1995, Spatial and temporal variability of late Neogene equatorial Pacific carbonate

T. K. Hagelberg  
*University of Rhode Island*

Nicklas G. Pisias  
*Oregon State University*

Larry A. Mayer  
*University of New Hampshire*, larry.mayer@unh.edu

N. J. Shackleton  
*Cambridge University*

Alan C. Mix  
*University of New Brunswick*

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INTRODUCTION AND SCIENTIFIC BACKGROUND

Because the oceanic carbon reservoir is 60 times the size of the atmosphere and is in direct exchange with the atmosphere, orbital scale atmospheric CO$_2$ changes must ultimately be explained by changes in the oceanic carbon cycle (Broecker and Peng, 1982). Evidence from ice cores indicates that the CO$_2$ increase at the ends of glacial periods occurred rapidly, on the order of a thousand years. To understand this process, it is necessary to understand the mechanisms controlling the carbon cycle on glacial/interglacial time scales. While ice cores provide a direct measure of changes in atmospheric CO$_2$, examination of the marine sediment record is necessary to obtain a history beyond that of the last glacial cycle.

Efforts to understand the mechanisms through which the oceanic carbon cycle changes atmospheric CO$_2$ have led to years of carbon cycle modeling. Hypotheses for glacial/interglacial pCO$_2$ changes were grouped into four categories by Heine et al. (1991) in a recent overview: three kinds of oceanic carbon pumps that are capable of influencing atmospheric CO$_2$ (a solubility pump, a nutrient pump, and a CaCO$_3$ pump), and changes in the oceanic velocity field and ventilation rates. Evidence for oceanic carbon pool changes that is needed to evaluate hypotheses is derived primarily from sedimentary δ$^{13}$C and CaCO$_3$ records. δ$^{13}$C indicates the efficiency of the ocean's biological pump and the nutrient content of deep and surface waters. The CaCO$_3$ concentration in sediments reflects productivity, dilution, and alkalinity changes.

The extent to which CaCO$_3$ sedimentation in equatorial Pacific Ocean sediments has been controlled by production or dissolution has been the subject of debate for more than 40 yr. Since Arrhenius (1952), studies have investigated the relationship of equatorial Pacific carbonate sedimentation to regional surface processes, global glacial/interglacial changes, and external orbital (Milankovitch) influence (e.g., Hays et al., 1969; Moore et al., 1977, 1982; Farrell and Prell, 1989; Rea et al., 1991). In the equatorial Pacific Ocean, large changes in sedimentary calcium carbonate concentration have occurred on long time scales (millions of years), as a response to oceanic boundary condition and geochemical mass budget changes, and on shorter time scales (tens of thousands of years), as a response to climatic influence.

Changing boundary conditions during the late Cenozoic have significantly influenced equatorial Pacific sediment composition. van Andel et al. (1975) estimated the initiation of the Equatorial Undercurrent (EUC) at about 11 to 12 Ma, when the northward movement of the Australia Plate cut off western Pacific-Indian Ocean circulation. This development may have caused the onset of a narrow, equatorially symmetric zone of carbonate and opal sedimentation having steep gradients away from the equator. Widespread orogeny during the late Miocene and Pliocene may have had a large influence on oceanic alkalinity budgets through increased chemical weathering (Raymo et al., 1988). This uplift may have also influenced surface ocean circulation in the equatorial Pacific through associated changes in atmospheric circulation (Ravelo et al., 1992). A modeling study has suggested that prior to closure of the Isthmus of Panama (3-4 Ma or earlier), North Atlantic Deep Water (NADW) production was reduced, carbonate preservation was increased because of less undersaturated Pacific deep waters, and eastern equatorial Pacific surface upwelling was unchanged (Maier-Raime et al., 1990). Initiation of large-scale continental glaciation in the Northern Hemisphere near 2.4 Ma may have altered oceanic alkalinity and productivity significantly.

van Andel et al. (1975) demonstrated that similar spatial patterns are present in bulk-sediment accumulation rates, carbonate accumulation rates, and carbonate concentration of central equatorial Pacific
Ocean sediments throughout the Cenozoic. A high correlation of these variables with latitude reflected the “bulge” of equatorial Pacific productivity. Longitudinal gradients in carbonate accumulation across the central equatorial Pacific were small by comparison. Their results implied that the equatorial calcite compensation depth (CCD) has fluctuated by only a few hundred meters over the past 35 m.y. This constancy in the depth of the CCD requires that large changes in the gradient of CaCO$_3$ dissolution be compensated for by variations in lysocline depth and carbonate supply from surface waters.

On shorter time scales, regional studies have demonstrated that the depth and slope of the lysocline in the central equatorial Pacific Ocean changed dramatically over glacial and interglacial cycles. During the late Pleistocene in the equatorial Pacific, glacial intervals recorded higher CaCO$_3$ concentration and accumulation than interglacial periods (Farrell and Prell, 1989; Berger, 1973). If changes in deep ocean carbonate ion saturation produced the glacial increase, according to Archer’s model (1991a), a large (20-40 µm) change in the gradient of CaCO$_3$ saturation must have occurred over a fairly small (~100 m) depth interval. On the other hand, if changes in surface water carbonate production produced the glacial carbonate increase, the gradient in CaCO$_3$ saturation would have remained relatively homogeneous with depth. To explain increased glacial carbonate concentration and accumulation through productivity changes, Archer’s model requires a doubling of productivity during the last glacial relative to the present.

This estimate is consistent with independent estimates of glacial surface production in equatorial oceans (e.g., Mix, 1989; Finney et al., 1988; Pratih et al., 1989; Lyle et al., 1988).

Regardless of whether deep ocean or surface ocean processes are primarily responsible for the observed changes in equatorial Pacific carbonate sedimentation on 10- to 100-k.y. scales, a temporal association with Pleistocene glacial/interglacial changes is well known. Moore et al. (1977) documented a 6 to 8-k.y. lag between ice volume and CaCO$_3$ concentration in the equatorial Pacific. Farrell and Prell (1989) documented a similar relationship in a reconstruction of late Pleistocene central equatorial Pacific lysocline changes. This lag is thought to result from the response time of oceanic CO$_2$ compensation of glacial/interglacial TCO$_3$ changes (Broecker and Peng, 1982). However, in detail the ice volume-carbonate relationship is not straightforward. While some studies have shown carbonate variations during the late Pleistocene to be coherent with ice volume variations (Rea et al., 1991; Farrell and Prell, 1989; Luz and Shackleton, 1975), others have indicated either no relationship or a complex one (e.g., Moore et al., 1982; Pisas and Rea, 1988, Lyle et al., 1988). Spatial variability in carbonate sedimentation, as noted by Murray (1987) and Farrell and Prell (1989, 1991), probably contributes to the different conclusions reached by studies that focused on selected locations.

Time series studies spanning intervals beyond the late Pleistocene indicate a complex relationship between equatorial Pacific carbonate and ice volume. For instance, a comparison of carbonate and ice volume proxy ($\delta^{18}O$) records spanning the past 4 m.y. (Farrell and Prell, 1991) suggests that the relationship between equatorial Pacific lysocline fluctuations and ice volume is not straightforward. It is well known that a change in the amplitude of global ice volume variability from principally 41 k.y. oscillations to 100 k.y. oscillations occurred near 1Ma (Ruddiman et al., 1989; Raymo et al., 1989; Pisas and Moore, 1981). A corresponding change in the amplitude of CaCO$_3$ variability is not evident from Farrell and Prell’s data. Although low sedimentation rates and uncertainties in chronology have made it difficult to determine how CaCO$_3$ and $\delta^{18}O$ have covaried over the past 4 m.y. in the equatorial Pacific, the records compared by Farrell and Prell suggest that the relationships observed in late Pleistocene age sediments were not present during the early Pleistocene and Pliocene.

