolution in the equatorial Indian Ocean (Peterson and Prell, 1985a). In the late Pleistocene the CCFI variations are mainly associated with glacial-interglacial changes. They show strong 100 and 41 kyr periodicities but no clear precession-related periodicities. As proposed earlier by Peterson and Prell (1985a), the lack of precession frequencies may suggest that the regional carbonate dissolution signal is driven by changes in deepwater circulation. We cannot totally reject the possibility, however, that low temporal resolution and/or bioturbation degrade somehow the precessional signal at ODP sites 722 and 758. In contrast, spectral density of dissolution cycles in the giant (53 m long) piston core MD900963 (Maldives area) displays clear maxima centered on the precession frequencies (23 and 19 kyr^{-1}) as well as on the 29-kyr^{-1} frequency but shows little power at the 100-kyr^{-1} frequency. These high-frequency changes most probably result from changes in surface productivity associated with monsoon variability. Dissolution at this site may be ultimately controlled by the oxidation of organic matter which appears to be incorporated into the sediments in greater quantity during periods of weak SW monsoon and/or increased dry NE monsoon.

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Introduction

Deep-sea carbonate sediments cover about one half of the total oceanic floor [*Berger et al.*, 1976; *Biscaye et al.*, 1976; *Kolla et al.*, 1976] and act as a large and reactive reservoir for carbon dioxide [*Broecker and Peng*, 1982, 1987; *Sundquist and Broecker*, 1985]. Understanding temporal and spatial changes in carbonate preservation is of key importance for testing the numerous models which seek to explain past changes in atmospheric pCO_2 through changes in the oceanic carbon-carbonate system [*Broecker, 1982*; *Broecker and Peng*, 1982; *Boyle*, 1988; *Keir*, 1988].

Since the pioneering work of *Arrhenius* [1952], it has been clearly established that carbonate contents of equatorial pelagic sediments show cyclic variations associated with Pleistocene glacial-interglacial changes [e.g., *Arrhenius*, 1952; *Broecker*, 1971; *Berger*, 1973; *Gardner*, 1975; *Moore et al.*, 1982; *Crowley*, 1985; *Peterson and Prell*, 1985a; *Vincent*, 1985; *Arrhenius*, 1988; *Farrell and Prell*, 1989; *Berger*, 1992]. These fluctuations have been correlated over large distances in the Pacific [*Hays et al.*, 1969; *Vincent*, 1981; *Karlin et al.*, 1992] and into the Indian Ocean [*Volat et al.*, 1980; *Vincent*, 1985] and have been used as powerful geochemical correlation tools. First interpreted in terms of changing surface productivity [*Arrhenius*, 1952], the Indo-Pacific carbonate fluctuations have been later reinterpreted in terms of dissolution changes, with increased dissolution during interglacials and enhanced preservation during glacials [*Berger*, 1973; *Thompson and Saito*, 1974; *Volat et al.*, 1980; *Farrell and Prell*, 1989] (see the review by *Berger* [1992]). In the Atlantic and Southern Oceans, dissolution generally intensified during glacial periods [*Gardner*, 1975; *Bé et al.*, 1976; *Crowley*, 1983; *Howard and Prell*, 1990]. This asymmetry between the Atlantic and the Indo-Pacific has been attributed to changes in basin-basin fractionation resulting from variations in North Atlantic Deep Water (NADW) formation [*Berger*, 1970; *Volat et al.*, 1980; *Crowley*, 1985].

Most studies dealing with carbonate dissolution history have focused on deep-sea records below the calcite lysocline, where past changes in carbonate saturation of seawater have resulted in high-amplitude changes in carbonate content or in calcareous microfossil preservation [*Berger*, 1973; *Thompson and Saito*, 1974; *Luz and Shackleton*, 1975; *Adelseck*, 1977; *Mayer*, 1979; *Peterson and Prell*, 1985a; *Farrell and Prell*, 1989]. There is far less data available for carbonate dissolution cycles in "shallower" water depths (above the lysocline) [*Droxler et al.*, 1983, 1990; *Haddad*, 1986]. Although supralysoclinal waters are supersaturated with respect to calcite, sediments deposited above the lysocline undergo significant carbonate dissolution. It has been proposed that this dissolution results from the decay of organic matter [*Emerson and Bender*, 1981; *Peterson and Prell*, 1985b], with the release of CO_2 reducing the pore water carbonate ion concentration to a level lower than at the equivalent water depth in the water column. Carbonate dissolution patterns at intermediate and shallow water depths are of key importance in determining whether past changes in the oceanic carbon-carbonate system indicate a redistribution of properties within the ocean during glacial/interglacial oscillations [e.g., *Boyle*, 1988], or indicate transfers between global carbon or carbonate reservoirs (shelf, soil, and biosphere) that affect the entire water column in a similar way [e.g., *Berger*, 1970; *Shackleton*, 1977; *Broecker*, 1982].

With increasing dissolution, foraminiferal tests tend to break down into small fragments [*Berger*, 1968; *Bé et al.*, 1975; *Thunell*, 1976]. Detailed depth transect studies conducted on the Ontong Java Plateau (western equatorial Pacific) and upon the Ninetyeast Ridge (eastern equatorial Indian Ocean) have clearly established that coarse fraction contents show a systematic relationship with water depth, which mainly results from the fragmentation of foraminifera and the subsequent transfer of fragments to the finer size fraction as dissolution proceeds [*Johnson et al.*, 1977; *Berger et al.*, 1982; *Peterson and Prell*, 1985b]. Decrease in coarse fraction content is a much more sensitive indicator of dissolution than carbonate content, because foraminiferal tests can break down after a small amount of carbonate has been lost to dissolution. For instance, some 60% of the whole sand-sized planktonic foraminifera has already broken up at the lysocline level upon Ninetyeast Ridge, whereas no more than 20 to 30% of total carbonate has been estimated to have been lost to dissolution [*Peterson and Prell*, 1985b]. Thus coarse fraction records should provide relatively easy to obtain qualitative information on the dissolution history in supralysoclinal carbonate deposits. The purpose of this study is to present a detailed coarse fraction stratigraphy of Pleistocene carbonate sediments from the tropical Indian Ocean to test the potential of grain size data as a correlation tool and as a proxy for studies of supralysoclinal carbonate dissolution.

Study Area, Material, and Methods

We selected four giant piston cores (30 to 53 m in length) retrieved during the SEYMAMA expedition of the French R/V *Marion Dufresne* (cores MD900938, MD900940, MD900949, and MD900963) and two Ocean Drilling Program (ODP) sites (sites 722 and 758). These pelagic carbonate sections are distributed over much of the tropical Indian Ocean in water depths ranging from 2030 to 3700 m (Figure 1 and Table 1). They provide long, continuous, and undisturbed Plio-Pleistocene coarse fraction records.

Sites 722 (Owen Ridge, Arabian Sea) and 758 (Ninetyeast Ridge, eastern equatorial Indian Ocean) are especially suitable for studying temporal changes in coarse fraction content above the lysocline. Sites 722 (2030 m) and 758 (2925 m) lie about 1300 and 900 m above the present foraminiferal lysocline in the Arabian Sea [*Cullen and Prell*, 1984] and in the eastern equatorial Indian Ocean [*Peterson and Prell*, 1985b], respectively. On Ninetyeast Ridge the present foraminiferal lysocline (at 3800 m) is close to the saturation horizon with respect to calcite in the water column [*Peterson and Prell*, 1985b]. The lysocline position in the Ninetyeast Ridge area varied between about 2800 m and 4200 m during the last 800 kyr, that is to say, plus 400 m or minus 1000 m around its present position in the water column [*Peterson and Prell*, 1985a]. The lysocline shoaled to the position of site 758 only during the interglacial intervals that occurred between about 300 and 500 ka. If we consider that the lysocline position at Owen Ridge may have roughly fluctuated with the same amplitude, then site 722 probably remained in a supralysoclinal position during the entire late Pleistocene.

At sites 722 and 758, continuous sedimentary sections have been previously constructed by splicing across recovery gaps the 9.5-m-long hydraulic piston cores retrieved in adjacent holes from the same site [*Murray and Prell*, 1991; *Farrell and Janecek*, 1991, respectively]. Sedimentological data (proportion of fraction > 150 μm obtained by wet-sieving and $CaCO_3$ content) and oxygen isotope data obtained on planktonic foraminifera were also taken from *Murray and Prell* [1991] and *Farrell and Janecek* [1991].

In the SEYMAMA cores, sediments were sampled at a 10-cm interval. Carbonate content was measured on cores MD900938, MD900949, and MD900963 using a modified "Bernard calcimeter." We estimated the accuracy of the method

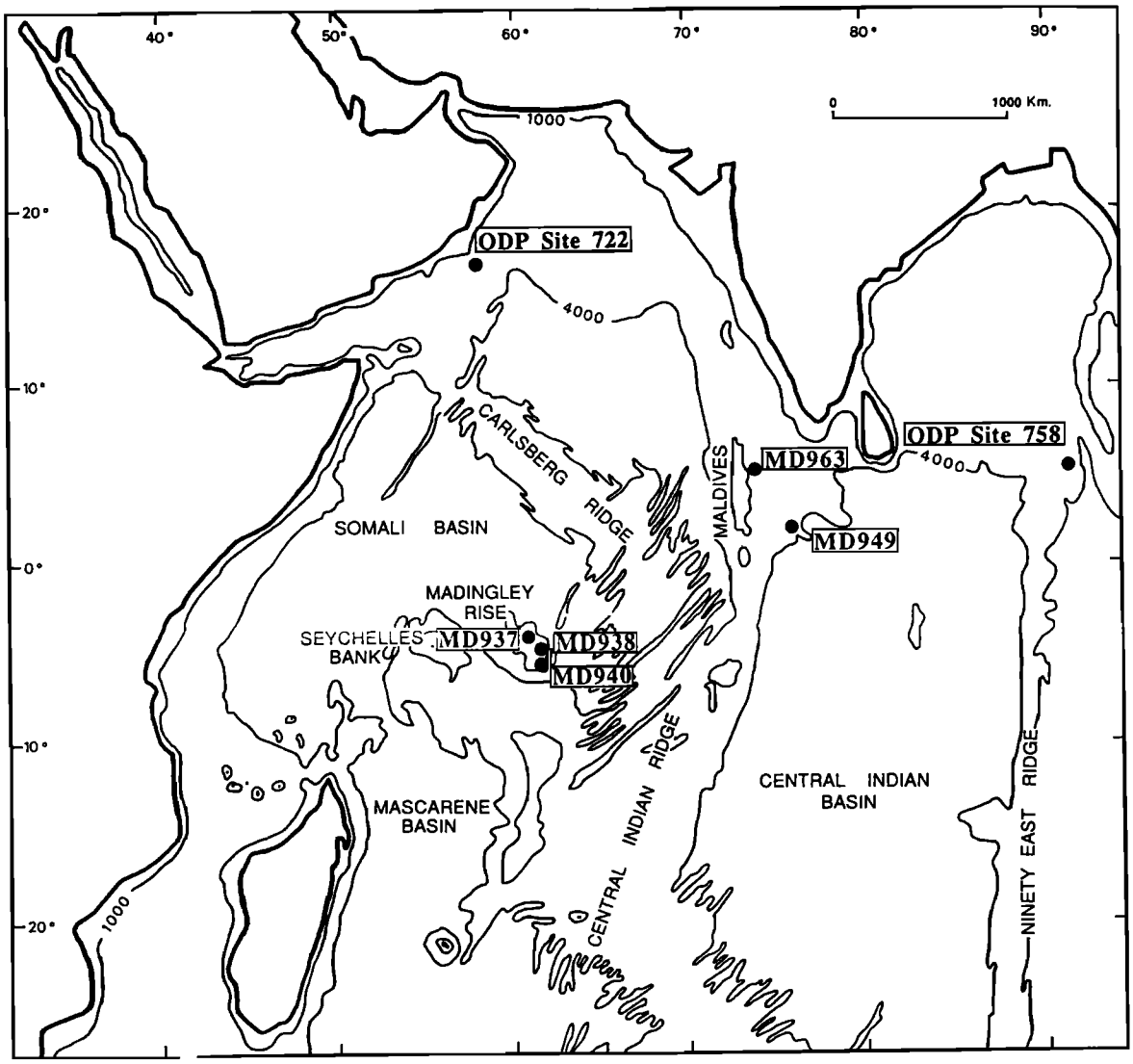

Figure 1. Location map showing positions of SEYMAMA giant piston cores and Ocean Drilling Program (ODP) sites used in this study. Bathymetric contours are in meters. Note that identification numbers of SEYMAMA cores were shortened in this figure for convenience (e.g., MD938 instead of MD900938).

