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Late Neogene changes in diatom sedimentation in the North Pacific

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Abstract—During the late Neogene, North Pacific diatom sedimentation underwent major changes in response to high-latitude cooling and changes in surface and deep water circulation. At 9 Ma diatom mass accumulation rates (MARs) increased in the NW Pacific and off northeast Japan, possibly due to shoaling of the Isthmus of Panama, which lead to an enrichment of nutrients in North Pacific deep waters. During the latest Miocene, diatom MARs increased progressively off southem California (6.5 Ma), at high latitudes of the North Pacific (6.2 Ma), and off northeastern Japan (5.5 Ma), presumably in response to high latitude cooling. At about 4.5 Ma diatom sedimentation abruptly increased in the NW Pacific but declined off Japan and California, coincident with the onset of a prolonged period of high-latitude warmth. Enhanced upwelling of nutrient-rich deep waters in the NW Pacific probably stimulated diatom production there. A major step in high latitude cooling at 2.7 Ma caused a reversal of these mid Pliocene diatom sedimentation patterns. Upwelling of deep, nutrient-rich waters waned at higher latitudes, leading to a decline in diatom productivity; while wind-driven, coastal upwelling increased off southern California and stimulated diatom growth.

Introduction

Biogenic sediments in the modern Pacific Ocean are silica-rich in comparison with the carbonate-rich sediments that typify the Atlantic. This is largely because deep waters in the Pacific are relatively old, depleted in oxygen, and rich in the nutrients including phosphorus, nitrate, and silica that are necessary for diatom growth (Berger 1970). Increased concentrations of carbon dioxide in the "old" Pacific deep waters also mean that the calcium carbonate compensation depth (CCD) occurs at relatively shallower depths in the Pacific than it does in the Atlantic, resulting in greater widespread dissolution of carbonate on the seafloor. These Atlantic–Pacific differences are due to the present day isolation of the two ocean basins and to the fact that bottom waters are created in the North Atlantic (North Atlantic deep water—NADW), but not in the North Pacific (Broecker and Peng 1982).

Prior to the middle Miocene, however, deep water interchange occurred between the Caribbean and eastern equatorial Pacific across the Central American Seaway, so that bottom waters of the Pacific and Atlantic were more similar in character and there was less of a difference in the nature of biogenic sediments being deposited on the seafloor of the two ocean basins (Keller and Barron 1983; Woodruff and Savin 1989). Ocean general circulation modeling suggests that an open isthmus would have caused lower-salinity waters from the Pacific to dilute North Atlantic surface waters by $>1‰$, restricting the potentiality for thermohaline overturn in the North Atlantic and reducing the production of any deep waters in the North Atlantic to near zero (Maier-Reimer *et al.* 1990). Reduced production of deep waters in the North Atlantic would have hindered the global "conveyor belt" of deep-water circulation and lead to decreased nutrient levels in North Pacific deep waters (Broecker and Peng 1982; Maier-Reimer *et al.* 1990). Because upward diffusion of deep waters is presently responsible for renewing nutrients in near-surface waters north of 40°N by [see Fig. 3 of GEOSECS data in Craig *et al.* (1981)], it is therefore likely that the nutrient concentrations in surface waters and diatom production in the North Pacific may have been significantly lower than at present prior to the middle Miocene (Woodruff and Savin 1989; Maier-Reimer *et al.* 1990).

It therefore appears likely that the late Neogene history of diatom sedimentation in the North Pacific was strongly influenced by the strength of the deep-water "conveyor belt", and ultimately by the closing of the Central American Seaway.

Materials and methods

In 1992 Ocean Drilling Program (ODP) Leg 145 drilled 25 holes at seven sites in a west to east transect of the subarctic North Pacific (Fig. 1), providing excellent material for the study of the late Neogene history of diatom sedimentation in the North Pacific. Three of the Leg 145 sites (882, 883 and 887) were drilled atop seamounts, where thick sections of biogenic sediments, relatively free of terrigenous debris, were recovered. Thick Neogene sections were also recovered at Site 884, on the east flank of the Detroit Seamount, and at Site 881 on the ocean basin floor east of the Kuril Islands (Rea *et al.* 1993). Sites 885 and 886 in the low biologic productivity region of the central North