**RESEARCH STRATEGY**

The following generalities can be made from the above information: (1) despite many studies, the relationship between equatorial Pacific carbonate sedimentation and global climate changes over the past several million years is not well defined. Strong spatial variability probably contributes; (2) the debate over production vs. dissolution as a control on eastern equatorial Pacific carbonate sedimentation is not resolved; and (3) most previous studies were limited to data sets restricted in either spatial coverage or temporal resolution.

The suite of sites sampled during Leg 138 provide additional high resolution carbonate and $\delta^{18}O$ records from the equatorial Pacific which can clarify these relationships. These records provide (1) spatial coverage of a very dynamic region; (2) high temporal sampling resolution (order of 1-2 k.y.); (3) long temporal coverage (>6 m.y.); (4) an internally consistent chronology; and (5) high-quality proxy records.

The objective of the present study is to determine if effects on carbonate sedimentation resulting from surface ocean processes, which will be more regional in nature, can be discriminated from effects caused by changes in deep ocean chemistry, which will be more uniform in nature. Two longitudinal transects of sites cross the high productivity region of the equatorial Pacific, allowing for comparison of latitudinal and (to a limited extent) longitudinal gradients of this system (Fig. 1). The latitudinal and longitudinal positions of the drilled sites have changed by no more than 0.5° latitude and 4° longitude over the past 6 m.y. due to west-northwest movement of the Pacific Plate (Pisas et al., this volume). The sites are at water depths within the lysocline (~3200 m, Parker and Berger, 1971; Adelseck and Anderson, 1978) and above the calcite saturation horizon (~4000 m, Broecker and Peng, 1982) (Table 1). While data are located at different depths, no correlation exists for depth vs. latitude that would complicate determination of whether changes in carbonate result from changes in productivity or from changes in lysocline depth (Fig. 2 and Table 1). These sites are shallower than the sites studied by Farrell and Prell (1991). However, because the equatorial Pacific lysocline also shoals to the east, these two data sets are comparable with respect to upper lysocline fluctuations.

In this study, carbonate proxy records from gamma-ray attenuation porosity evaluator (GRAPE; Boyce, 1976) wet-bulk density measurements at Sites 846 through 853 were analyzed. Site 844, located in the Costa Rica Dome, and Site 845, located in the Guatemalan Basin, represent different oceanographic environments and were dominantly siliceous clay during much of the late Neogene and were not considered in the present study. The continuity of the records from Sites 846 through 853 is documented in detail over the last 6

![Figure 1. Present-day location of Leg 138 Sites 844 through 854, superimposed on a schematic of the general circulation of the eastern equatorial Pacific Ocean. Solid arrows indicate surface currents, and dashed arrow indicates the Equatorial Undercurrent (EUC). Shaded regions illustrate the latitudinal extent of the South Equatorial Current (SEC) and North Equatorial Current (NEC). NECC = North Equatorial Countercurrent; CAC = California Current; CHC = Chile Current; and PC = Peru Current.](image)
GRAPE records from each hole at Sites 846 through 852 were analyzed. Finally, these results were used to examine variations over the past 4.6 m.y.

The second term in the above equation reflects the linear grain-density relationship, and the last term reflects the nonlinear packing factor. Herbert and Mayer (1991) demonstrated that the sensitivity of sedimentary wet-bulk density to \( \text{CaCO}_3 \) concentration is controlled by the high porosity of the noncarbonate (opal) fraction. Thus, although biogenic opal has a lower grain density than carbonate (2.1-2.3 g/cm\(^3\), as compared to 2.6-2.7 g/cm\(^3\); Mayer, 1979), porosity differences are most important.

Stacked GRAPE records developed from the adjacent holes at each site (Hagelberg et al., this volume; Fig. 3) and a spliced record from Site 853 were used in the carbonate predictions. The effectiveness of GRAPE data as a carbonate predictor is indicated by the scatter plots in Figure 4, which compare shipboard \( \text{CaCO}_3 \) measurements to predicted values from matching samples in Sites 846 through 853. The sample cross correlation, or model skill, that gives the fraction of variance explained by the predictive model, is shown in Figure 4. The temporal evolution of carbonate sedimentation was studied in 1-m.y. time steps. Analysis of individual frequency bands allowed estimation of carbonate concentration (%), which is influenced by dilution from opal and carbonate variability. These results are related to sedimentary carbonation independent of opal fluctuations, carbonate MAR time series were analyzed. Finally, these results were used to examine interactions between carbonate sedimentation and global ice volume variations over the past 4.6 m.y.

METHODS

Continuous Records

Each of the 11 Leg 138 drill sites was multiple-APC cored in an effort to document recovery of a continuous sedimentary sequence. At each site, records of magnetic susceptibility, GRAPE wet-bulk density, and sediment color reflectance were used to monitor section recovery and to construct a composite depth section (Hagelberg et al., 1992; Mayer, Pisias, Janecek, et al., 1992). Multiple measurements of these sedimentary parameters on adjacent holes provide multiple realizations of the same sedimentary process at every site. The GRAPE records from each hole at Sites 846 through 852 were 'stacked' to provide a statistically robust and less noisy estimate of sediment bulk density (Hagelberg et al., this volume). The average sampling interval of the stacked, smoothed GRAPE records is 2 cm (Fig. 3). The temporal resolution of the records is on the order of 1000 yr.

GRAPE Data as a Carbonate Proxy

Sample resolution traditionally has been a problem when studying pre-Pleistocene spatial and temporal climatic variations at high resolution because of time and cost restrictions. During Leg 138, this problem was addressed through the use of \( \text{CaCO}_3 \) proxy records derived from GRAPE bulk density at near-continuous resolution at all sites. Development of proxy \( \text{CaCO}_3 \) estimates was possible because equatorial Pacific sediments are primarily composed of two components: biogenic calcite and opal. Owing to this composition, bulk density is a good predictor of carbonate in this region (Mayer, 1991; Herbert and Mayer, 1991). Wet-bulk density measurements record two primary differences between carbonate and opal: their grain densities and their packing properties. These relationships led to development of an empirical relationship that predicts carbonate on this basis (Mayer, 1979, 1991). After correction to account for porosity rebound and depth, Mayer's (1991) equation was used to predict carbonate concentration at Sites 846 through 853:

\[
\%\text{CaCO}_3 = -835.52 + 1112.69p - 332.54p^2
\]

The second term in the above equation reflects the linear grain-density relationship, and the last term reflects the nonlinear packing factor. Herbert and Mayer (1991) demonstrated that the sensitivity of sedimentary wet-bulk density to \( \text{CaCO}_3 \) concentration is controlled by the high porosity of the noncarbonate (opal) fraction. Thus, although biogenic opal has a lower grain density than carbonate (2.1-2.3 g/cm\(^3\), as compared to 2.6-2.7 g/cm\(^3\); Mayer, 1979), porosity differences are most important.
The correlatability of GRAPE records over the Leg 138 sites contributed to development of an internally consistent chronostratigraphic framework for Leg 138 sites during the cruise (Shackleton et al., 1992). A series of events, identified primarily on the basis of biostratigraphy and magnetostratigraphy, were related to patterns in the GRAPE bulk density records. Correlations between the GRAPE bulk density records were used to identify these events at every site. During post-Leg 138 study, Shackleton et al. (this volume) demonstrated that very high stratigraphic resolution is possible if the assumptions of an orbitally driven system are considered. Variations in GRAPE and δ¹⁸O were correlated to variations in 65°N solar insolation (as determined by Berger and Loutre, 1988) to develop a chronology that is both internally consistent and astronomically calibrated over the past 6 m.y. (Shackleton et al., this volume).