Table 1. Location and Water Dept of Piston Cores Retrieved During the SEYMANA Cruise of the French R/V *Marion Dufresene* in the Equatorial Indian Ocean, and of Ocean Drilling Program Site 572 (Equatorial Pacific), and 722 and 758 (Tropical Indian Ocean)

Cores/Sites	Latitude	Longitude	Water Depth, m
MD900938	04°57'90" S	60°04'44" E	3430
MD900940	05°33'53" S	61°40'12" E	3190
MD900949	02°05'43" N	76°06'96" E	3700
MD900963	05°03'30" N	73°52'60" E	2446
ODP site 722	16°37'30" N	59°47'80" E	2030
ODP site 758	05°23'05" N	90°21'67" E	2925
ODP site 572	01°26'09" N	113°50'52" W	3900

to be within ± 2-3% for carbonate contents ranging from 50 to 100%. Coarse fraction contents were obtained in all SEYMAMA cores through wet-sieving over a 63-µm mesh sieve. In core MD900938, we also measured relative abundances of the > 250- and >350-µm fractions to test if there are important differences in the interpretation of coarse fraction records based on a different size. It is important to address correctly this point, since we will compare > 150-µm grain size records (sites 722 and 758) with > 63-µm records (SEYMAMA cores).

In cores MD900938, MD900949, and MD900963, we subsieved the fraction coarser than 63 µm using a 250-µm and a 355-µm mesh sieve in order to isolate large foraminifera for later stable isotopic studies. In these three cores, we examined the fragmentation of planktonic foraminifera in the 250- to 355-µm size fraction. Results are expressed in the form of a "preservation index" calculated as the percentage of whole

foraminifera to fragments plus whole foraminifera. In core MD900938 and MD900949, fragmentation was measured at 20-cm intervals, and in core MD900963 it was measured at 10-cm intervals.

Foraminiferal fragmentation is one of the most reliable indices of carbonate dissolution [*Bé et al.*, 1975; *Thunell*, 1976; *Peterson and Prell*, 1985b; *Le and Shackleton*, 1992]. The large size of particles in the 250- to 355-μm size fraction makes the counting of fragments a relatively easy task to perform. However, we were aware that foraminiferal fragments in such a small size range may not be numerous enough to get a good statistical measure of dissolution intensity. In core MD900938, to test the sensitivity of our index, we compared results obtained in the 250- to 355-μm size fraction with fragmentation measured in the > 150-μm size fraction, since this latter fraction is the most commonly used in foraminiferal fragmentation studies [e.g., *Peterson and Prell*, 1985b].

In core MD900963, oxygen stable isotopes were measured at the isotope geochemistry laboratory of the Centre des Faibles Radioactivités, Gif-sur-Yvette, France, on the planktonic foraminifera *Globigerinoides ruber* (white) picked from the 250- to 355-μm size fraction. The mass spectrometer is a Finnigan MAT-251 model with an automated carbonate preparation system (individual reaction chambers device). The data are all reported with respect to the Pee Dee belemnite standard. (Carbonate content, grain size, foraminifer fragments, and $\delta^{18}O$ data are available in the Calcium Carbonate Database, National Geophysical Data Center, Boulder, Colorado, at paleo@mail.ngdc.noaa.gov).

Paleomagnetic measurements were performed on standard cubes (8 cm^3) every 75 cm along core MD900963, with the mean sampling interval then reduced to about 12 cm around the Brunhes/Matuyama magnetic reversal. The measurements and stepwise demagnetizations of the natural remanent magnetization (NRM) were performed within the shielded room of the Institut de Physique du Globe de Paris, France. NRM values are low but declination is interpretable.

In the uppermost 10 m of core MD900963, organic carbon measurements were carried out at intervals of 10 cm at the Department of Geoscience of the University of Bremen, Germany, using a Carbone-Hydrogen-Nitrogen analyzer. Sample preparation and technical details are described elsewhere [*Müller et al.*, 1994].

Chronological Framework

Chronological Framework Based on Biostratigraphy and Magnetostratigraphy

At all sites, we were able to develop a coarse chronological framework for the past 2500 kyr based on magnetic reversal stratigraphy and/or calcareous microfossil events (Table 2).

Table 2. Calcareous Microfossil Datums and Magnetostratigraphic Events Used for the Depth-to-Time Conversions in SEYMANA Cores MD900938, MD900940, MD900949, and MD900963, and ODP Sites 722 and 758

Stratigraphic Events	Age, Ma		MD938, m	MD940, m	MD949, m	MD963, m	Site 722, m	Site 758, m
Acme *Emiliania huxleyi*	0.073	(1)	...	...	...	3.80	...	...
FO *Emiliania huxleyi*	0.28	(1)	...	...	...	...	7.95	...
LO *Pseudoemiliania lacunosa*	0.47	(1)	7.85	6.15	13.35	21.70	18.90	...
Brunhes/Matuyama	0.78	(2)	...	...	...	35.10	28.00	12.17
LO *Reticulofenestra asanoi*	0.88	(3)	13.55	11.15	22.35	38.38	...	...
Top Jaramillo	0.99	(2)	...	...	...	...	...	15.57
Rejuven. *Gephyrocapsa oceanica*	1.028	(4)	15.95	12.55	25.71	...	...	...
Base Jaramillo	1.07	(2)	...	...	...	...	40.35	16.77
LO large *Gephyrocapsa oceanica*	1.24	(4)	18.06	15.15	30.55	...	...	...
LO *Helicosphaera sellii*	1.47	(4)	19.17	17.71	35.00	41.09	...	...
LO *Calcidiscus macintyrei*	1.59	(4)	20.62	18.56		41.09	55.55	...
FO *Gephyrocapsa oceanica*	1.67	(4)	...	...		41.09	...	...
LO *Globigerinoides fistulosus*	1.73	(5)	...	...		41.20	...	...
Top Olduvai	1.77	(2)	...	...		41.20	...	...
LO *Discoaster brouweri*	1.95	(4)	24.45	20.65		51.00	69.50	...
LO *Discoaster assymetricus*	2.26	(6)	29.45	...			...	...
Matuyama/Gauss	2.60	(2)		...			89.40	38.53

Biostratigraphic analyses for SEYMAMA cores were performed at the Laboratoire de Géologie du Quaternaire (Centre National de la Recherche Scientifique, Marseille, France); biostratigraphic and magnetostratigraphic events for ODP sites are from the shipboard parties of legs 117 and 121, respectively [*Peirce et al.*, 1991; *Prell et al.*, 1991]. Depth positions in the final, spliced depth scales of sites 722 and 758 were taken from *Murray and Prell* [1991] and *Farrell and Janecek* [1991], respectively. FO, first occurrence, LO, last occurrence. Ages are (1) from *Berggren et al.* [1985], (2) from *Shackleton et al.* [1990], (3) from *Wei* [1993], (4) from *Raffi et al.* [1993], (5) estimated according to the disappearance of *G. fistulosus* at site 677 and the orbital chronology developed at this site by *Shackleton et al.* [1990], and (6) calculated by linear interpolation between ages of magnetic reversals from *Shackleton et al.* [1990] with position of the datum given by *Berggren et al.* [1985].

For sites 722 and 758, biostratigraphic and magnetostratigraphic events are those determined during shipboard analysis. Positions of these selected stratigraphic events in the spliced depth scale are given by *Murray and Prell* [1991] and *Farrell and Janecek* [1991]. In the SEYMAMA cores, the positions of calcareous microfossil events have been determined with a 10-cm sampling interval. The Brunhes/Matuyama magnetic polarity reversal has been identified in core MD900963 at a depth of 35 ± 0.3 m.

Time-to-depth conversions (Figure 2) were performed by assuming constant sedimentation rates between control points (Table 2). Ages of magnetic reversals are from estimates by *Shackleton et al.* [1990]. Ages of calcareous nannofossil events are derived from recent calibration of nannofossil biostratigraphy to oxygen isotope stratigraphies [*Raffi et al.*, 1993; *Wei*, 1993]. Some nannofossil events have been shown to be diachronous between low- and mid-latitude sites [*Raffi et al.*, 1993]. Because we are dealing with low-latitude sites in this paper, we use age estimates based on calibration to oxygen isotope stratigraphy from Ocean Drilling Program (ODP) site 677. The last occurrence of *Globigerinoides fistulosus* occurs in the oxygen isotopic stage 61 (at 70.6 m) in site 677 and is dated at 1.73 Ma based on the orbitally derived chronology developed at this site by *Shackleton et al.* [1990]. Depth-age curves for the SEYMAMA cores are shown in Figure 2.

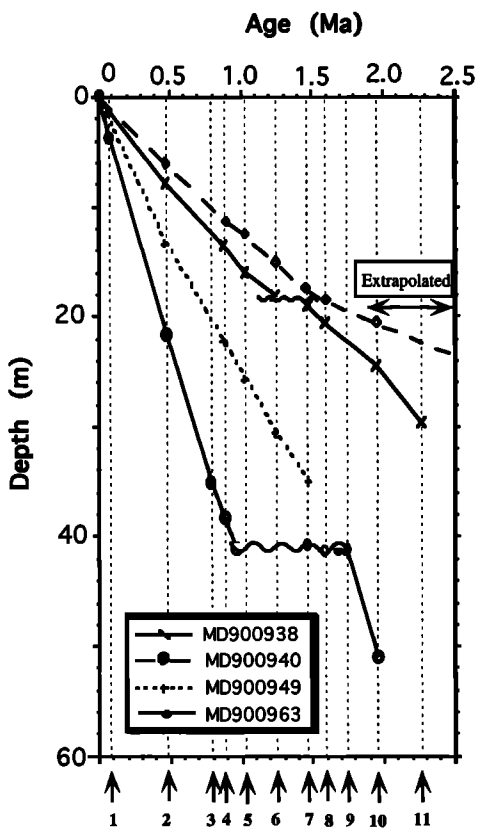

Figure 2. Age-depth plots showing sedimentation rates in cores MD900938 and MD900940 (Madingley Rise, western equatorial Indian Ocean) and in cores MD900949 and MD900963 (central equatorial Indian Ocean): 1, first occurrence (FO) *Emiliania huxleyi*; 2, last occurrence (LO) *Pseudoemiliania lacunosa*; 3, Brunhes/Matuyama reversal; 4, LO *Reticulofenestra asanoi*; 5, rejuvenation of large *Gephyrocapsa oceanica*; 6, decrease of the abundance in large *G. oceanica*; 7, LO *Helicosphaera sellii*; 8, LO *Calcidiscus macintyrei*; 9, LO *Globigerinoides fistulosus*; 10, LO *Discoaster brouweri*; 11, LO *Discoaster assymetricus*.

The depositional history of core MD900963 records a major hiatus at about 41 m (Figure 2), as shown by the concomitant disappearance at this level of *Helicosphaera sellii*, *Calcidiscus macintyrei*, and *G. fistulosus* and by the first appearance of *Gephyrocapsa oceanica*. The ages of these biostratigraphic datums span the time interval from 1.73 Ma to 1.47 Ma, which gives the minimum duration of the hiatus. We interpret, however, the hiatus as spanning 770,000 years between 1.73 Ma at its base (dated by the last occurrence of *G. fistulosus*) and 0.96 Ma at its top (dated by extrapolating downcore the sedimentation rates calculated between the Brunhes/Matuyama reversal and the last occurrence of *Reticulofenestra asanoi*).

In core MD900938, a step in the depth-age curve is observed between the last occurrence of *H. sellii* (1.47 Ma at 19.17 m) and the disappearance of large *Gephyrocapsa oceanica* (1.24 Ma at 18.06 m) (Figure 2). It is unlikely that there is continuous sedimentation in this short depth interval. We interpret this interval as containing a hiatus at about 18.60 m, spanning almost totally the time interval (230,000 years) between the two biostratigraphic events (Figure 2).

Oxygen Isotope Stratigraphy and Orbital Chronology

We constructed more detailed age models based on planktonic foraminiferal $\delta^{18}O$ in SEYMAMA core MD900963 and ODP sites 722 and 758. These age models will permit more precise intersite comparisons of the coarse fraction records from these three series. The chronology for these isotopic age models was developed from site 677 [*Shackleton et al.*, 1990]. Thus age models from sites 722 and 758 differ from those proposed by *Murray and Prell* [1991] and *Farrel and Janecek* [1991], respectively, which rested upon correlation to the SPECMAP curve [*Imbrie et al.*, 1984] and to the orbitally tuned record from ODP site 607 [*Ruddiman et al.*, 1989].

We aligned features in our $\delta^{18}O$ records with features in the site 677 record using the program "LINAGE" developed by D. Paillard and L. Labeyrie at the Centre des Faibles Radioactivités. In our correlation processes, the $\delta^{18}O$ transitions (stage boundaries) were given higher priorities than the centers of glacials and interglacials. The sedimentation rates are assumed to be constant between selected age control points and to change more or less abruptly at these control points. Such an assumption is probably quite correct when control points correspond to isotopic transitions, where environmental changes are supposed to affect the sedimentation rates decisively. Furthermore, as stated recently by *Le and Shackleton* [1992], locating a rapid change with a large amplitude (stage boundary) is probably more accurate than positioning a peak or trough with considerable width and small amplitude in comparison with analytical noise (centers of glacials or interglacials).