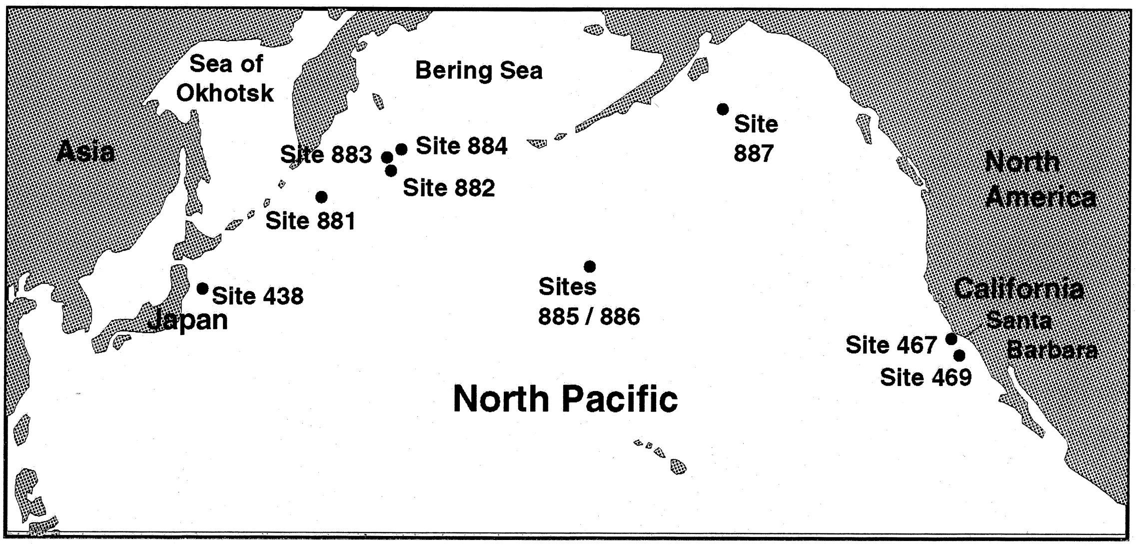

Fig. 1. Location of Deep Sea Drilling Project and Ocean Drilling Program sites studied in the North Pacific and the Santa Barbara coastal area off southern California.

Pacific cored only a thin section (50 m) of Neogene biosiliceous sediment. Leg 145 provided the first temporally lengthy Miocene magnetostratigraphic records in the North Pacific, extending back through magnetic polarity Subchron C5En (18.817–18.317 Ma) from sediments recovered at ODP Sites 884 (51°27'N, 1 68°20.2'E; water depth, 3836 m) and 887 (54°21.9'N, 1 48°26.8'W, water depth, 3645 m). Correlation of diatom biostratigraphy to these magnetostratigraphic records allows for the first time the accurate construction of age models for the last 18 million years (Barron *et al.* 1995).

Diatom mass accumulation rates (MARs) (g/cm^2 k.y.) for the Leg 145 sites were calculated by multiplying the percentage diatoms in the sediment by the dry bulk density of the sediment (g/cm^3) and then by the sedimentation rate (cm/k.y.). Percentage estimates of diatoms, as determined by shipboard scientists through the examination of smear slides (Rea *et al.* 1993), are not a weight percent measurement as is normally required in estimating mass accumulation rates. At best, such estimates represent a percent of volume that should be proportional to a weight percent estimate. Nevertheless, it can be argued [see Table 3 in Shipboard Scientific Party (1990)] that visual estimates of the abundance of biosilica generally fall within a factor of 2 of estimates of biosilica that are made by chemical means. Certainly, weight percent opal data would be preferable to use for MARs; however, such opal data has presently only been completed at Site 887 (Rea and Snoeckx 1995) and on the past 3.2 m.y. of Site 882 (Haug *et al.* 1995).

Due to the subjective nature of the smear slide data, it was decided to average data and the resulting MAR estimates for each core (~9.5 m length); meaning that about 2 to 7 smear slide percent estimates were averaged per core. Dry bulk densities were estimated by the Gamma Ray Attenuation Porosity Evaluator (GRAPE) by shipboard scientists (Rea *et al.* 1993) and were also averaged per core. The sedimentation rate was taken from the magnetostratigraphic and biostratigraphic compilations of Barron *et al.* (1995), and the geomagnetic time scale of Cande and Kent (1992) has been applied throughout this report.