Because no a priori model was available for predicting appropriate time constants for the response of equatorial Pacific carbonate to orbital forcing, zero phase between insolation and GRAPE was initially assumed (Shackleton et al., this volume). The further assumption that correlatable GRAPE events at all sites were in phase with one another was necessary to correlate the records at high resolution. These assumptions have bearing on the resulting time-varying spatial variability in the carbonate records. As measurements of benthic δ¹⁸O from multiple sites emerged, providing independent measurements that were spatially in phase, the assumption of no phase difference between individual GRAPE events was modified and the astronomically calibrated chronology was adjusted accordingly (Shackleton et al., this volume). Because the GRAPE-calibrated (δ¹⁸O independent) chronology and the chronology developed using δ¹⁸O data have slight differences in some intervals (on the order of a few thousand years), Shackleton et al. (this volume) presented two alternative chronologies for the upper 1 m.y. at the Leg 138 sites. In this study, the alternative chronology that makes use of the δ¹⁸O data is used. The resulting time series of predicted CaCO₃ concentration from Sites 846 through 853 for the interval from 0 to 6 Ma is given in Figure 5.

**Mass Accumulation Rates**

Although a great deal of information can be gained from analysis of CaCO₃ concentration alone, carbonate concentration is not the most sensitive indicator of carbonate dissolution (Heath and Culber-son, 1970). Dilution by biogenic opal and other sedimentary components influences carbonate concentration in sediments. Farrell and Prell (1989, 1991) assumed that opal sedimentation in the central equatorial Pacific was sufficiently negligible and that CaCO₃ concentration sufficiently approximated carbonate mass burial. In the eastern tropical Pacific Ocean, however, opal is often a significant sedimentary component. Thus, carbonate mass accumulation rates were estimated to evaluate objectively the role of opal dilution when determining spatial variability in CaCO₃ concentration records.

Calculation of mass accumulation rates incorporates several sources of error. Standard errors in carbonate prediction were estimated in the preceding section (Table 2) and on average are about 7%. Second, the dry-bulk density of the sediment must be known. Third, a highly constrained chronology is required to determine sedimentation.

**Figure 4. Scatter plots of predicted (from GRAPE) vs. measured %CaCO₃ from the top 6 m.y. of Sites 846 through 853.**

**Chronology**

**Figure 5. The 6-m.y. records of predicted %CaCO₃ for Sites 846 through 853.**
Table 3. Grain density statistics, Sites 846 through 853.

<table>
<thead>
<tr>
<th>Site</th>
<th>N</th>
<th>Mean</th>
<th>Std. dev.</th>
</tr>
</thead>
<tbody>
<tr>
<td>846</td>
<td>398</td>
<td>2.50</td>
<td>0.12</td>
</tr>
<tr>
<td>847</td>
<td>375</td>
<td>2.54</td>
<td>0.10</td>
</tr>
<tr>
<td>848</td>
<td>115</td>
<td>2.59</td>
<td>0.08</td>
</tr>
<tr>
<td>849</td>
<td>317</td>
<td>2.59</td>
<td>0.07</td>
</tr>
<tr>
<td>850</td>
<td>260</td>
<td>2.60</td>
<td>0.07</td>
</tr>
<tr>
<td>851</td>
<td>248</td>
<td>2.63</td>
<td>0.07</td>
</tr>
<tr>
<td>852</td>
<td>135</td>
<td>2.72</td>
<td>0.05</td>
</tr>
<tr>
<td>853</td>
<td>75</td>
<td>2.73</td>
<td>0.05</td>
</tr>
</tbody>
</table>

Estimates of sedimentation rates have been derived directly from tie points in the astronomically calibrated chronology of Shackleton et al. (this volume). Average spacing of age control points is 25 to 30 k.y. with two exceptions: Sites 848 and 853, which have much lower sedimentation rates. Estimates of errors in the sedimentation rates are difficult to obtain. However, comparison of the tuned chronology with variations typically seen between adjacent holes (Hagelberg et al., this volume) suggests that variability in sedimentation rates on the order of 15% are typical. Assuming that the three sources of error estimated above are uncorrelated, the total error of MAR estimates at most 7% + 8% + 15% = 30%. This is a rough approximation, but it serves to indicate the higher error introduced into MAR estimates. Estimates of carbonate MAR for Sites 846 through 853 for the interval 0 to 6 m.y. are given in Figure 7.

**EOF Analyses**

Each record from Sites 846 through 853 displays some level of variability that is common to all of the sites, as well as variability that is unique to that site. The common variability was used by Shackleton et al. (1992) as a means of correlating biostratigraphic and magnetostratigraphic events among sites, as described above. The goal of this study is to partition patterns of variability in equatorial Pacific Ocean carbonate sedimentation in a quantitative manner.

The numerical approach used to partition the eight time series from Sites 846 through 853 into a smaller set of spatial and temporal

![Figure 6](image-url)
Table 4. Statistics for predicted values of dry-bulk density compared with measured values, Sites 846 through 853.

<table>
<thead>
<tr>
<th>Site</th>
<th>N</th>
<th>Mean</th>
<th>Std. dev.</th>
<th>Mean</th>
<th>Std. dev.</th>
<th>Skill</th>
<th>Crit. val.</th>
<th>RMS</th>
</tr>
</thead>
<tbody>
<tr>
<td>846</td>
<td>398</td>
<td>0.59</td>
<td>0.14</td>
<td>0.59</td>
<td>0.14</td>
<td>0.90</td>
<td>0.15</td>
<td>0.059</td>
</tr>
<tr>
<td>847</td>
<td>375</td>
<td>0.55</td>
<td>0.15</td>
<td>0.54</td>
<td>0.14</td>
<td>0.91</td>
<td>0.16</td>
<td>0.059</td>
</tr>
<tr>
<td>848</td>
<td>115</td>
<td>0.57</td>
<td>0.09</td>
<td>0.58</td>
<td>0.09</td>
<td>0.83</td>
<td>0.28</td>
<td>0.049</td>
</tr>
<tr>
<td>849</td>
<td>317</td>
<td>0.67</td>
<td>0.13</td>
<td>0.65</td>
<td>0.12</td>
<td>0.83</td>
<td>0.17</td>
<td>0.086</td>
</tr>
<tr>
<td>850</td>
<td>260</td>
<td>0.63</td>
<td>0.12</td>
<td>0.61</td>
<td>0.11</td>
<td>0.83</td>
<td>0.19</td>
<td>0.059</td>
</tr>
<tr>
<td>851</td>
<td>248</td>
<td>0.63</td>
<td>0.10</td>
<td>0.62</td>
<td>0.10</td>
<td>0.89</td>
<td>0.19</td>
<td>0.045</td>
</tr>
<tr>
<td>852</td>
<td>135</td>
<td>0.58</td>
<td>0.12</td>
<td>0.57</td>
<td>0.11</td>
<td>0.86</td>
<td>0.26</td>
<td>0.045</td>
</tr>
<tr>
<td>853</td>
<td>74</td>
<td>0.52</td>
<td>0.09</td>
<td>0.55</td>
<td>0.09</td>
<td>0.71</td>
<td>0.35</td>
<td>0.063</td>
</tr>
</tbody>
</table>

The model skill gives the fraction of variance explained by the predictive mode.

Crit. val. = 0.99 level for significant skill.

RMS = root-mean-square-error.

Spectral Analyses

Power spectra and cross-spectra of the amplitude time series from the EOF analyses indicate the frequency distribution of variance of each EOF, as well as the extent to which each EOF is coherent with Milankovitch band variability. Spectral estimates were made using the Blackman-Tukey lagged auto- (or cross-) covariance method (Jenkins and Watts, 1968). Sampling intervals were 1000 yr. For analyses of the entire 6 m.y. records, 600 lags of the auto and cross-covariance and a cosine taper window were used for smoothed estimates having at least 27 degrees of freedom. For analyses of 1 m.y. intervals, 400 lags of the auto or cross covariance were used for smoothed estimates having at least 7 degrees of freedom.

RESULTS AND DISCUSSION

General Patterns of Carbonate Variability (0–6 Ma)

One can generally conclude from the patterns of carbonate concentration given in Figure 5 that multiple processes, including productivity and dissolution, are important for determining the space-time variability of equatorial Pacific Ocean carbonate sedimentation and preservation. The variations of these multiple processes are best quantified using EOF analysis with one time series and one spatial map describing each independent mode of variation. An EOF analysis of the records for 0 to 6 Ma for Sites 846 through 853 resolves two modes of variability that together explain 76% of the total variance. The spatial patterns of these two modes are presented in map form in Figure 8, and the amplitude time series of the two spatial modes are given in Figure 9. The influence of dissolution relative to carbonate productivity when determining these patterns is discussed in the section following presentation of the results.