Correlations are easy in the Brunhes chronozones because of the characteristic glacial-interglacial 100-kyr cycles. In the

early Pleistocene, however, the low-amplitude 41-kyr $\delta^{18}O$ cycles typical of this time interval [e.g., *Prell*, 1982; *Ruddiman et al.*, 1986] are much more difficult to correlate with confidence. In this interval, the base of the Jaramillo subchronozone, recognized at sites 722 and 758, provides an additional control point for ensuring intersite $\delta^{18}O$ correlations.

Results and Discussion

Coarse Fraction Records From the Equatorial Indian Ocean and Carbonate Dissolution Fluctuations

Except for a few ash layers encountered at site 758 [*Farrell and Janecek*, 1991], carbonate contents range roughly from about 40 to 80% in cores MD900949 and MD900963 and in sites 722 and 758 (Figure 3). Core MD900938 shows higher carbonate contents, ranging from about 80 to 95%. Sediments mainly consist of calcareous nannofossil oozes with varying amounts of foraminifera. Sediments retrieved at site 722 and core MD900963, however, are not "typical" open sea pelagic carbonates. Site 722 was cored in an area strongly affected by monsoonal winds and is close to the Arabian coast. Consequently, carbonate contents reflect the dilution by terrigeneous material, largely eolian in origin, coming from nearby arid land masses [*Murray and Prell*, 1991]. Core MD900963 was retrieved on the eastern shoulder of the Maldives Ridge. Besides calcareous planktonic particles, the sediments at this location also contain bank-derived aragonite needles (aragonite content averages 8% with a standard deviation of 10% (G. Haddad, personal communication, 1992)) which are typical of periplatform carbonate deposits [e.g., *Kier and Pilkey*, 1971; *Schlager and James*, 1978; *Droxler et al.*, 1983, 1990; *Boardman and Neumann*, 1984].

Coarse fraction records from all sites are plotted versus age for the last 2500 kyr (Figure 4). Ages are those derived from biostratigraphic and magnetostratigraphic age models (Table 2). In sediments older than about 1500 ka, grain size varied with only low amplitudes. The fluctuations do not appear to be easily correlatable among all the sites (except between cores MD900938 and MD900940, which were retrieved about 100 km apart on Madingley Rise (Figure 1 and Table 1)). At about 1300-1500 ka, coarse fraction records show the onset of a high-amplitude, long-term oscillation. Minima and maxima of this oscillation appear to be roughly correlatable across the entire tropical Indian Ocean. Minima are centered at about 300-500 ka and 900-1000 ka, and maxima are centered at about 100-200 ka, 700-800 ka, and 1100-1300 ka.

The coarse fraction from these sediments consists almost exclusively of sand-sized foraminifera (radiolarians are usually < 10%). Past changes in the coarse fraction abundance in these sediments may reflect (1) ecological factors, such as past changes in the mean size of foraminifera or changes in the foraminifera to nannofossils ratio [*Briggs et al.*, 1985; *Mienert and Bloemendal*, 1989], or (2) sedimentological factors, such as carbonate dissolution [*Berger et al.*, 1982; *Peterson and Prell*, 1985b], changes in the winnowing intensity at the seafloor which preferentially removes fines and leaves coarse particles [*Wu and Berger*, 1991; *Mayer et al.*, 1993], or changes in dilution by fine-grained, nonbiogenic particles.

A positive relationship between coarse fraction content and foraminiferal preservation has been previously observed at site 758 [*Chen and Farrell*, 1991], thus clearly arguing for a control by carbonate dissolution. We examined the fragmentation of planktonic foraminifera in the 250- to 355-μm size fraction in cores MD900938, MD900949, and MD900963. In core MD900938, fragmentation was also measured in the >150 μm fraction. The strong correlation between our preservation

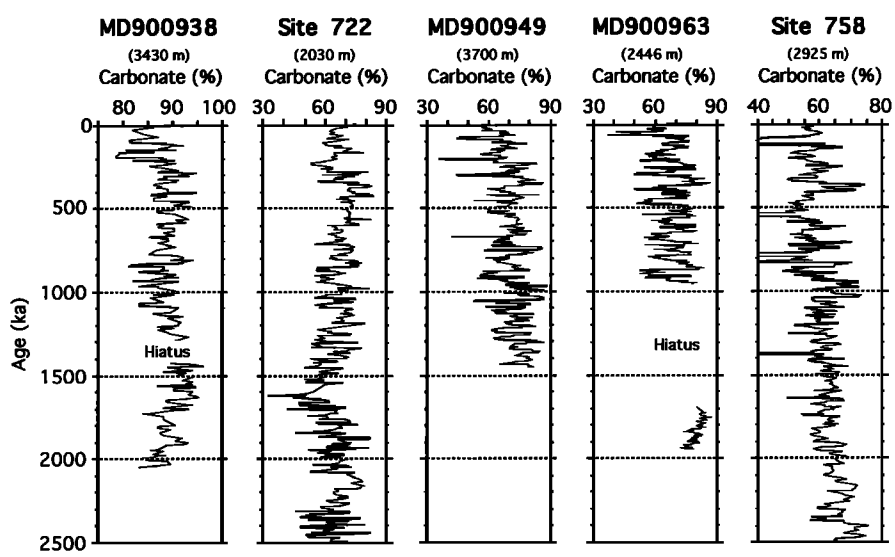

Figure 3. Plots of carbonate contents versus age for cores MD900938 and MD900940 (Madingley Rise, western equatorial Indian Ocean), cores MD900949 and MD900963 (Maldives area, central equatorial Indian Ocean), and sites 722 (Owen Ridge, Arabian Sea) and 758 (Ninetyeast Ridge, eastern equatorial Indian Ocean). Carbonate contents data from sites 722 and 758 are from *Murray and Prell* [1991] and *Farrell and Janecek* [1991], respectively.

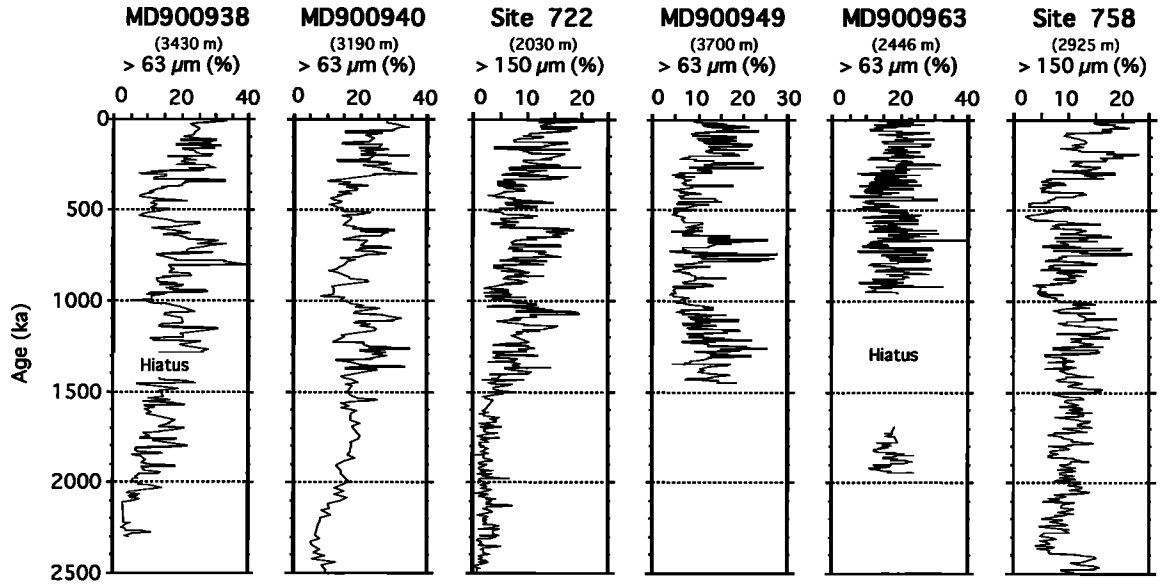

Figure 4. Plots of coarse fraction contents versus age for cores MD900938 and MD900940 (Madingley Rise, western equatorial Indian Ocean), cores MD900949 and MD900963 (Maldives area, central equatorial Indian Ocean), and sites 722 (Owen Ridge, Arabian Sea) and 758 (Ninetyeast Ridge, eastern equatorial Indian Ocean). Coarse fraction data from sites 722 and 758 are from *Murray and Prell* [1991] and *Farrell and Janecek* [1991], respectively.

index (measured in the 250- to 355-μm fraction) and the index measured in the >150-μm fraction (Figure 5) indicates that our preservation index is a valuable tool for estimating carbonate preservation/dissolution despite the smaller size range of the 250- to 355-μm fraction and the relatively low number of fragments it contains compared to the >150-μm fraction.

A good relationship can be seen in Figures 6 and 7 between coarse fraction content and foraminiferal preservation in cores MD900938, MD900949, and MD900963 (despite the fact that the sampling resolution of the foraminiferal fragmentation analyses in cores MD900938 and MD900949 is lower than the sampling resolution of the grain size analyses). This

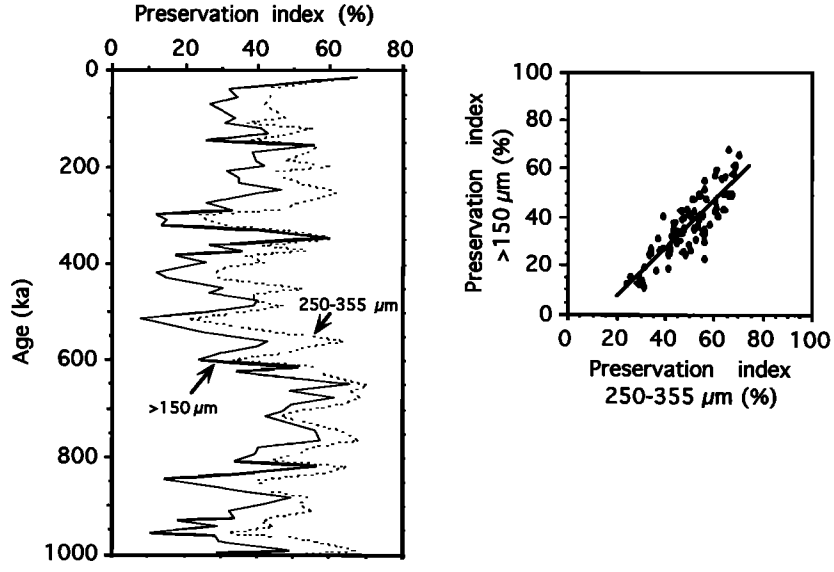

Figure 5. (left) Plots of carbonate preservation indices versus age for core MD900938 over the last 1000 kyr. Whole foraminifera/whole foraminifera plus foraminiferal fragments, in percentages, measured in the 250- to 355-μm size fraction (dashed line) and the > 150-μm size fraction (solid line). (right) Plot of the >150-μm preservation index (PI) versus the 250- to 355-μm index. In the samples studied there appears to be a strong linear correlation between the two indices. $PI_{150} = (1.01 x PI_{250\text{-}355}) - 13.7$ (R^2=0.74).

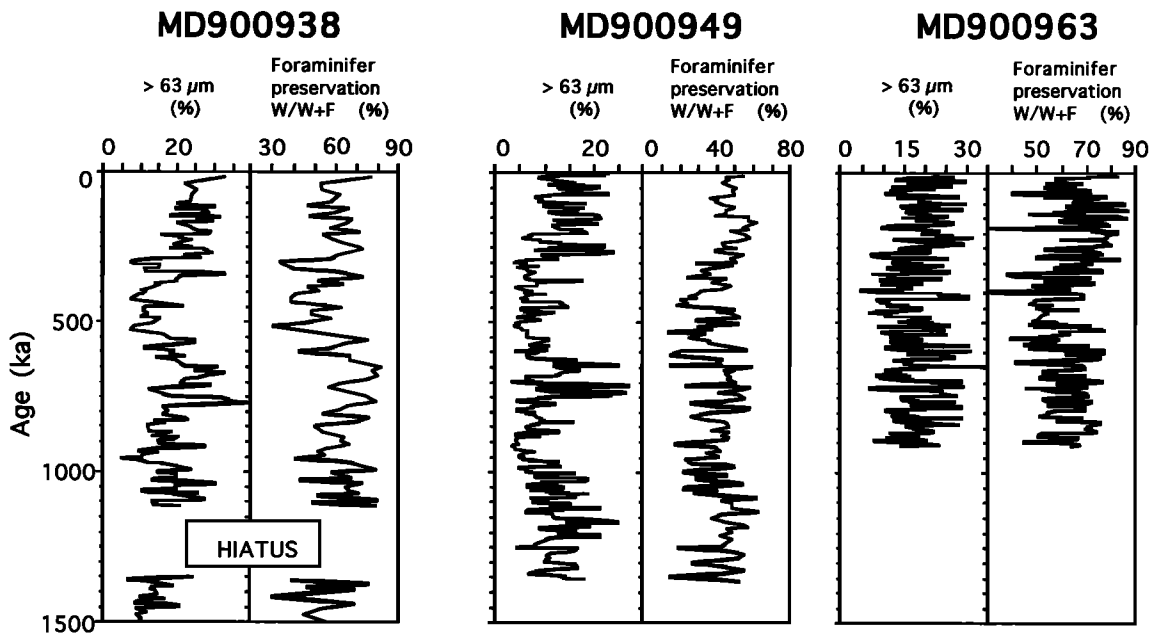

Figure 6. Coarse fraction content (> 63 μm) and foraminiferal preservation index (calculated as the percentage of whole foraminifera (W) to fragments plus whole foraminifera (W+F)) for SEYMAMA cores MD900938, MD900949, and MD900963. The good visual correlation between the two curves suggests that most of the changes in grain size, and especially the long-term oscillations observed in Figure 4, are controlled by carbonate dissolution.

relationship is particularly clear in the two shallowest cores MD900938 (3430 m) and MD900963 (2446 m). Because dissolution becomes a more dominant sedimentary process as water depth increases, we expected to find the strongest relationship between foraminiferal fragmentation and coarse fraction content in core MD900949, which is the deepest SEYMAMA section studied (3700 m). Surprisingly, the correlation is less obvious in this core, although it is still noticeable. The following scenario could well explain such results: In the early stages of dissolution (little loss of carbonate and shallow water depths), foraminiferal fragmentation results in a net transfer of material from the coarse to the fine fractions, whereas there is only a little loss of the finest particles through total dissolution. Thus grain size changes and foraminiferal fragmentation are strongly related. With increasing water depth, as loss of carbonate becomes more important, grain size distribution results from two processes that act in opposite directions: foraminiferal fragmentation tends to enrich the finest size fractions relative to the coarser ones, whereas at the same time, increased dissolution of the finest particles tends to reduce significantly the relative contribution of the fine fraction to the total sediment. This significant loss of the finest fraction may well explain the loose relationship between foraminiferal fragmentation and the coarse fraction content at great depths.