Results

Figure 2 shows diatom mass accumulation rates (MARs) for the Oligocene to Holocene at North Pacific ODP Sites 883, 884 and 887. Although poorly preserved diatoms are recorded in the lowermost Oligocene of Site 884, diverse and moderately well preserved diatoms first occur in the upper part of the lower Oligocene (ca. 30 Ma) (Gladenkov and Barron 1995). Nevertheless, Fig. 2 shows that diatom MARs were quite low ($\ll$1 g/cm^2 k.y.) at Site 884 during the late Oligocene and early Miocene until about 18.5 Ma. Unfortunately, the upper Oligocene and most of the lower Miocene are removed at an unconformity at Site 883 (Barron *et al.* 1995), but relatively common diatoms are present immediately above this unconformity in lower Miocene sediments that are estimated to be about 21 Ma in age. At northeast Pacific Site 887 on the Patton-Murray Seamount, diatoms first appear at about 18 Ma, an age that approximates the age of base of the diatom-rich Monterey Formation of California and the onset of a widespread expansion of diatomaceous sediments in the North Pacific (Barron and Baldauf 1990). Western North Pacific Sites 883 and 884 generally have higher diatom MARs than northeast Pacific Site 887, except for the last 1 m.y. Site 883 on the top of Detroit Seamount (2407 m water depth) typically has higher MARs than Site 884 (3838 m water depth) on the flank of the seamount, especially for the past 10 m.y. Presumably this difference reflects winnowing of smaller, more delicate diatoms off the flank of the seamount.

It is apparent from Fig. 2 that the main changes in diatom MAR's occurred after 10 Ma. A major increase in diatom sedimentation is recorded at about 9 Ma (event A) at Sites 883 and 884 and is followed by a second increase between about 6.5 and 6.0 Ma (event B). The 6 Ma increase is also clearly present at Site

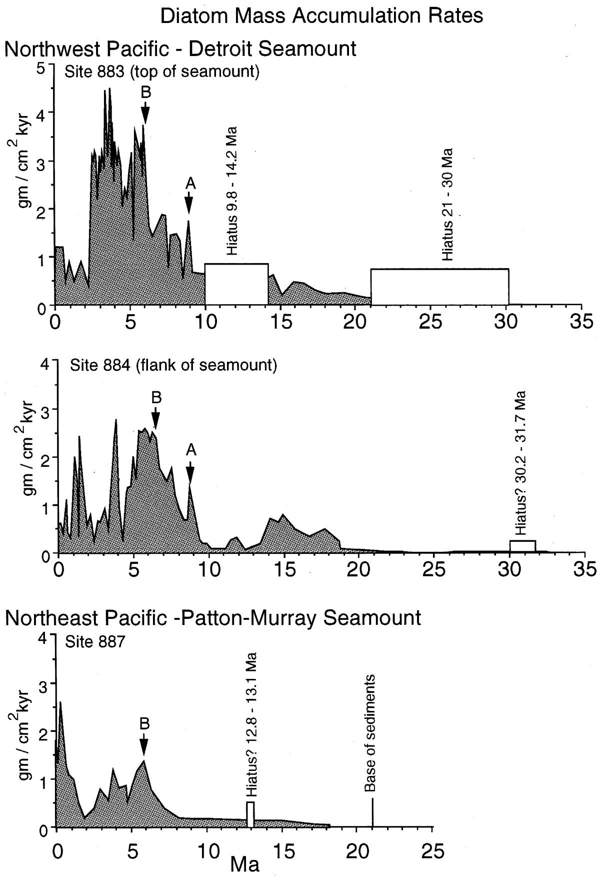

Fig. 2. Diatom mass accumulation rates for North Pacific Sites 883, 884 and 887 for the past 35 million years.

887, while the earlier 9 Ma increase is not evident there.

Let us examine in more detail the late Neogene history of diatom sedimentation in the North Pacific region cored by Leg 145. On Fig. 3 diatom MARs for the past 10 m.y. are compared between Detroit Seamount Sites 883 and 882 (water depth of 3255 m, about 90 km to the south of Site 883), Patton-Murray Seamount Site 887, and seafloor Site 881. The Site 884 record has been omitted, because diatoms there are much more poorly preserved and fragmented than they are at nearby Sites 882 and 883. Similarly, the records for Sites 885 and 886 in the low biologic productivity region of the central North Pacific are not included, because only a thin section (50 m) of Neogene biosiliceous sediment was recovered there. Sedimentation rates are also shown on Fig. 3 for comparison with the MARs.