Figure 7. CaCO₃ mass accumulation rates in g CaCO₃/cm²/ky., 0 to 6 Ma for Sites 846 through 853. See text for description of calculations. y-axis scale is given in the bottom panel.

Figure 8. EOF-1 (top) and EOF-2 (bottom) of %CaCO₃, 0 to 6 Ma. Values contoured are the loadings for each EOF (elements of the eigenvector), normalized by the length of the eigenvector so that the average loading is 1.0.
The first EOF for the 0- to 6-Ma CaCO$_3$ record accounts for 57% of the total variance. The spatial map of EOF-1 (Fig. 8, top) has highest loading at eastern transect Sites 846 and 847. On the western transect, the highest loading is at Site 849, with loadings decreasing to the north and south. This pattern suggests that the dominant mode of CaCO$_3$ variability may be related to equatorial divergence.

The amplitude time series (Fig. 9, top) and power spectrum (Fig. 10) of EOF-1 indicate dominance by a combination of low frequency (200-250 k.y.) and higher-frequency Milankovitch band oscillations. Very high coherence is present between insolation at 65°N (Berger and Loutre, 1988) and EOF-1 at periods of 41, 23, 19, and 17 k.y. (Fig. 10 and Table 5). 65°N insolation and EOF-1 are in phase where coherent, with the exception of the 100-k.y. period, where phase is near -180°. The near-zero phase at the dominant insolation frequencies reflects the orbital calibration of the chronology.

The spatial map for EOF-2 accounts for more than 19% of the total variance. The highest loadings are off-equator on the 110°W transect, at Sites 846, 852, and 853, and at eastern transect Sites 846 and 847, where loadings are of the opposite sign (Fig. 8, bottom). The western transect equatorial sites have minimum loading. Thus, EOF-2 appears to represent a mode of variability having both north-south and east-west asymmetry.

The amplitude time series for EOF-2 (Fig. 9, bottom) indicates a distinct step in amplitude between 3 and 4 Ma, at approximately 3.6 Ma. This change in the variability of EOF-2 might be related to changes in external boundary conditions that affect carbonate sedimentation, such as the closure of the Panamanian isthmus. Changes in the spatial mode of variability in EOF-2, described below, support such a change in EOF-2 during this time interval.

Variance in the amplitude time series of EOF-2 (Fig. 10) is dominated by low frequencies that are not coherent with insolation variations. Although significant (0.80 level) coherence is present at some Milankovitch frequencies (Fig. 10; Table 5), at a 0.90 level, none of the frequencies in the EOF-2 time series are coherent with insolation. In the precession band where insolation and EOF-1 are also coherent, insolation and EOF-2 have a phase close to 180° (143 ±26° for 23 k.y., and 170 ±21° for 19 k.y.). This separation of Milankovitch band variability into two modes may imply that multiple processes are operating at the same frequencies.

**Evolution of CaCO$_3$ Variability**

As discussed in the "Introduction" section (this chapter), large changes in both climatic and tectonic boundary conditions have occurred over the past 6 m.y. In accord with previously documented changes, large changes in carbonate and bulk sedimentation rates occurred at all sites from 6 to 4.7 Ma (Mayer et al., 1992). It is not well understood how these changes have influenced the modes of variability in the equatorial Pacific CaCO$_3$. Evolutive spectra of CaCO$_3$ variability also suggest a transition near 1 Ma, which coincides with changes in global ice volume (Mayer et al., this volume).

### Table 5. Results of cross-spectral analyses between 65°N insolation and EOFs 1 and 2, 0 to 6 Ma, %CaCO$_3$.

<table>
<thead>
<tr>
<th>Period (k.y.)</th>
<th>Coh.</th>
<th>Phase (degrees)</th>
<th>Period (k.y.)</th>
<th>Coh.</th>
<th>Phase (degrees)</th>
</tr>
</thead>
<tbody>
<tr>
<td>125</td>
<td>0.43</td>
<td>-176 ±28</td>
<td>100</td>
<td>0.43</td>
<td>-137 ±29</td>
</tr>
<tr>
<td>41</td>
<td>0.86</td>
<td>-9 ±9</td>
<td>67</td>
<td>0.54</td>
<td>-51 ±22</td>
</tr>
<tr>
<td>36</td>
<td>0.81</td>
<td>-156 ±11</td>
<td>23</td>
<td>0.47</td>
<td>143 ±26</td>
</tr>
<tr>
<td>23.8</td>
<td>0.92</td>
<td>-10 ±6</td>
<td>19</td>
<td>0.45</td>
<td>-175 ±27</td>
</tr>
<tr>
<td>22.2</td>
<td>0.93</td>
<td>-5 ±6</td>
<td>17</td>
<td>0.95</td>
<td>96 ±5</td>
</tr>
<tr>
<td>19</td>
<td>0.88</td>
<td>-5 ±8</td>
<td>12</td>
<td>0.51</td>
<td>-30 ±24</td>
</tr>
<tr>
<td>17</td>
<td>0.96</td>
<td>-44 ±4</td>
<td>11</td>
<td>0.62</td>
<td>-36 ±18</td>
</tr>
<tr>
<td>15</td>
<td>0.66</td>
<td>64 ±19</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>14</td>
<td>0.46</td>
<td>27 ±26</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>12</td>
<td>0.52</td>
<td>22 ±23</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>10</td>
<td>0.75</td>
<td>0 ±13</td>
<td></td>
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</tr>
</tbody>
</table>

*Coh. = Coherence for periods significant at 0.80 level.

To account for these changes and also to determine the modes of variability that characterize carbonate sedimentation, separate EOF analyses were performed on successive 1-m.y. time intervals.

The spatial maps of the two dominant EOFs for each time slice are shown in Figure 11. The Leg 138 sites in each spatial map have been backtracked and are located at approximate paleolatitudes, according to the rotation of Cox and Engebretson (1985; note that the present-day geography is superimposed on the backtracked site locations in Fig. 11). In each time interval, two modes dominate. The total amount of variance explained by these two modes is consistent, with EOF-1 representing from 44% to 67%, and EOF-2 representing from 15% to 32% of the overall variance. In only one time interval, from 5 to 6 Ma, is the variance in the third EOF (not shown) greater than 10%.

With the exception of Site 852, which has a high loading from 3 to 4 Ma and from 5 to 6 Ma, the spatial pattern of EOF-1 in each time slice shows coupling to the equator (Figs. 11A–11F). As with the results from 0 to 6 Ma, the spatial pattern of EOF-1 in each interval suggests an influence of surface ocean processes on variations in CaCO$_3$ concentration. The EOF loadings are shown as a function of latitude across the 110°W transect in Fig. 12A. High loadings in EOF-1 occur at the equator, with a secondary peak at 4° to 5°N (Site 852). The relationship between modern water depth and the EOF...
loadings for each time slice is shown in Figure 12B. Although relationships are difficult to resolve because of close depth spacing of Sites 848, 849, and 852, the loadings of EOF-1 do not show any overall correlation with depth.

The spatial pattern of EOF-2 has similar features in the time intervals from 0 to 1 Ma through 3 to 4 Ma (Figs. 11G-11L). Like the analysis for the period from 0 to 6 Ma, a pattern of out-of-phase east-west loadings is present, with eastern transect Sites 846 and 847 having loadings opposite in sign of the western transect loadings. On the western transect, equatorial Sites 849 and 850 have minimum loadings, while the off-equatorial sites have higher loadings. The highest positive loadings are at Site 852, and highest negative loadings are at Site 846. Prior to 3 to 4 Ma, a different spatial pattern is present. The interval from 4 to 5 Ma, while showing a high loading at Site 852, does not display the east-west symmetry of the interval from 0 to 4 Ma. The spatial pattern of EOF-2 from 5 to 6 Ma does not appear to have any similarities with the later time intervals. These features also are evident in Figures 12C and 12D, which show latitudinal and depth gradients for EOF-2. Figure 12D indicates a correlation between water depth and EOF-2 loadings, particularly between the eastern and western transects.