The intersite comparison between the different coarse fraction curves (Figure 4) clearly shows that the same long-term signal is recorded whatever the coarse size fraction (> 63 μm in SEYMAMA cores and >150 μm in ODP sites). In core MD900938, we studied the relative weight contributions of the >63-μm, >250-μm, and >355-μm fractions to the total sediment. Our data confirm that changes in grain size composition mainly result from variations in the relative importance of the finest and coarsest end-members (<63 μm and >355 μm), whereas the relative abundances of intermediate fractions (63-250 μm and 250-355 μm) show little change (Figure 8). Thus grain size changes controlled by carbonate dissolution imply a net transfer of foraminiferal fragments from the coarsest fractions to the finest ones. In the intermediate fractions, as suggested by *Berger et al.* [1982], weight losses due to the transfer of small fragments to the finer fractions were almost balanced by gains due to large fragments coming from the coarser fractions. Thus when dissolution increases, the number of fragments in intermediate fractions increases, although the weight contribution of these fractions to the total sediment does not change significantly.

Records from core MD900938, and more particularly from core MD900963, show that not only the long-term oscillation but also the higher-frequency fluctuations in coarse fraction content are mimicked by the preservation index (Figures 6 and 7). The high-resolution age models developed at sites 722 and 758 allow us to test if the higher-frequency fluctuations in coarse fraction records are correlatable among these sites.

Over the last 1000 kyr, the two coarse fraction curves bear a strong resemblance when plotted versus age (Figure 9 (middle)). Not only the large-scale fluctuations but also most of the smaller-scale changes (occurring at glacial-interglacial and higher frequencies) are correlatable among these two remote sites. Before about 1000 ka, however, correlations are usually poorer, with intervals where the grain size signals appear almost anticorrelated (see below for discussion). We have no detailed foraminiferal fragmentation records for sites

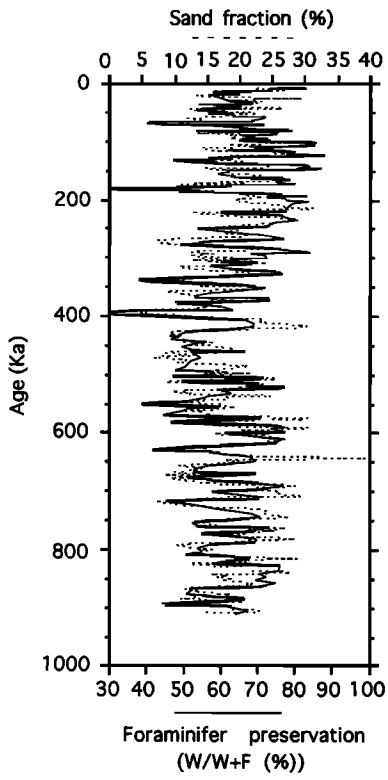

Figure 7. Superimposition of coarse fraction content (> 63 μm) and preservation index measured in core MD900963 shows that not only the long-term oscillation (see also Figures 4 and 6) but also the high-frequency changes in the coarse fraction record are controlled by changes in carbonate dissolution/preservation.

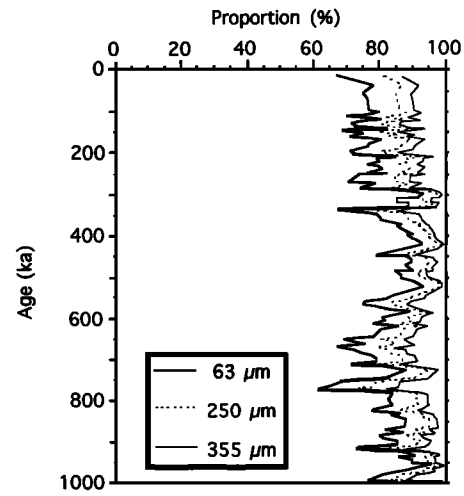

Figure 8. Abundance distribution of size classes as a function of age in core MD900938. We measured the weight percent of the >63-μm fraction (thick solid line), we subsieved this sand fraction to measure the weight percent of the > 250-μm fraction (dashed line), and finally we measured the weight contribution of the > 355-μm fraction by subsieving the > 250-μm fraction (thin solid line).

722 and 758. Nevertheless, a positive relationship has been observed between coarse fraction content and foraminiferal preservation at site 758 [*Chen and Farrell*, 1991] and in piston core RC27-61 retrieved in the vicinity of site 722 (*Murray and Prell*, 1992). Furthermore, sites 722 and 758 lie in very different environments, some 3700 km apart (Figure 1 and Table 1); thus it is unlikely that changes in surface productivity, dilution by nonbiogenic material, or winnowing at the seafloor could explain such an impressive, long-distance correlation of coarse fraction records over the last 1000 kyr. As concluded for the long-term oscillation, these regionally correlatable "high-frequency" changes in grain size are most probably controlled by carbonate dissolution pulses. In the late Pleistocene, the coarse fraction changes are consistent with the "classical" Indo-Pacific dissolution pattern [*Berger*, 1973; *Volat et al.*, 1980; *Vincent*, 1985], showing increased carbonate dissolution (coarse fraction minima) during interglacial intervals and enhanced carbonate preservation (coarse fraction maxima) during glacial intervals (Figure 9). However, the dissolution/preservation pattern is not a simple glacial-interglacial dichotomy. In many intervals, the coarse fraction cycles appear slightly shifted with respect to the glacial-interglacial cycles, in agreement with observations made in deep-sea carbonate records [e.g., *Luz and Shackleton*, 1975; *Moore et al.*, 1977; *Peterson and Prell*, 1985a; *Farrell and Prell*, 1989].

As noted above, although the correlation between coarse fraction records of sites 722 and 758 is fairly good, there are noticeable differences in these records. These discrepancies are especially noteworthy in the early Pleistocene. Between about 1100 ka and 1250 ka, for instance, the two grain size curves appear almost completely anticorrelated (Figure 9, (middle)). Prior to about 900 ka, the lack of large-amplitude 100-kyr glacial-interglacial oscillations in carbonate dissolution (which dominate the grain size records of sites 722 and 758 in the late Pleistocene) may explain why small-amplitude, local grain size changes are able to perturb more efficiently the regional (global) signal. Also, in the early Pleistocene, we cannot rule out the possibility that discrepancies between grain size records may result from inaccuracies in our age models, which rest upon the intersite correlation of low-amplitude 41-kyr $\delta^{18}O$ cycles. In the late Pleistocene, however, there is little doubt concerning the accuracy of the $\delta^{18}O$ stratigraphy. Consequently, minor discrepancies between the coarse fraction records from sites 722 and 758 can be confidently ascribed to the superimposition of local signals (e.g., winnowing and productivity) on the regionally correlatable dissolution signal.

Unlike grain size records, carbonate contents of sites 722 and 758 cannot be readily correlated (Figure 3). Variations in carbonate content at site 758 (2925-m water depth) primarily reflect changes in surface productivity and terrigeneous input [*Farrell and Janecek*, 1991], whereas those at site 722 mainly reflect dilution by terrigeneous material and do not reflect dissolution; for example, low $CaCO_3$ content observed in glacial intervals is associated with a good preservation of foraminiferal tests [*Murray and Prell*, 1991, 1992]. In some areas, carbonate content is a good proxy for carbonate dissolution and can be used for long-distance intersite correlations [e.g., *Hays et al.*, 1969]. However, our

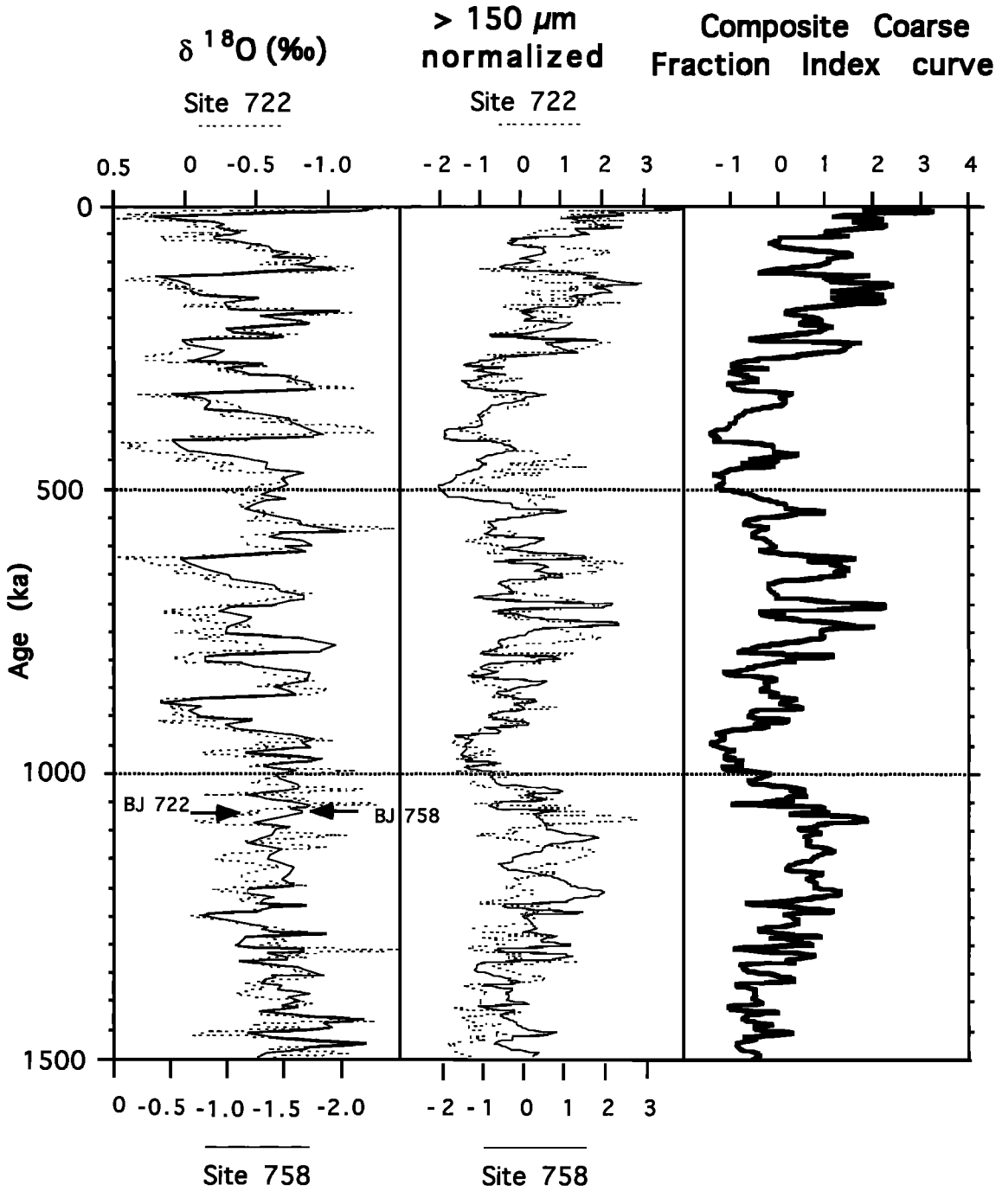

Figure 9. (left) Detailed $\delta^{18}O$ records from sites 722 and 758 [*Murray and Prell*, 1991; *Farrell and Janecek*, 1991]. These isotopic records were independently correlated to the orbitally tuned isotopic record from ODP site 677 [*Shackleton et al.*, 1990]. This age model allows a detailed comparison between coarse fraction records from both sites. (middle) Normalized coarse fraction records from sites 722 and 758. The very good match suggests that not only the long-term oscillation (see also Figures 4 and 6) but also most of the higher-frequency fluctuations are correlatable from the Owen Ridge (Arabian Sea, site 722) to the Ninetyeast Ridge (eastern equatorial Indian Ocean, site 758), some 3700 km away. (right) Composite Coarse Fraction Index curve (CCFI) obtained by stacking the normalized coarse fraction records. This method tends to diminish local signals and enhance regionally correlatable signals.

observations confirm that it is not always true. In the tropical Indian Ocean, coarse fraction content constitutes a better tool for intersite correlations and appears to be a much more sensitive indicator of dissolution than carbonate content in relatively shallow sites (above the lysocline).

The same carbonate dissolution pattern is observed during the Pleistocene in the Indian Ocean and the Pacific Ocean, with increased carbonate dissolution during interglacial intervals and enhanced carbonate preservation during glacial intervals [e.g., *Volat et al.*, 1980; *Vincent*, 1985]. Therefore, some interoceanic correlations between coarse fraction records should be possible if global signals have not been masked by local or regional processes. To test the possibility of interoceanic correlations, we selected the Deep Sea Drilling Project (DSDP) site 572 (central equatorial Pacific) in which a good quality coarse fraction record (> 150 μm) and a detailed $\delta^{18}O$ curve are available [*Farrell*, 1991]. We developed an age model at site 572 by aligning features of the $\delta^{18}O$ record with the features in the orbitally tuned $\delta^{18}O$ record from site 677 [*Shackleton et al.*, 1990]. In Figure 10, the coarse fraction records from sites 572, 722, and 758 are plotted versus age for comparison. At every site, coarse fraction contents generally increase during glacial intervals and decrease during interglacial intervals. These fluctuations, which we interpret as reflecting changes in carbonate dissolution, are in agreement with the Indo-Pacific dissolution-preservation history observed in deep-sea carbonate content records [*Berger*, 1973; *Vincent*, 1985; *Farrell and Prell*, 1989]. As was already noted for sites 722 and 758 (see above), in numerous places the coarse fraction cycles in site 572 appear slightly shifted with respect to the glacial-interglacial cycles: preservation maxima (peaks in the coarse fraction curve) are centered on the intervals of most rapid deglaciation rather than at the glacial maxima (e.g., coarse fraction peaks at about 330 and 420 ka). *Berger and Vincent* [1981] have reported such preservation spikes coinciding with deglaciation, in calcareous microfossil assemblages from piston cores of the equatorial Pacific. The preservation spike corresponding to the last deglaciation is

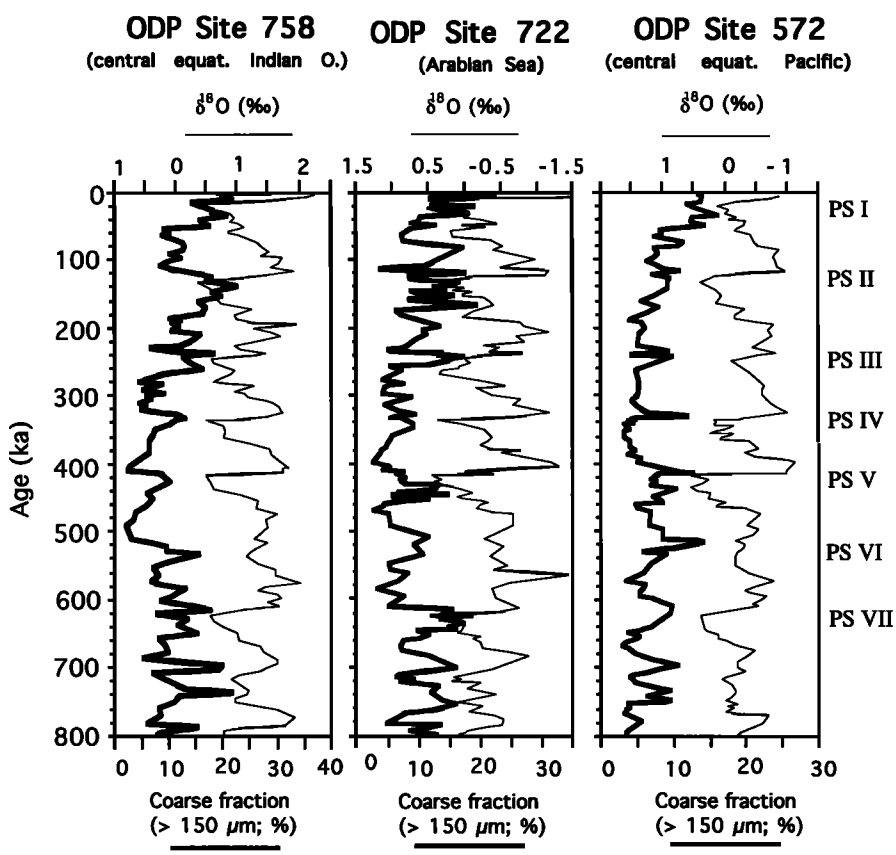

Figure 10. Coarse fraction contents and $\delta^{18}O$ values plotted versus age at ODP sites 758 and 722 (tropical Indian Ocean) and 572 (equatorial Pacific) for the last 800 kyr (data from *Farrell and Janecek* [1991], *Murray and Prell* [1991], *Farrell* [1991], respectively). At all three sites, coarse fractions generally increase during glacial intervals and decrease during interglacial intervals. We interpret this pattern in terms of carbonate dissolution (e.g., increased dissolution during interglacial reduces coarse fraction content through fragmentation of foraminifera). Preservation spikes (PS I, PS II, etc.) at right are those noted in piston cores of the equatorial Pacific by *Berger and Vincent* [1981]. Many of these preservation spikes are clearly visible here in the three coarse fraction records and, as already noted by Berger and Vincent, coincide with deglaciation events.

considered to be a global phenomenon [*Berger*, 1977], although *François et al.* [1990] recently suggested that this preservation spike is limited to the Atlantic Ocean.

Construction of a Composite Coarse Fraction Index Curve for the Tropical Indian Ocean

Coarse fraction records from sites 722 and 758 were normalized and linearly interpolated at a regular 2-kyr time interval (Figure 9 (middle)). We stacked these normalized coarse fraction curves to produce a reference coarse fraction record in which local signals tend to cancel out. The resulting Composite Coarse Fraction Index curve (CCFI) is shown in Figure 9 (right) and the values are given in Appendix Table A1[1]. (The CCFI used in this paper corresponds to the Composite Grain Size Index (CGSI) used in previously published studies [Bassinot, 1993a, b; Bassinot et al., 1992]. Although this terminology change may introduce some confusion, it is more proper to use the term "coarse fraction" (rather than "grain size") to refer to one distinct size fraction obtained by wet sieving).

Peterson and Prell [1985a] have provided a 800-kyr record of bathymetric and temporal variations of carbonate preservation in the equatorial Indian Ocean. They used a quantitatively defined Composite Dissolution Index (CDI) measured on a set of 7 piston cores retrieved from water depths of 2900 m to 4400 m on the Ninetyeast Ridge near 6°S. Core V34-53 was given special attention because of its sensitive location (3812 m) near the depth of the present-day foraminiferal and hydrographic lysoclines in the equatorial Indian Ocean (3800 m). For intercomparison between our CCFI curve and the CDI curve from core V34-53, we had to modify the initial age model developed by *Peterson and Prell* [1985a]. This age model rested upon the correlation of the $\delta^{18}O$ record from core V34-53 to the SPECMAP stack [*Imbrie et al.*, 1984]. The orbital chronology developed from site 677 introduces three additional precessional cycles in isotopic stages 17 and 18 compared to the SPECMAP solution [*Shackleton et al.*, 1990]. The new age model for core V34-53 was developed by correlating the $\delta^{18}O$ record of core V34-53 (data from *Peterson and Prell* [1985a]) to the orbitally tuned $\delta^{18}O$ record from site 677 [*Shackleton et al.*, 1990] using the program "LINAGE".

There is a generally excellent match between our CCFI curve and the CDI curve from core V34-53 [*Peterson and Prell*, 1985a] (Figure 11). Over the interval from about 350 ka to 600 ka, however, low accumulation rates (around 0.5 cm kyr^{-1}) due to intensified dissolution and loss of dissolved carbonate from the sediments tend to smooth the isotopic record from core V34-53. This smoothed part of the $\delta^{18}O$ record is difficult to correlate confidently to the $\delta^{18}O$ curve from site 677, which probably explains slight discrepancies (leads/lags) between the CCFI and the CDI records.

Despite these slight discrepancies, the generally good agreement between the CCFI curve and the CDI curve clearly indicates that the CCFI provides a reliable though qualitative record of past changes in carbonate dissolution in the tropical Indian Ocean. It also extends back in time (to 1500 ka) the record of dissolution first provided by the CDI record. Additional high-resolution coarse fraction records are necessary to test the reliability of the CCFI curve over the interval from 900 to 1500 ka. However, over this time interval, the CCFI curve seems to provide a good reference for the long-term oscillation.

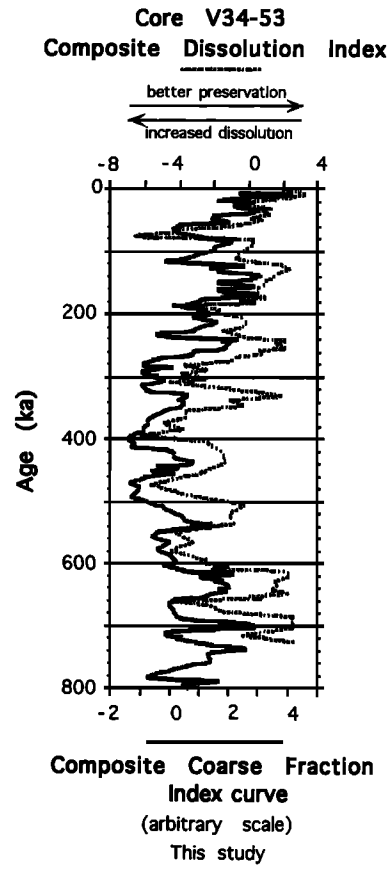

Figure 11. Comparison of the Composite Coarse Fraction Index curve (CCFI) (this study) with the Composite Dissolution Index curve from core V34-53 (Ninetyeast Ridge) [*Peterson and Prell*, 1985a). The impressive match between the two curves demonstrates unambiguously that coarse fraction changes that are correlatable over long distances in the tropical Indian Ocean are controlled by regional (or global) changes in carbonate dissolution. The CCFI curve may be used as a reference curve to develop a "coarse fraction stratigraphy" for the last 1500 kyr.

In deep-sea carbonate deposits, changes in *P* wave velocity parallel changes in coarse fraction content [e.g., *Johnson et al.*, 1977; *Hamilton et al.*, 1982]. Therefore, in the tropical Indian Ocean, *P* wave velocity profiles can be tied to a precise stratigraphic and chronostratigraphic framework by correlating them to the CCFI curve [*Bassinot*, 1993a, b; *Bassinot et al.*, 1992]. Because of the potential for near-continuous measurements of acoustic velocity on unsplit cores, "sonostratigraphy" may become a rapid means of determining the stratigraphic position of core samples and of getting some indication of paleoceanographic conditions that prevailed during sediment deposition [*Bassinot*, 1993a, b; *Bassinot et al.*, 1992].

[1]Appendix Table A1 Is available with entire article on microfiche. Order from the American Geophysical Union, 2000 Florida Avenue, N.W., Washington, D.C. 20009. Document P94-001; $2.50. Payment must accompany order.