The change in the spatial pattern of EOF-2 between 3 to 4 Ma and 4 to 5 Ma may be related to changes associated with closure of the eastern and western transects.

Panamanian isthmus. This interpretation is highly dependent on the EOF loading at Site 847 and, thus, should be taken with caution. However, Figure 12C also indicates a different spatial pattern of EOF-2 on the 110°W transect from 3 to 4 Ma and 5 to 6 Ma than in later periods. This difference is not dependent on Site 847. In addition, the amplitude time series of EOF-2 for the entire 0- to 6-Ma interval (Fig. 9) suggests a change in character between 3 and 5 Ma. Together, these patterns suggest that EOF-2 indicates a response of carbonate sedimentation to a change in oceanographic boundary conditions between 3 to 4 Ma and 4 to 5 Ma.

The amplitude time series for EOF-1 and EOF-2 for each 1-Ma, time slice are given in Figures 13A through 13F. In general, the EOF-1 time series have more distinct concentrations of variance than that in the EOF-2 time series, and the EOF-2 time series have little orbital-band variability (Fig. 14). Coherence with insolation is extremely high for EOF-1 in each time interval (Figs. 14A-14F and Table 6) and is significant at a 0.80 level (coherence = 0.67) as well as at a 0.90 level (coherence = 0.88). These strong linear relationships indicate that the dominant variability in the CaCO₃ record over the past 6 m.y. includes Milankovitch variability. EOF-2 has high coherence with precession band variations only from 3 to 4 Ma and 5 to 6 Ma (Figs. 14G-14L). As with the previous EOF analysis of the 0 to 6 Ma interval, near-zero phase between 65°N insolation and EOF-1 reflects the constraint imposed by the orbital calibration of the time scale. Where coherence between EOF-2 and insolation is significant, the phase indicates that two processes are operating at the same Milankovitch frequencies, each of which has a different relationship to insolation. Most of the coherence observed between EOF-2 and insolation is not coherent at a 0.90 level, and occurs at "non-Milankovitch" frequencies. Thus, most of the variability in EOF-2 is not linearly related to insolation.

**Patterns in Carbonate Mass Accumulation Rates**

The role of noncarbonate dilution must be considered before one makes any conclusions with respect to the role of carbonate production relative to carbonate dissolution in driving the spatial patterns of CaCO₃. Is the carbonate concentration record reflecting dilution by noncarbonate sedimentary components (primarily biogenic opal) more than processes related to carbonate sedimentation? Analysis of carbonate MAR time series can address this issue. If spatial patterns similar to the carbonate concentration EOFs are present, then dilution...
can be ruled out as the primary cause of the spatial variability in carbonate. An EOF analysis was done using the 0- to 6-Ma carbonate MAR time series (Fig. 7). As with CaCO$_3$ concentration data, two EOFs explain 77% of the total variance in the time series for Sites 846 through 853. Spatial maps for EOF-1, which explains 65% of the total variance, and EOF-2, which explains 12% of the total variance, are given in Figure 15.

The carbonate MAR EOFs confirm the inferences drawn from the CaCO$_3$ concentration results. A strong coupling to the equator is present in the spatial pattern of EOF-1, indicating the domination of the influence of surface water on carbonate sedimentation. The pattern is similar to EOF-1 of carbonate concentration (Fig. 8). While in some short intervals of Sites 846 through 853 low carbonate concentration is indeed a result of dilution by noncarbonate, for example, time intervals dominated by monospecific diatom ooze (Kemp and Baldauf, this volume), these intervals are not representative of the dominant variability over all records, and the EOF analyses represent the time-integrated modes of variability. Thus, although dilution may play some role in influencing CaCO$_3$ variability, it is not the dominant process. The second EOF for the carbonate MARs displays strong east-west asymmetry. The spatial pattern is not similar to the second EOF for the CaCO$_3$ concentration, suggesting that they are reflecting different oceanographic processes.

The loadings of EOF-1 across the 110°W transect (Fig. 16A) highlights the strong coupling to the equator. These loadings parallel variations in sediment thickness (and thus sedimentation rates). However, a normalized sedimentation rate comparison between sites has indicated that attenuation of records at the lower sedimentation rate sites (848, 852, and 853) is not simply a function of sedimentation rate (Pisias et al., 1992). Thus, the equatorial pattern displayed by EOF-1 cannot be entirely a function of sedimentation rate. The latitudinal transect for EOF-2 of carbonate MAR (Fig. 16C) does not indicate a clear relationship. Neither EOF-1 nor EOF-2 of carbonate MAR appears to have a correlation with water depth (Figs. 16B and 16D).

The frequency resolution of MAR time series is limited by the resolution of the sedimentation rates of the records. If the low sedimentation rate resolution at Site 853 (~200 k.y.) is taken as the limiting factor, spectral analyses are not appropriate for the carbonate MAR EOFs. Even so, cross-spectral analyses among the CaCO$_3$ MAR EOFs (not shown) indicated high coherence between the EOF-1 amplitude time series and 65°N insolation. Comparisons between EOF-1 and EOF-2, respectively, of the CaCO$_3$ concentration and MAR records indicates that EOF-1 of carbonate concentration and carbonate MAR are highly positively correlated. In the frequency domain, the two EOFs are highly coherent with one another (often greater than 0.95) at a broad range of frequencies. From these results, it is reasonable to conclude that the dominant mode of variability in carbonate concentration and carbonate MAR reflect the same fundamental oceanographic process and that over the past 6 m.y., the CaCO$_3$ concentration spatial pattern is not driven dominantly by dilution.

EOF-2 of CaCO$_3$ concentration and MAR appear to reflect different processes. The two modes of variability are not highly correlated from 0 to 6 Ma. The spatial maps (Figs. 8 and 15) also differ. While EOF-2 of CaCO$_3$ concentration may be influenced by dissolution processes, EOF-2 of carbonate MAR indicates a process operating in
The carbonate system that is independent of both the dominant modes of variability (EOF-1) and EOF-2 of carbonate concentration.

**Frequency Domain Analyses**

The observation that much of the variability represented in EOF-1 is strongly coherent with solar insolation raises yet another question: does each frequency band have the same spatial structure? The analyses presented above consider all frequencies of each time series together, to determine empirically the independent modes of variability. As demonstrated by Imbrie et al. (1993), the processes controlling climatic variations operate differently in the primary Milankovitch frequency bands. It is physically meaningful to examine three of the dominant frequency bands of the carbonate time series separately. For instance, the 23- to 19-k.y. band associated with precession variations may have a different spatial representation than the 41-k.y. band associated with orbital obliquity. In addition, at frequencies lower than 41 k.y., where little variance is present in insolation, most of the variance in carbonate variability is concentrated.

Figure 14. A–F. Power and coherence spectra for EOF-1 and 65°N insolation, for the 1-m.y. time slice amplitude time series given in Figure 13. G–L. Power and coherence spectra for EOF-2 and 65°N insolation for the amplitude time series given in Figure 13. Symbols are as described in the caption for Figure 10.

Figure 15. EOF-1 (top) and EOF-2 (bottom) of the CaCO$_3$ MAR time series given in Figure 7.