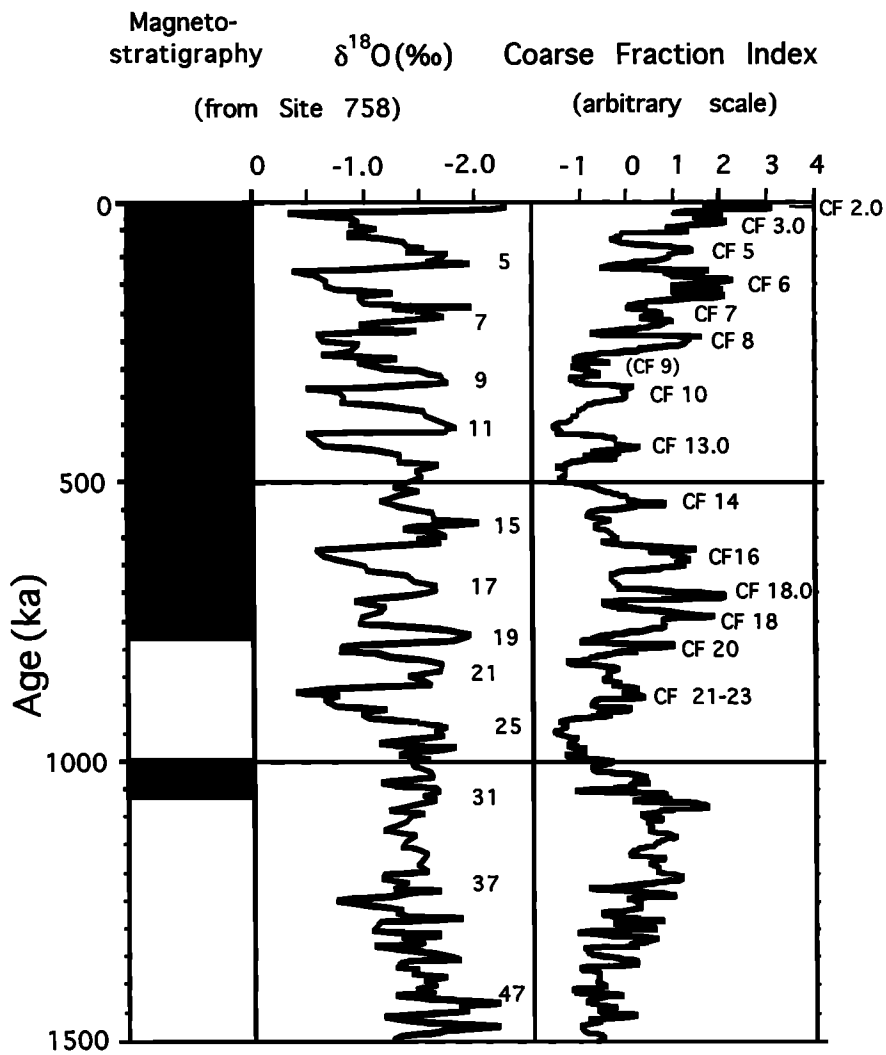

Figure 12. Composite Coarse Fraction Index curve (CCFI) tied into a precise stratigraphic framework using the magnetostratigraphy and isotope stratigraphy of site 758. Major peaks of the CCFI curve have been labeled according to their position relative to the $\delta^{18}O$ stratigraphy (see text for details).

In Figure 12, the CCFI curve is shown with the oxygen isotope stratigraphy and the magnetostratigraphy from site 758. In the 0-900 ka interval, we have assigned numbers to major CCFI peaks (i.e., preservation events) based on their respective positions relative to the $\delta^{18}O$ stratigraphy. Peaks were labeled "CF" followed by the oxygen isotopic stage number in which they reach their maximum. Higher CCFI values are generally observed in glacial intervals (e.g., peaks CF6, CF8, CF10, CF14, and CF16) or are shifted toward early stages of deglaciation (e.g., CF18.0). Nevertheless, peaks also appear in some interglacial intervals (e.g., CF5 and CF7). These latter preservation peaks, which generally have lower maximum values than peaks in adjacent glacial intervals, are bracketed by sharp dissolution events. For instance, the peak CF5, which reaches lower values than peaks CF3.0 and CF6, is bracketed by the important dissolution event that occurs at the transition between oxygen isotope stage 5 and 4 (a global event which has been recognized in the Indo-Pacific Ocean as well as in the Atlantic Ocean [*Crowley*, 1985]) and a dissolution event occurring in substage 5.5. In the CCFI record, both dissolution events have roughly the same magnitude, whereas in the CDI record, the dissolution event at the stage 4/5 transition is much more clearly expressed (Figure 11).

Spectral Analysis of the Composite Coarse Fraction Index Record

We analyzed the CCFI variability in the frequency domain over the last 900 kyr using the Blackman-Tukey approach with 50% lags [*Jenkins and Watts*, 1968]. Much of the total variance in the CCFI record (Figure 13, solid line) occurs in the low-frequency end of the spectrum and is associated with the long-term oscillation. Observations in the time domain over the last 1500 kyr suggest that the long-term oscillation has an irregular cycle length of the order of ~500 kyr (Figures 9 and 14). Minima in the CCFI curve (maxima of dissolution) are found approximately at 400 ka, 950 ka, and 1400 ka, and maxima are found at about 150 ka, 700 ka, and 1125 ka

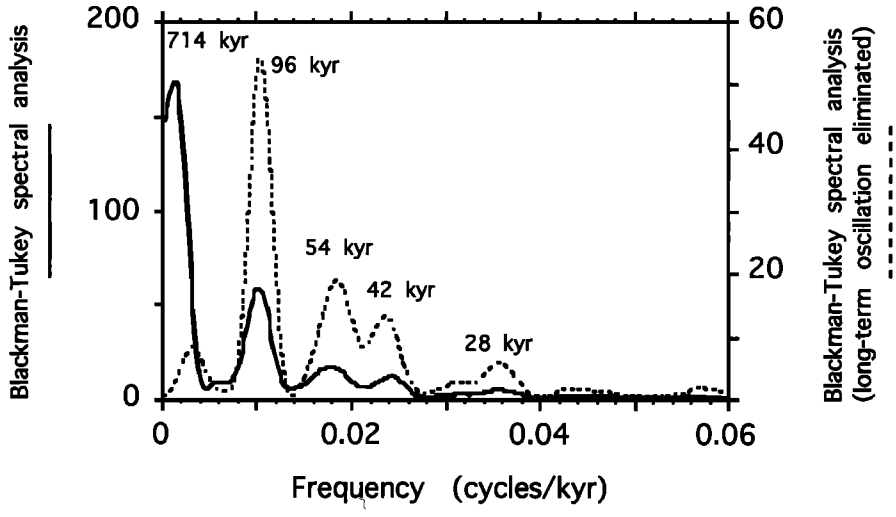

Figure 13. Blackman-Tukey spectral analysis of the Composite Coarse Fraction Index curve (CCFI) over the last 900 kyr. Dashed line indicates the power spectrum calculated after the long-term oscillation of the CCFI record has been eliminated to resolve more accurately the high-frequency cycles (see text for details).

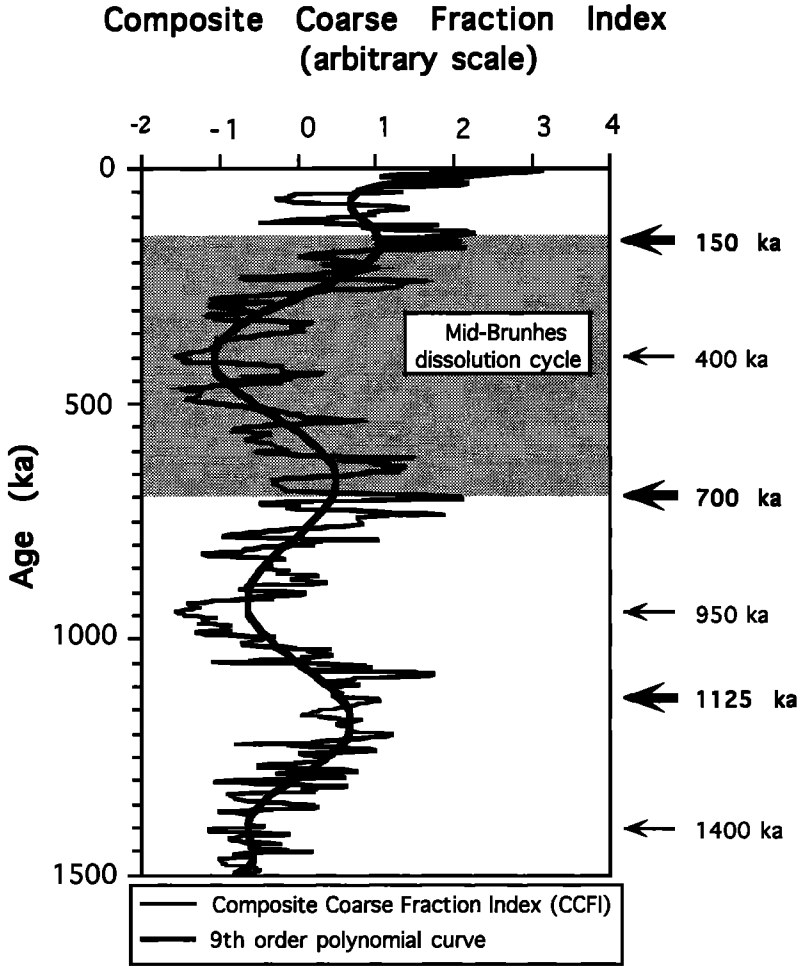

Figure 14. Composite Coarse Fraction Index curve smoothed with a 30-point window running average to enhance the long-term dissolution oscillation that has an irregular wavelength of the order of ~500 kyr. Thin and thick arrows indicate dissolution maxima and minima, respectively.

(Figures 9 and 14). Although the long-term oscillation in the CCFI record has an irregular wavelength, we will refer to it as the "~500-kyr oscillation" for convenience.

In series dominated by high-amplitude low-frequency oscillations, the high-frequency part of the power spectra is distorted by variance transfer from the low-frequency peaks. Prewhitening techniques are commonly used to reduce this distortion [*Jenkins and Watts*, 1968]. However, in the CCFI power spectrum, we noticed that prewhitening not only reduced the amplitude of the ~500-kyr cycle but also significantly reduced the power estimate of the ~100-kyr oscillation. In order to define better the high-frequency oscillations without disturbing the amplitude ratios between the primary Milankovitch bands, we decided to eliminate long-term oscillations by fitting a polynomial curve to the CCFI record and then subtracting this curve from the bulk record. An order of 9 polynomial curve satisfactorily described the long-term oscillation (Figure 14).

In the resulting detrended CCFI record, variance clearly occurs in the eccentricity (96 kyr) and obliquity (42 kyr) frequency bands, but the absence of discrete spectral peaks at 23 and 19 kyr (Figure 13, dashed line) suggests that the CCFI curve does not record a precessional response. The lack of a precession signal in the CCFI record may be due either to the attenuation of the high-frequency signals during the stacking procedure or to the true absence of precession signals in the records of coarse fraction variations.

Spectral Analyses of the Coarse Fraction Records From Sites 722 and 758

We eliminated the long-term oscillation from the individual coarse fraction records of sites 722 and 758 using the same technique used for the CCFI record (see above), and we performed Blackman-Tukey spectral analyses (Figure 15). Strong 98-kyr cycles in the coarse fraction records of sites 722 and 758 reflect fluctuations associated with late Pleistocene glacial-interglacial changes, as evidenced by the comparison of $\delta^{18}O$ and coarse fraction records in the time domain (Figure 9). Variance is also clearly found in the obliquity band (41 kyr) at site 758 but is weaker at site 722. Power spectra of coarse fraction records also display frequencies that are not associated with the primary Milankovitch bands (Figure 15) (27, 32, and 60 kyr for site 722 and ~30 and 53 kyr for site 758). These periods are close to those of 29, 35, and 54 kyr found in low-latitude deep-sea records of past climatic changes (e.g., eolian grain size records in the Pacific, sea surface temperature (SST) in the southern tropical Indian Ocean, and indices of monsoon strength [*Pisias and Rea*, 1988; *Clemens and Prell*, 1991a, b; *Murray and Prell*, 1991]). These periods may be related to a nonlinear response of the low-latitude climatic system to insolation forcing at the primary Milankovitch bands [*Clemens and Prell*, 1991a, b].

The $\delta^{18}O$ and coarse fraction records at site 758 show a small peak at a period of about 23 kyr (Figure 15). At site 722, no precession-related peaks were observed in the power spectrum of the coarse fraction record although a 23-kyr cycle is observed in the power spectrum of the $\delta^{18}O$ record, indicating that the sequence probably permits the recovery of a precession signal when it exists (Figure 15). Thus, given these data, and as concluded earlier by *Peterson and Prell* [1985a], it seems that there are no clear regional precession-related oscillations in the carbonate dissolution pattern from the tropical Indian Ocean. Based on the weak linkage of the CDI curve and the $\delta^{18}O$ record from core V34-53 over the precessional frequencies, and the high coherencies at the 100-kyr^{-1} and 41-kyr^{-1} frequencies, *Peterson and Prell* [1985a] suggested that deepwater circulation changes, driven by high-latitude climatic forcing in the north Atlantic, were the principal source of variability in the preservation record of the equatorial Indian Ocean.

However, additional high-resolution data on Ninetyeast Ridge and Owen Ridge are necessary to check our results. Based on the sedimentation rates and sampling intervals, the lowest sampling resolution encountered should be about 10 kyr for both sites 722 and 758 (mean sampling intervals over the last 900 kyr are 5.5 and 6.5 kyr for sites 722 and 758, respectively). Theoretically, a sampling interval of 10 kyr is just high enough to resolve a 20-kyr cycle (the spectral estimates covering a frequency range up to the Nyquist frequency of 0.05 cycles kyr^{-1}). However, relatively low sedimentation rates at sites 722 and 758 (i.e., at site 758, sedimentation rates range roughly from 1.2 to 2.5 cm kyr^{-1}, with a mean value of 1.8 cm kyr^{-1}) most probably result in a significant reduction in the variance at high frequencies, as predicted by bioturbation models [i.e., *Pisias*, 1983].