Figure 16. A. EOF-1 loadings of the CaCO$_3$ MAR EOFs on the 110°W transect. B. EOF-1 loadings as a function of present water depth. C. EOF-2 loadings on the 110°W transect. D. EOF-2 loadings as a function of water depth.
The CaCO$_3$ concentration time series from Sites 846 through 853 were reanalyzed with this objective in mind. Each record was low- and band-pass filtered to isolate three groups of frequencies: frequencies lower than 45 cycles/k.y. ("low-pass"), frequencies from 45 to 35 cycles/k.y. ("tilt band"), and frequencies from 25 to 18 cycles/k.y. ("precession band"). The "low-pass" filter used a cosine taper having a half-amplitude at 0.016 cycles/k.y. (62.5 k.y.), which excluded energy at frequencies higher than 45 cycles/k.y. The tilt band filter was centered on 0.024 cycles/k.y. (41 k.y.), and had a bandwidth of 0.02 cycles/k.y. The precession band filter was centered at 0.048 cycles/k.y. (20.8 k.y.) and had a bandwidth of 0.03 cycles/k.y.

The results for EOF-1 of each of these analyses are shown in Figures 17 through 20 (EOF-2 of each frequency band represented substantially less of the total variance and is not discussed here). The spatial patterns of EOF-1 of the low-pass carbonate time series from 0 to 6 Ma are similar to the unfiltered EOFs (Figs. 11A–11F). This is expected as the dominant variability for the unfiltered EOF-1s were at low frequencies (Figs. 14A–14F). A high loading at Site 852 relative to Sites 851 and 853 is present from 3 to 4 Ma and from 5 to 6 Ma (Fig. 17). With the exception of 3 to 4 Ma and 5 to 6 Ma, the maximum loading among the 110°W transect sites is at the equator (Fig. 18A). Loadings as a function of depth (Fig. 17B) are relatively constant and are not strongly correlated with depth.

The spatial patterns of EOF-1 in the tilt band have a spatial pattern with high loadings on the equator and at Site 852 at ~6°N (Fig. 19). Comparison to the low-pass results indicate a higher coupling to the equator in this band than in the low-frequency band. Unlike the low-pass filtered records, the latitudinal transect (Fig. 18C) clearly

![Figure 18](image)

**Figure 18.** A. EOF-1 loadings of low-pass filtered %CaCO$_3$ time slices on the 110°W transect. B. Low-pass EOF-1 loadings for each time slice as a function of present-day water depth. C. Tilt band %CaCO$_3$ EOF-1 loadings on the 110°W transect. D. Tilt band EOF-1 loadings as a function of water depth. E. Precession band %CaCO$_3$ EOF-1 loadings on the 110°W transect. F. Precession band EOF-1 loadings as a function of water depth.

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**Table 6. Summary of cross-spectral results between 65°N insolation and CaCO$_3$: EOFs 1 and 2.**

<table>
<thead>
<tr>
<th>Period (ky.)</th>
<th>Coh.</th>
<th>Period (degrees)</th>
<th>Coh.</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-1 Ma EOF-1</td>
<td>0.86</td>
<td>-101 ± 18</td>
<td>0.89</td>
</tr>
<tr>
<td>4-5 Ma EOF-2</td>
<td>0.71</td>
<td>54 ± 27</td>
<td></td>
</tr>
<tr>
<td>1-2 Ma EOF-1</td>
<td>0.79</td>
<td>102 ± 23</td>
<td>0.83</td>
</tr>
<tr>
<td>2-3 Ma EOF-1</td>
<td>0.97</td>
<td>-13 ± 8</td>
<td>0.79</td>
</tr>
<tr>
<td>3-4 Ma EOF-1</td>
<td>0.92</td>
<td>47 ± 13</td>
<td>0.78</td>
</tr>
<tr>
<td>4-5 Ma EOF-1</td>
<td>0.83</td>
<td>-3 ± 20</td>
<td>0.81</td>
</tr>
<tr>
<td>5-6 Ma EOF-1</td>
<td>0.81</td>
<td>64 ± 27</td>
<td>0.80</td>
</tr>
<tr>
<td>0.81</td>
<td>64 ± 27</td>
<td>0.80</td>
<td>-125 ± 22</td>
</tr>
<tr>
<td>0.81</td>
<td>64 ± 27</td>
<td>0.80</td>
<td>-125 ± 22</td>
</tr>
</tbody>
</table>

---

**Figure 17.** A–F. EOF-1 of the low-pass filtered %CaCO$_3$ times series in 1-m.y. time slices, 0 to 6 Ma.
error estimates for the EOF loadings are necessary to determine if equator and strong gradients away from the equator (Fig. 20). Variability at depth is evident (Fig. 18D).

6°N (Site 852). No strong correlation between loadings and water depth at Site 852 is diminished relative to previous analyses. In every 1-m.y. time slice, Site 852 relative to Sites 851 and 853 has the least coupling to the equator and to near 6°N (Site 852). No strong correlation between loadings and water depth at Site 852 is diminished relative to previous analyses. In every 1-m.y. time slice, Site 852 relative to Sites 851 and 853 has the least coupling to the equator and to surface ocean processes. The high loadings at Site 852 in the unfiltered data set primarily reflect processes that are operating at lower frequencies. In the tilt band, variability appears to reflect surface oceanographic processes operating on the equator and near 6°N, near the NECC boundary. Finally, variability in the precession band appears to be associated only with equatorial divergence and surface oceanographic processes.

Carbonate and Ice Volume

The relationships of the carbonate EOFs to ice volume over the past 4.5 m.y. were also examined. Previous studies (e.g., Farrell and Prell, 1991) suggested that the relationship between carbonate and ice volume over the late Pleistocene was not present during the Pliocene and early Pleistocene. A comparison between benthic δ18O and carbonate concentration EOFs may clarify this relationship. Comparisons were made between a spliced 849/846 δ18O record (Mix et al., this volume; Shackleton et al., this volume) and CaCO3 EOFs 1 and 2 for each 1-m.y. time slice. (Comparisons having similar results were also made between δ18O and CaCO3 MAR EOFs, but are not shown here.) Overall, δ18O and CaCO3 EOFs 1 and 2, respectively, are not highly correlated in any 1-m.y. interval, having correlation coefficients no higher than 0.23. Heavy isotopes (glacial) are correlated with positive EOF loadings (higher carbonate) for EOF-1, and with negative EOF loadings for EOF-2.

Although overall correlation is low, correlation is high in some frequency bands. Significant coherence between δ18O and the carbonates in the tilt band, variability appears to reflect surface oceanographic processes operating on the equator and near 6°N, near the NECC boundary. Finally, variability in the precession band appears to be associated only with equatorial divergence and surface oceanographic processes.

Table 7. Summary of cross-spectral results at Site 846/849.

<table>
<thead>
<tr>
<th>Period (k.y.)</th>
<th>Coh.</th>
<th>Phase (degrees)</th>
<th>Period (k.y.)</th>
<th>Coh.</th>
<th>Phase</th>
</tr>
</thead>
<tbody>
<tr>
<td>0-1 Ma</td>
<td>EOF-1</td>
<td></td>
<td>100</td>
<td>0.87</td>
<td>59 ± 17</td>
</tr>
<tr>
<td></td>
<td>EOF-2</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>71</td>
<td>0.87</td>
<td>9 ± 17</td>
<td>45</td>
<td>0.76</td>
<td>37 ± 26</td>
</tr>
<tr>
<td>41</td>
<td>0.88</td>
<td>107 ± 16</td>
<td>40</td>
<td>0.90</td>
<td>84 ± 16</td>
</tr>
<tr>
<td>23</td>
<td>0.83</td>
<td>149 ± 20</td>
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<td>0.75</td>
<td>73 ± 24</td>
</tr>
<tr>
<td>16</td>
<td>0.77</td>
<td>183 ± 24</td>
<td>13</td>
<td>0.82</td>
<td>143 ± 21</td>
</tr>
<tr>
<td>1-2 Ma</td>
<td>EOF-1</td>
<td></td>
<td>111</td>
<td>0.91</td>
<td>57 ± 14</td>
</tr>
<tr>
<td></td>
<td>EOF-2</td>
<td></td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>77</td>
<td>0.86</td>
<td>54 ± 18</td>
<td>111</td>
<td>0.94</td>
<td>61 ± 11</td>
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<tr>
<td>43</td>
<td>0.73</td>
<td>156 ± 27</td>
<td>83</td>
<td>0.76</td>
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<tr>
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<td>0.83</td>
<td>0 ± 22</td>
<td>17</td>
<td>0.91</td>
<td>129 ± 28</td>
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<td>0.82</td>
<td>15 ± 21</td>
</tr>
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<td>12</td>
<td>0.89</td>
<td>10 ± 16</td>
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<tr>
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<td></td>
<td></td>
<td>10</td>
<td>0.90</td>
<td>33 ± 15</td>
</tr>
<tr>
<td>2-3 Ma</td>
<td>EOF-1</td>
<td></td>
<td>41</td>
<td>0.87</td>
<td>126 ± 17</td>
</tr>
<tr>
<td></td>
<td>EOF-2</td>
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<td></td>
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<td></td>
</tr>
<tr>
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<td>125</td>
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<td>20 ± 18</td>
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<td>44 ± 11</td>
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<td>-60 ± 14</td>
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<td>-161 ± 15</td>
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<td>EOF-2</td>
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<tr>
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<td>0.90</td>
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<td>0.84</td>
<td>169 ± 20</td>
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<td>0.84</td>
<td>69 ± 19</td>
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<td>0.77</td>
<td>-47 ± 24</td>
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<td>11</td>
<td>0.92</td>
<td>72 ± 13</td>
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<td>333</td>
<td>0.90</td>
<td>-160 ± 12</td>
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<tr>
<td></td>
<td>EOF-2</td>
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<td></td>
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<td>125 ± 22</td>
<td>14</td>
<td>0.81</td>
<td>-57 ± 18</td>
</tr>
</tbody>
</table>

Table 7: Summary of cross-spectral results at Site 846/849.

$δ^{18}$O data from Shackleton et al. (this volume; Site 846) and Mix et al. (this volume; Site 849).
VARIABILITY OF LATE NEOGENE PACIFIC CARBONATE

Figure 21. A-E. Power and coherence spectra for oxygen isotopes and δCaCO₃ EOF-1, for the 1-m.y. time slice amplitude time series given in Figure 13. Solid line indicates the EOF power spectra, dotted line is the insolation power spectrum. F-J. Power and coherence spectra for oxygen isotopes and EOF-2 for the amplitude time series given in Figure 13. Symbols are as described in the caption for Figure 10.

Implications

Taken at face value, the EOF patterns resolved above indicate a strong influence by surface ocean processes on eastern equatorial Pacific carbonate sedimentation. In every analysis, the equator is the focus of most of the variability in EOF-1, suggesting a relation to equatorial divergence and, thus, carbonate productivity. In the CaCO₃ concentration EOFs alone, the possibility of opal dilution cannot be ruled out. However, the results from EOF analysis of carbonate MAR time series exclude the possibility that the equatorial signal resolved in EOF-1 is mainly associated with noncarbonate dilution. EOF-1 includes variability that is highly coherent with insolation variations. Frequency domain analyses suggest a coupling to surface ocean processes at a range of frequencies. Regional variability is stronger in the tilt and precession bands than at low frequencies.

One independent line of evidence supports changes in carbonate productivity as the primary factor controlling carbonate sedimentation. Radiolarian faunal data from intervals containing high and low carbonate concentrations, respectively, were studied by Pisias et al. (this volume). Radiolarians associated with increased productivity and stronger equatorial divergence correlate with increased CaCO₃, while radiolarians associated with warmer, tropical waters correlate with increased CaCO₃. As carbonate concentration and carbonate mass flux are positively correlated, this indicates that increased opaline productivity coincides with increased carbonate production in the equatorial Pacific, and vice versa. Lyle et al. (1988) arrived at similar conclusions in a late Pleistocene study of equatorial Pacific sedimentation. They postulated that oscillations in production of two planktonic communities (one community of both opal and calcite-secreting plankton) and one community of opal-secreting plankton, and, finally, radiolarians were shown to have derived mainly from changes in carbonate flux out of the water column, rather than from carbonate dissolution.

An important result of this study is the constancy of the dominant mode of carbonate sedimentation and the apparent relationship to surface processes in the presence of other large-scale boundary condition changes. A similar result was found by Pisias et al. (this volume), who determined that fundamental relationships between
carbonate saturation state and associated carbonate dissolution be responsible for the variability resolved above? Although the EOF analyses strongly suggest that surface ocean processes are primary, is a dissolution signal embedded? To answer this, carbonate dissolution related to changes in the chemistry of the deep ocean must be considered, as well as dissolution related to regional organic carbon oxidation, a byproduct of increased productivity. Each of these dissolution processes is discussed below. In addition, comparison to other dissolution indicators is necessary.

If dissolution caused by global oceanic chemistry changes drives equatorial Pacific CaCO$_3$ sedimentation, a pattern of variability similar to modern productivity gradients would not be expected. The water depths of five of the sites studied above (848–852; Table 1) fall within a 100-m range. If dissolution were the dominant process and all other factors were held constant, these five sites would respond similarly to changes in corrosiveness of the overlying waters, and they would have approximately the same variance. However, variability over these sites is significant on time scales up to 20 k.y. Archer’s model (Archer, 1991a) predicted that a very large gradient in carbonate saturation would be required over a depth interval of 200 m to account for central equatorial Pacific CaCO$_3$ patterns by changes in ocean chemistry. This is unlikely given the relative homogeneity of the present-day gradient (Archer, 1991a). It is more unlikely that large corrosivity changes on precessional time scales would occur over an even smaller depth range.

If dissolution owing to global ocean chemistry variation is dominant, a correlation between water depth and carbonate sedimentation would be present. Because all of the sites are within the upper lysocline, records at deeper water depths would be expected to have more variability than the shallowest sites. The transects of EOF loadings over depth (Figs. 12B, 16B, 18B, 18D, 18F) indicate that the opposite is true for EOF-1. Sites 846 and 847, which are the shallowest sites, have on average higher loadings in EOF-1 for CaCO$_3$ concentration and MAR than Sites 848 through 853, which are more than 400 m deeper. In some cases, such as the low-frequency band EOF loadings for CaCO$_3$, no relationship with depth can be determined. The only instance where a positive correlation between EOF loadings and depth is seen is in CaCO$_3$ EOF-2. Carbonate dissolution must be considered as a possible dominant factor determining the spatial pattern of EOF-2, and as one of several factors controlling EOF-1 of the low frequency band.

Variability in carbonate sedimentation at low frequencies may be related to carbonate dissolution. Evidence lies in the slightly lower gradients with latitude resolved in the low frequency EOFs relative to the tilt band and precession band EOFs, and the low spatial variability from 0 to 1, 1 to 2, and 4 to 5 Ma (Figs. 17, 19, and 20). If true, influence from dissolution is embedded in the spatial variability of EOF-1 of CaCO$_3$, because low frequency variance is dominant in all of the CaCO$_3$ records, and EOF-1 accounts for more than 50% of the total variance. Additional evidence comes from comparison of EOF-1 with benthic $^{87}$Sr, where at low frequencies, variations in ice volume and EOF-1 are coherent (Fig. 21). However, with the exception of the time intervals from 3 to 4 Ma and 5 to 6 Ma, the low-pass filtered CaCO$_3$ EOFs show a correlation with latitude, and the highest loadings are focused on the equator. Thus, even at low frequencies, surface oceanographic processes exert influence on carbonate sedimentation variability.

If dissolution is a primary component affecting all Leg 138 sites to the same degree, an attenuation as a function of carbonate sedimentation rate can be expected. If all other factors affecting carbonate sedimentation are held constant except for carbonate sedimentation rates, higher sedimentation rate sites will show less of a response to the same amount of changes in dissolution. This effect was modeled by Pisas and Prell (1985), using carbonate records from DSDP Leg 85 sites. At the Leg 138 sites, the maximum carbonate accumulation rates are at the equatorial sites, and the minimum carbonate accumulation rates are located away from the equator (Sites 848, 852, and 853). According to the Pisas and Prell model, dissolution should exert a stronger control on the variability in Sites 848, 852, and 853 relative to equatorial Sites 849, 850, and 851. Although inconsistent with EOF-1, such a pattern is consistent with EOF-2 on the western transect. This is another line of evidence supporting EOF-2 as an indicator of carbonate dissolution.