High-Resolution 900-kyr Record From Core MD900963 (Maldives Area): Dissolution Changes Associated With the Decay of Organic Carbon

Core MD900963 offers the best opportunity to look at high-frequency signals. Although core MD900963 does not extend into the lower Pleistocene, the youngest part of the long-term dissolution cycle ("mid-Brunhes dissolution event") is clearly observed in the coarse fraction and the foraminiferal preservation records (e.g., Figures 4, 6, and 7), with maximum dissolution centered at about 400 ka. On shorter scales, however, the record from core MD900963 differs markedly from what we observed at sites 722 and 758. Although many dissolution-preservation events can be correlated between the CCFI curve and the coarse fraction record from core MD900963 (arrows on Figure 16), the coarse fraction curve from core MD900963 appears to be clearly dominated by higher-frequency changes than coarse fraction records from sites 722 and 758. The strong correlation between the coarse fraction record and the preservation index curve in core MD900963 (Figure 7) again indicates that these high-frequency grain size changes are also primarily controlled by carbonate dissolution. Spectral analysis of the foraminiferal preservation record shows that beside the large amount of variance associated with the long-term oscillation, much of the variability occurs at the periods associated with precession (23 and 19 kyr) and at 29 kyr (Figure 17). These periodicities of 23-19 and 29 kyr, which are also clearly expressed in the $\delta^{18}O$ record of this core (Figure 17), have been shown to be associated with climatic variability at low latitudes and, especially, monsoon variability [e.g., *Clemens and Prell*, 1991a, b]. Little concentration of variance in the preservation record of core MD900963 is associated with the Milankovitch eccentricity and obliquity periods. As was already noted for

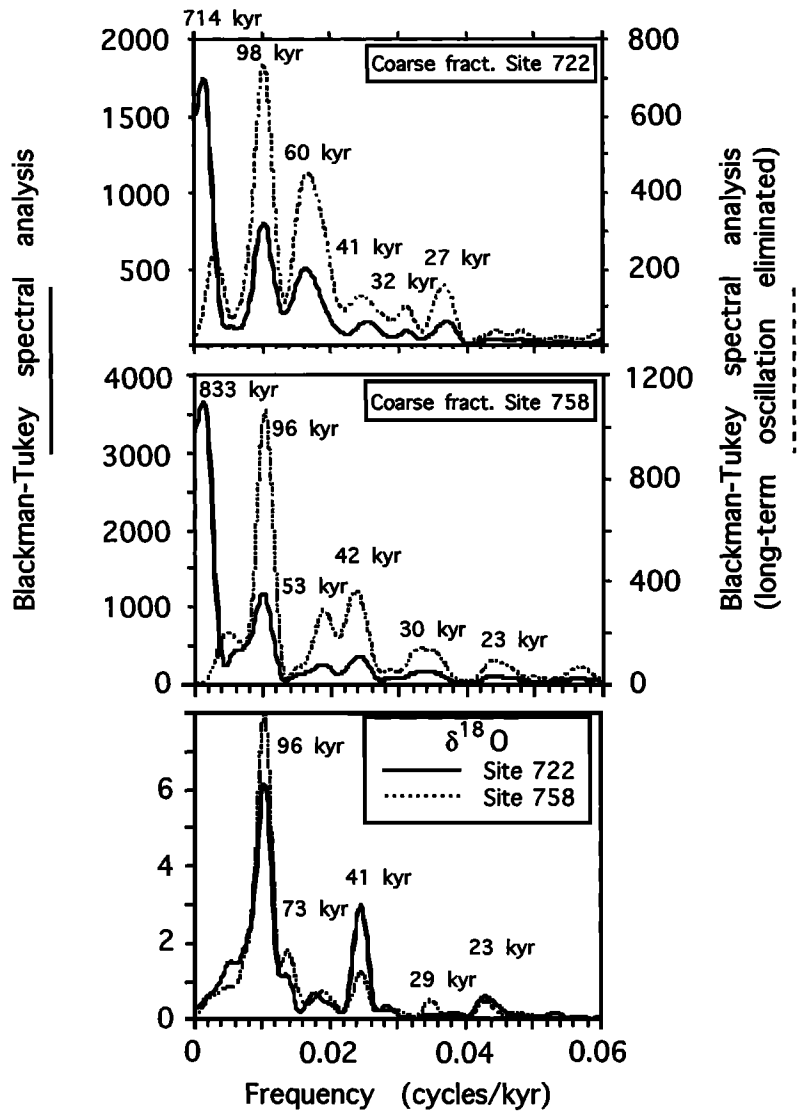

Figure 15. Blackman-Tukey spectral analysis of 900-kyr $\delta^{18}O$ and coarse fraction records from sites 722 and 758. Dashed lines in the coarse fraction spectral analyses indicate the power spectra calculated after the long-term oscillation has been eliminated to resolve more accurately the high-frequency cycles (see text for details).

sites 722 and 758, dissolution has only a subordinate effect on carbonate content in core MD900963. Hence the power spectrum of the carbonate content record differs clearly from that of the preservation index, with the highest spectral density found in the eccentricity band (95 kyr) (Figure 17). This suggests that high-frequency cycles in the preservation and the coarse fraction records from core MD900963 occur with little total loss of $CaCO_3$ to dissolution.

Organic carbon (C_{org}) measurements in the top 10 m of core MD900963 (last 170 kyr) show that the high-frequency dissolution pulses observed in the foraminiferal fragmentation (or in grain size) are associated with changes in C_{org} content (e.g., high dissolution is observed in levels with high C_{org} content) (Figure 18). Dissolution maxima and high organic carbon contents are associated with even-numbered $\delta^{18}O$ stages and a particularly well-marked maximum is shown in glacial stage 6 (Figure 18). The occurrence of a dissolution maximum in glacial stages is opposite to what has been observed at sites 722 and 758.

Strong evidence supports the hypothesis that fluctuations in organic carbon content in the sediments reflect changes in paleoproductivity [e.g., *Müller and Suess*, 1979; *Bertrand and Lallier-Vergès*, 1993]. However, only a small amount of the C_{org} rain to the seafloor is ultimately preserved in the sediments (about 0.1 to 2% in sediments accumulating at 2 to 13 cm kyr^{-1} [*Müller and Suess*, 1979]). Thus we suggest that high-frequency dissolution pulses observed in core MD900963 may be controlled ultimately by the oxidation of organic matter incorporated into the sediments in greater quantities during periods of high productivity [*Emerson and Bender*, 1981]. Sedimentation rates in core MD900963 are higher during odd-numbered $\delta^{18}O$ stages than during even-numbered stages, thus better preservation of C_{org} through rapid burial cannot be invoked to explain higher C_{org} contents. Further

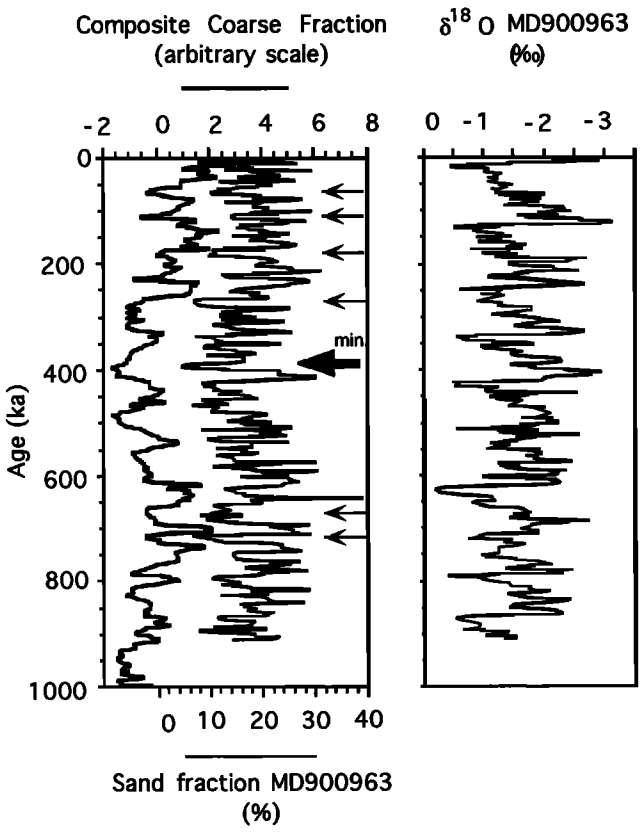

Figure 16. (left) Plots versus age of the Composite Coarse Fraction Index curve (CCFI) and the high-resolution coarse fraction record from core MD900963 over the last 900 kyr. (right) Plot of $\delta^{18}O$ versus age for core MD900963.

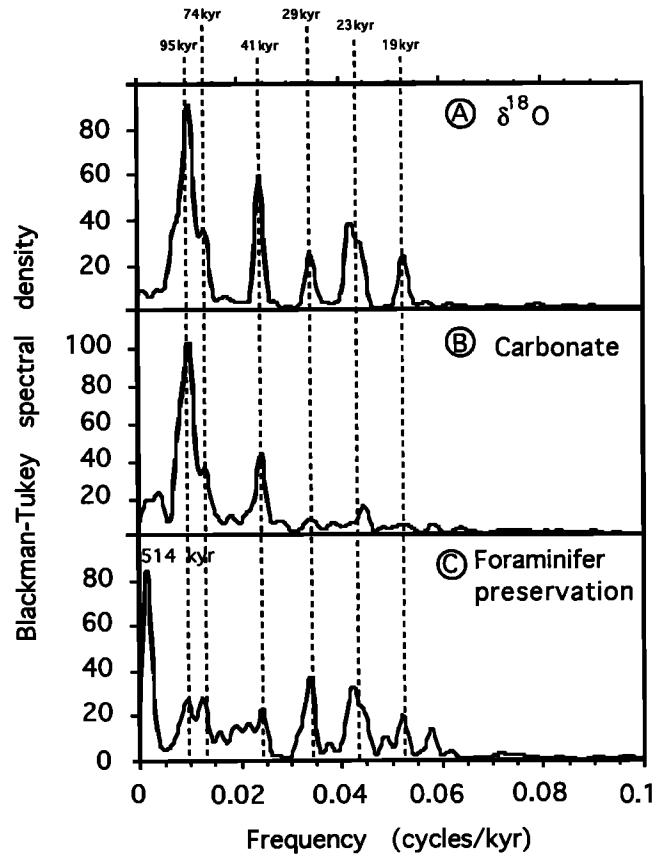

Figure 17. Blackman-Tukey spectral density of $\delta^{18}O$, carbonate content, and foraminiferal preservation records in core MD900963.

work is necessary to test this hypothesis and especially to quantify the amount of carbonate dissolved by such a process. Nevertheless, we contend that the amount required is probably small, since changes in dissolution, which are clearly reflected in foraminiferal fragmentation and grain size, do not noticeably affect the carbonate content of core MD900963 (see above).

Additional work is required to understand what ultimately controls those changes in paleoproductivity. At present, the C_{org} content record of core MD900963 is not long enough (last 170 kyr) to permit a precise spectral analysis. However, the strong precession-related cycles and the 29-kyr cycle we found in the carbonate dissolution record (foraminiferal fragmentation) argue for a low-latitude process which is most probably related to changes in the monsoonal system [*Murray and Prell*, 1992]. Using a combination of oxygen isotopes and alkenone-derived temperatures measured in the uppermost 10 m of core MD900963, *Rostek et al.* [1993] reconstructed the sea surface salinity signal for the Maldives area over the last 170 kyr. Today, the precipitation over India and the Indian Ocean is related to the strength of the SW monsoon [*Cadet and Reverdin*, 1981]. Hence *Rostek et al.* [1993] interpreted the variations in salinity recorded in core MD900963 as linked to past changes in the Indian monsoon. They found low salinity during stage 5e and the early Holocene, thus suggesting that the wet SW monsoon was stronger during these time intervals. Higher salinity were observed during glacial stages 6 and 2 and substage 5b, indicating reduced SW monsoon and/or increased dry SE monsoon.

Based on results from *Rostek et al.* [1993], it appears therefore that productivity generally increased (high C_{org} content) during periods of weak SW monsoon and/or strengthened NE monsoon. Whether the changes in productivity are regional or local is still under debate. Studies of primary productivity conducted in the modern Indian Ocean [e.g., *Krey*, 1973] indicate that south of India, the C_{org} fluxes are highest during the SW monsoon and seem to be related to an upwelling cell which is clearly observed in SST records [*Wyrtki*, 1971]. *Rao and Jayaraman* [1966], however, showed that local upwelling cells may also develop during the NE monsoon season in the vicinity of the Maldives area. Thus if productivity changes are local in extent, we suggest that increased productivity during even-numbered $\delta^{18}O$ stages may be related to the intensification of upwelling cells driven by NE monsoon winds. This hypothesis is in good agreement with the results of *Fontugne and Duplessy* [1986], who have shown that productivity during the last glacial period was greater in areas where production depends on the NE monsoon (such as upwelling off the NE coast of India).

It is puzzling that such a high-frequency dissolution signal related to C_{org} decay is not observed at site 722. The C_{org} content of site 722 [*Murray and Prell*, 1991] has a higher mean

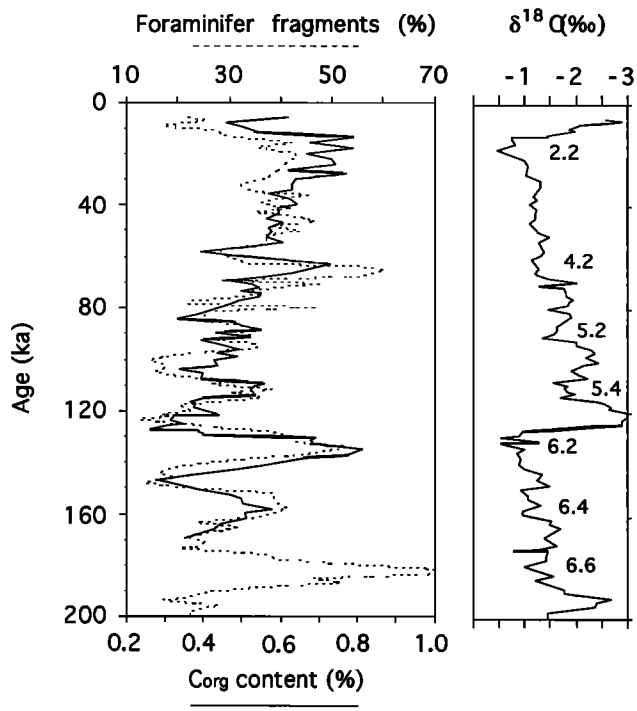

Figure 18. Plots of age versus C_{org} content and percent of foraminiferal fragments (left) and versus $\delta^{18}O$ (right) over the last 200 kyr in core MD900963.

value and shows higher-amplitude variability than that from core MD900963 (mean value is ~1% with a standard deviation of 0.43% for site 722 and 0.5% with a standard deviation of 0.12% for core MD900963). Furthermore, site 722 is situated in an area where primary productivity shows important seasonal changes which result from fluctuations of the monsoonal wind-driven Arabian upwelling. Thus we might expect to find precession-related changes in the C_{org} supply and the dissolution record from site 722. Nevertheless, the precession-related dissolution signal which is clearly observed in the grain size record of core MD900963 is absent in the grain size record from site 722. Both sections were sampled at roughly 10-cm intervals. The lower sedimentation rates at site 722 result in lower temporal resolution which most probably degrades the high-frequency signals. However, as discussed above, the spectral analysis of the oxygen record of site 722 shows a 23-kyr peak, thus indicating that the sedimentary section seems to permit the recovery of precession-related signals when they exist. Additional studies are necessary to know if the high-frequency dissolution signal observed in core MD900063 as well as the good correlation between coarse fraction content, foraminiferal fragmentation, and C_{org} content are observed at other locations.

The Long-Term Dissolution Oscillation

The major characteristic of coarse fraction records over the last 1500 kyr is the high-amplitude, ~500-kyr oscillation which reflects long-term changes in carbonate dissolution (Figures 4 and 9). This ~500-kyr oscillation is superimposed on a gradual increase in mean coarse fraction content which is observed over the last 2500 kyr in cores MD900938 and MD900940, at site 758, and especially at site 722 (Figure 4). *Farrell and Prell* [1991], who observed such a long-term increase of carbonate preservation in pelagic sediments from the Pacific, proposed that it results from a gradual lowering of sea level which fractionates carbonate from the world's shelves to the basins [e.g., *Berger*, 1970; *Berger and Winterer*, 1974].

A late Pleistocene ~500-kyr dissolution oscillation has been previously observed by *Vincent* [1981] and *Gardner* [1982] in carbonates from the deep Pacific and was recently found in aragonite-bearing periplatform deposits from the Bahamas [*Droxler et al.*, 1988] and the Maldives [*Droxler et al.*, 1990]. The last 700 kyr of this long-term oscillation correspond to the well-known "mid-Brunhes dissolution cycle" [*Adelseck*, 1977], with its characteristic dissolution maximum (grain size minimum) centered at about 400 ka and bounded by preservation maxima (grain size maxima) at about 100 ka and 700 ka. This cycle has been recognized in the Pacific [*Vincent*, 1981, 1985; *Farrell and Prell*, 1989], the north Atlantic [*Crowley*, 1985], and the equatorial Indian Ocean [*Peterson and Prell*, 1985a], clearly indicating its global character. Mid-Brunhes climatic cycles have been also observed in numerous other marine and continental records [*Jansen et al.*, 1986]. Data from *Vincent* [1981] and *Droxler et al.* [1990] suggest that the mid-Brunhes interval may be part of a long-term oscillation and not a steplike phenomenon as proposed recently by different authors [*Jansen et al.*, 1986; *Chuey et al.*, 1987; *Pisias and Rea*, 1988].

Droxler et al. [1990] indicate that a ~500-kyr carbonate dissolution supercycle is recorded for at least the past 2000 kyr in aragonite bearing periplatform sediments. However, our coarse fraction data suggest that high-amplitude oscillations that are correlatable across the tropical Indian Ocean apparently initiated at about 1400 ka. Prior to that time, only the site 758 record presents evidence for a long-term oscillation (Figure 4), but minima and maxima of this oscillation do not correlate well with minima and maxima of the aragonite supercycles described by *Droxler et al.* [1990]. For example, a maximum in the site 758 grain size record at about 2.0 Ma corresponds to a strong minimum in aragonite supercycles, and a minimum in grain size at about 1.75 Ma corresponds to a maximum in aragonite supercycles (it is interesting to note, however, that fine aragonite content of ODP hole 633A in the Bahamas shows a strong minimum at about 1.75 Ma [*Droxler et al.*, 1990]).

In the 0-1500 ka interval, the timing of minima and maxima in the aragonite supercycles [*Droxler et al.*, 1990] appears to correlate quite nicely with that of minima and maxima in the CCFI curve (Table 3). The CCFI curve, whose ages rest upon correlations to the recent age model developed from site 677 [*Shackleton et al.*, 1990], permits us to determine more accurately the ages of these minima and maxima. The long-term dissolution oscillation does not have a regular 500-kyr periodicity. Time intervals between two adjacent peaks or troughs vary from 425 to 550 kyr (Figure 14 and Table 3). We do not know yet if the long-term dissolution oscillation is a slightly irregular cycle, if it results from the distortion of a long-term supercycle by discrete events, or if it is the sum of independent events.

The high-amplitude dissolution oscillation implies important reorganizations of the oceanic carbon-carbonate

Table 3. Ages of Minima and Maxima in the Long-Term Carbonate Dissolution Oscillation as Proposed by *Droxler et al.* [1990] and in This Study

Maxima Carbonate Preservation Ma		Minima Carbonate Preservation Ma	
Droxler et al. [1990]	This Study	*Droxler et al.* [1990]	This Study
0.15	0.15	0.40	0.40
0.65	0.70	0.90	0.95
1.20	1.12	1.45	1.40

system in the early Pleistocene. At present, the mechanism involved in this large oscillation remains poorly understood. The onset of the long-term dissolution oscillation at about 1300-1500 ka corresponds roughly to the onset of large-amplitude oscillations of North Atlantic Deep Water (NADW) production, with important reductions of NADW during glacial intervals [*Raymo et al.*, 1990]. Today, NADW accumulates CO_2 oxidized from falling organic debris on its "grand journey" to the Pacific, resulting in an important asymmetry in CO_2 distribution and water acidity between the deep waters of the Atlantic and those of the Indo-Pacific. We cannot rule out the possibility that changes in NADW production may have an effect on the global deep-sea carbonate budget. However, if NADW production played a major role in the long-term dissolution oscillation, we would also expect a strong asymmetrical effect in the dissolution pattern of the Atlantic and the Indo-Pacific, resulting from basin-basin fractionation [*Berger*, 1970; *Volat et al.*, 1980; *Crowley*, 1985]. The fact that the "mid-Brunhes dissolution interval" and the dissolution intervals centered at about 950 and 1400 ka are also observed at intermediate water depths in the Indo-Pacific ocean as well as in the Atlantic [*Droxler et al.*, 1990] suggests that there is a control by transfers between global carbon-carbonate reservoirs rather than by changes in deep circulation.

There is some evidence in sequences of marine terraces for higher sea level at about ~400 ka (stage 11) and at about 950-1070 ka (stages 25 and 31) [*Pirazzoli et al.*, 1993]. However, the magnitude of sea level changes is still poorly constrained, and it is therefore highly speculative to explain the long-term dissolution oscillation as resulting from a simple shelf-basin carbonate fractionation. Thus the origin of the long-term dissolution oscillation is still unresolved. Any mechanism proposed to explain it, however, should fulfill two conditions: (1) It must affect the entire ocean, since the "mid-Brunhes dissolution cycle" and the dissolution intervals centered at about 950 and 1400 kyr have been observed in the Indo-Pacific as well as in the Atlantic, and (2) it has to affect the entire water column, since the long-term dissolution oscillation is recognized in shallow aragonite-bearing periplatform deposits as well as in deep-sea carbonates.

Long-term increases in the Ca^{++} flux to the oceans have been recently proposed to explain the progressive increase in carbonate preservation since the Pliocene [*Raymo et al.*, 1988; *Farrell and Prell*, 1989]. If we assume a "steady state" ocean, mass balance constraints require Ca^{++} input to be offset by Ca^{++} output via oceanic $CaCO_3$ sedimentation, thus explaining the better deep-sea carbonate preservation. *Raymo et al.* [1988] proposed that rapid tectonic uplift of mountains and plateau regions during the last 5 m.y. may explain such an overall increase of the Ca^{++} flux to the ocean via increased chemical and mechanical erosion. Although the details of past changes of Ca^{++} flux to the oceans and erosion budgets are not well known, we suggest that the ~500-kyr carbonate dissolution oscillation may be associated with changes in Ca^{++} flux to the ocean. These changes would result either from long-term isostatic adjustment or tectonic activity and/or past changes in the intensity of chemical weathering. The latter hypothesis finds some support in the fact that numerous marine and continental climatic time series show that the "mid-Brunhes dissolution cycle" developed during a time of important climatic changes [*Jansen et al.*, 1986].

Conclusions

Over the last 1500 kyr, major changes in coarse fraction content of pelagic carbonate sediments can be correlated over much of the tropical Indian Ocean from 2030-m to about 3700-m water depths. These changes in grain size are mainly controlled by carbonate dissolution through the breakdown of sand-sized foraminifera and the subsequent transfer of small fragments into the finer size fraction.

We constructed a Composite Coarse Fraction Index (CCFI) curve for the last 1500 kyr by merging the normalized coarse fraction records from ODP sites 722 (Owen Ridge, Arabian Sea) and 758 (Ninetyeast Ridge, eastern equatorial Indian Ocean). A precise chronostratigraphy of this composite curve was developed by correlating the $\delta^{18}O$ records of sites 722 and 758 to the orbital chronology recently developed from ODP site 677 [*Shackleton et al.*, 1990]. The CCFI curve, in which local grain size signals (winnowing and productivity) tend to cancel out, enhances the regional (or global) signal related to changes in carbonate dissolution. The last 800 kyr of the CCFI curve appear to compare extremely well with the Composite Dissolution Index (CDI) curve from core V34-53 (Ninetyeast Ridge), which unambiguously records past variations of carbonate dissolution in the equatorial Indian Ocean [*Peterson and Prell*, 1985a].

In the late Pleistocene, the CCFI variations are mainly associated with glacial-interglacial changes. Frequency spectra show strong 100-kyr and 41-kyr periodicities but little precession-related response. As proposed by *Peterson and Prell* [1985a], the lack of a clear signal in the precessional bands suggests that the regional carbonate dissolution signal is driven by changes in deepwater circulation. However, additional higher-resolution studies are necessary at ODP sites 722 and 758, since we cannot totally reject the possibility that bioturbation and the relatively low temporal resolution at which these series were sampled degrade somehow the precession-related signals.

In contrast, spectral analyses of dissolution cycles in the giant (53 m long) piston core MD900963 (Maldives area) display clear maxima centered on the precession frequencies (23 and 19 kyr^{-1}) as well as on the 29-kyr^{-1} frequency but show little power at the 100-kyr^{-1} frequency. These high-frequency changes most probably result from changes in surface productivity associated with monsoon variability. Dissolution at this site may be ultimately controlled by the oxidation of

organic matter which appears to be incorporated into the sediments in greater quantity during periods of weak SW monsoon and/or increased dry NE monsoon.

The CCFI record displays a long-term, high-amplitude dissolution oscillation, with the well-known "mid-Brunhes dissolution cycle" representing the last part of this oscillation. It appears that this long-term oscillation, whose origin is not yet fully known, is not a true cycle and shows an irregular wavelength (of the order of ~500 kyr but varying from 425 to 550 kyr). We suggest that this oscillation may be associated with long-term changes in Ca^{++} flux to the ocean.

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