Another line of evidence that argues against dissolution as a primary factor driving carbonate fluctuations at Leg 138 sites is a comparison to dissolution indexes. If dissolution is significant, a correlation between carbonate variability and dissolution indexes might be evident. Dissolution indexes, as estimated from foraminiferal fragmentation data, have been generated for Site 847 (Murray et al., this volume) and Site 846 (Le et al., this volume). Sites 846 and 847 have very high loadings in EOFs 1 and 2 for CaCO$_3$ concentration and MAR, but comparison of these records to fragmentation data shows little relationship. In some instances, carbonate and fragmentation indexes appear to be positively correlated, and in other instances, a negative correlation is evident. The correlation coefficients between the fragmentation records and carbonate concentration, carbonate MAR, and carbonate EOFs, respectively, are close to zero. Because of other evidence suggesting that EOF-2 of carbonate concentration reflects dissolution processes, this result may indicate that fragmentation indexes are not the best indicators of carbonate loss resulting from dissolution.

How important is the role of carbonate dissolution driven by organic carbon respiration? Equatorial sites located in regions of higher productivity would lie under waters more undersaturated in CO$_3$$_2$ than sites off of the equator, where production is lower. Accordingly, lower CO$_3$$_2$ concentration is documented beneath the equator in the eastern Pacific (Broecker and Peng, 1982, p.77). If this type of dissolution is a primary control on carbonate sedimentation, then any variability could be expected to be coupled to equatorial divergence, where productivity is highest. Archer (1991b) modeled the effect of variations of organic carbon degradation on lysocline shape. For the last glacial maximum, Archer (1991a) determined that the increase in carbonate accumulation observed by Farrell and Prell (1989) in the central equatorial Pacific in theory can be generated by decreasing the ratio of organic carbon to calcite and, thereby, diminishing respiration-driven dissolution. However, organic carbon accumulation in the equatorial Pacific coincides with higher carbonate accumulation during the Pleistocene (Doose et al., this volume; Lyle et al., 1988; Pedersen, 1983). By extension to the late Neogene, this rules out organic carbon respiration as a single dominating factor. However, modeling the effect of changes in the ratio of organic carbon to calcite flux will help to determine the role of this process in determining the spatial patterns resolved here.

If both global deep ocean chemistry changes and organic carbon-induced dissolution are considered together, then it becomes clear why resolving a single dissolution mode is difficult. One dissolution process would be coupled to water depth, and the other process would
be coupled to the surface productivity, all other factors held constant. In addition, while high productivity may lead to increased dissolution in one region of the equatorial Pacific, the increase in rain rate of material could lead to increased preservation in another region. As noted by Archer (1991b), the competition of these two effects may make resolution of either one of them difficult, especially if they are operating on the same time scales.

As noted previously, the spatial distribution of CaCO₃, EOF-2 may suggest a relationship to these multiple carbonate dissolution processes. For instance, if high oceanic productivity at the eastern boundary is associated with high loadings of EOF-2 on the eastern transect, this may indicate high dissolution induced by organic carbon. The same high productivity might lead to decreased dissolution at off-equator sites on the 110°W transect because of increased chances for preservation of carbonate. This might explain the out-of-phase relationship between eastern and western sites. Although amplitude time series of EOF-2 are not correlated to foraminiferal fragmentation data, there is evidence for a correlation with water depth (Fig. 12D). On the 110°W transect, the loadings of EOF-2 are consistent with influence from dissolution. However, the absence of a relationship to the fragmentation indexes remains puzzling.

One can make a similar argument regarding the role of noncarbonate material in controlling the spatial pattern of CaCO₃, EOF-2. Because the spatial pattern of EOF-2 in CaCO₃ concentration is different from the spatial pattern of EOF-2 of carbonate MAR, the concentration data must record a different process. The spatial pattern of CaCO₃ concentration EOF-2 might be indicating dilution related to opal production in the eastern transect sites, and also carbonate dissolution in the off-equator sites on the western transect. This possibility can be tested with records of opal and noncarbonate, non-opal accumulation and with opaline faunal counts, such as that of Piasas et al. (this volume).

The patterns resolved in EOF-2 of carbonate concentration may indicate a change in oceanographic boundary conditions near 3 to 4 Ma. A stepwise change in the amplitude time series of EOF-2 (Fig. 9) occurs at this time, and a change in the spatial pattern of EOF-2 in the time slices occurs between 3 to 4 Ma and 4 to 5 Ma. These changes may be associated with closure of the Panamanian isthmus. A modeling study (Maiher-Raimer et al., 1990) suggested that the closure of the Isthmus of Panama at about 3 to 4 Ma was related to significant changes in the intensity of deep ocean circulation and, thereby, deep ocean chemistry. The spatial pattern of EOF-2 may indicate a response of carbonate sedimentation consistent with this boundary condition change.

CONCLUSIONS

1. Spatial patterns in sediment composition in the eastern equatorial Pacific exhibit variability common to all of Leg 138 Sites 846 through 853, as well as variability that is associated with latitude. Large spatial variability is present over the past 6 m.y. on time scales from 1 m.y. to less than 10 k.y. The spatial patterns are remarkably consistent over the past 6 m.y. Although differences between individual time slices are evident, the fundamental spatial pattern that represents the bulk of the variability in the carbonate system is consistent over time periods in which large boundary condition changes take place. This implies a constant forcing mechanism of variability in carbonate preservation over the past 6 m.y.

2. Two independent modes of variability account for more than 70% of the variance in carbonate concentration and carbonate accumulation records over the past 6 m.y. The first mode, EOF-1, is coupled to the equator and is coherent with insolation changes. This pattern of variability appears to be associated with surface ocean processes. The second mode of variability in carbonate sedimentation may be associated with carbonate dissolution over the past 4 m.y. Coherence between carbonate EOF-2 and insolation is low.

3. Variability in carbonate sedimentation differs as a function of frequency, but in the precession band, the tilt band, and the low frequency band, high regional variability allows one to infer a coupling to surface ocean processes. Precession band variability displays the highest coupling to the equator, tilt band variability indicates stronger importance of Site 852 at 6°N, and variability in carbonate sedimentation at low frequencies displays the least coupling to equatorial processes.

4. A correlation of the loadings of EOF-1 of carbonate concentration and mass accumulation with latitude implies a strong linkage of carbonate sedimentation to surface ocean processes. The absence of a correlation between EOF-1 loadings and depth suggests that global alkalinity-induced dissolution is not dominant. The loadings of EOF-2 do indicate a relationship with water depth, suggesting that dissolution may play a role in forcing this mode of variability.

5. Benthic δ¹⁸O and carbonate are positively correlated throughout the past 4 m.y., with increased carbonate associated with positive δ¹⁸O. Over the past 4 m.y., the dominant mode of variability in carbonate concentration (EOF-1) is coherent with Site 849/846 δ¹⁸O at periods of 41, 23, and 19 k.y. The second independent mode of carbonate variability (EOF-2) is coherent with δ¹⁸O at 100 k.y. from 0 to 2 Ma. The two independent modes of variability observed in the carbonate system each have linear relationships to δ¹⁸O, but at different frequencies and/or phases. This implies linkages with variability in ice volume in both modes of carbonate variability.

6. Modeling of dissolution and dilution processes combined with EOF analyses and comparison of faunal and isotopic indicators from Leg 138 sites can provide additional constraints to the oceanographic interpretations from the EOFs presented here. Parameters indicating surface ocean processes, such as radiolarian and foraminiferal census data, will be a useful tool in determining the phase of the carbonate response with respect to insolation. Parameters that represent spatially coherent deep ocean changes, such as benthic isotopes, will constrain the spatial variance in the phase of the carbonate response.

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