Increased diatom accumulation rates are recorded at Site 883 between 6.2 and 2.7 Ma. At 6.2 Ma (event B), diatom MARs increase abruptly from <2 g/cm^2 k.y. to >3 g/cm^2 k.y.; they mostly remain between 2.5 g/cm^2 k.y. and 4 g/cm^2 k.y. until 2.7 Ma, where they sharply decline to values <1 g/cm^2 wk.y. (event D). In general diatom MARs parallel sedimentation rates, although an increase in sedimentation rate at 4.4 Ma (event C) from ca. 80 m/m.y. to 130 m/m.y. does not appear to be accompanied by a major increase in diatom MAR at Site 883. At nearby Site 882, however, diatom MARs increase significantly (from ca. 2 to 4–6 g/cm^2 k.y.) at 4.4 Ma, corresponding to an increase in the sedimentation rate from 66 m/m.y. to 140 m/m.y. The differences in diatom MARs between Sites 882 and 883 may reflect inaccuracies of estimating diatoms by smear slides.

The records at both Sites 882 and 883 show a major decline in diatom MARs and sedimentation rate at about 2.7 Ma. Detailed laboratory determinations by Haug *et al.* (1995) of opal for the past 3.2 m.y. at Site 882 reveal that a sharp decrease in opal MAR from >3 g/cm^2 k.y. to <1 g/cm^2 k.y. at 2.73 Ma coincided with a major increase in ice rafted detritus, signaling the onset of extensive Northern Hemisphere glaciation.

Site 887 has lower diatom MARs (generally ~1 g/cm^2 k.y.) than Sites 882 and 883, although comparable or higher diatom MARs characterize the last 0.5 m.y. As at Site 883, diatom MAR's increase at about 6.2 Ma (event B), but the increase is more gradual at Site 887. The abrupt increase in sedimentation rate from ca. 10 to >40 m/m.y. at 6.2 Ma, however, suggests that the gradual nature of this MAR increase may be due to the practice of averaging MAR values per core—one core (or about 9.5 m) is equivalent to about 1 m.y. duration prior to 6.2 Ma. Correspondingly, MAR's decline from about 1 g/cm^2 k.y. to <0.5 g/cm^2 k.y. between about 3.1 and 2.6 Ma, coincident with a decrease in sedimentation rate from 30 m/m.y. to 15 m/m.y. that is recorded at 2.7 Ma (event D).

Site 881, on the floor of the ocean basin, also shows increased diatom MARs (to >2 g/cm^2 k.y.) at 4.4 Ma (event C). Poor recovery of sediments older than 5 Ma limits interpretation of the diatom MAR. In contrast to the Detroit Seamount sites, diatom MARs increase abruptly (to >4 g/cm^2 k.y.) at about 2.6 Ma (event D), coinciding with an increase in sedimentation rate from 33 to 100 m/m.y. and a major increase in ice rafted detritus (Shipboard Scientific Party 1993). After 2.0 Ma, the sedimentation rate drops to about 66 m/m.y., and diatom MAR shows considerable fluctuation but is generally <2 g/cm^2 k.y. Seismic stratigraphy (Hamilton 1995) reveals that the section deposited between 2.6 and 2.0 Ma (165 and 113 m below seafloor) at Site 881 represents localized channel fill and is not regionally representative. The abundant presence of fragmented, robust diatoms in these sediments also argues for redeposition by bottom currents.

Comparison with other parts of the North Pacific

It would be instructive to make some preliminary comparisons of the Leg 145 diatom accumulation rates with those from middle latitude regions of the North Pacific.

Site 438 off northeast Japan

Deep Sea Drilling Project Site 438 (40°37.79' N, 143°14.15' E, water depth 1558 m) (Fig. 1) is a key reference section on the continental slope off northeast Japan that contains a thick sequence of Pliocene and Miocene diatomaceous sediments (Shipboard Scientific Party 1980). Detailed diatom biostratigraphic data compiled by Barron (1980) and Akiba (1986) allow the construction of an age model for Site 438 using the

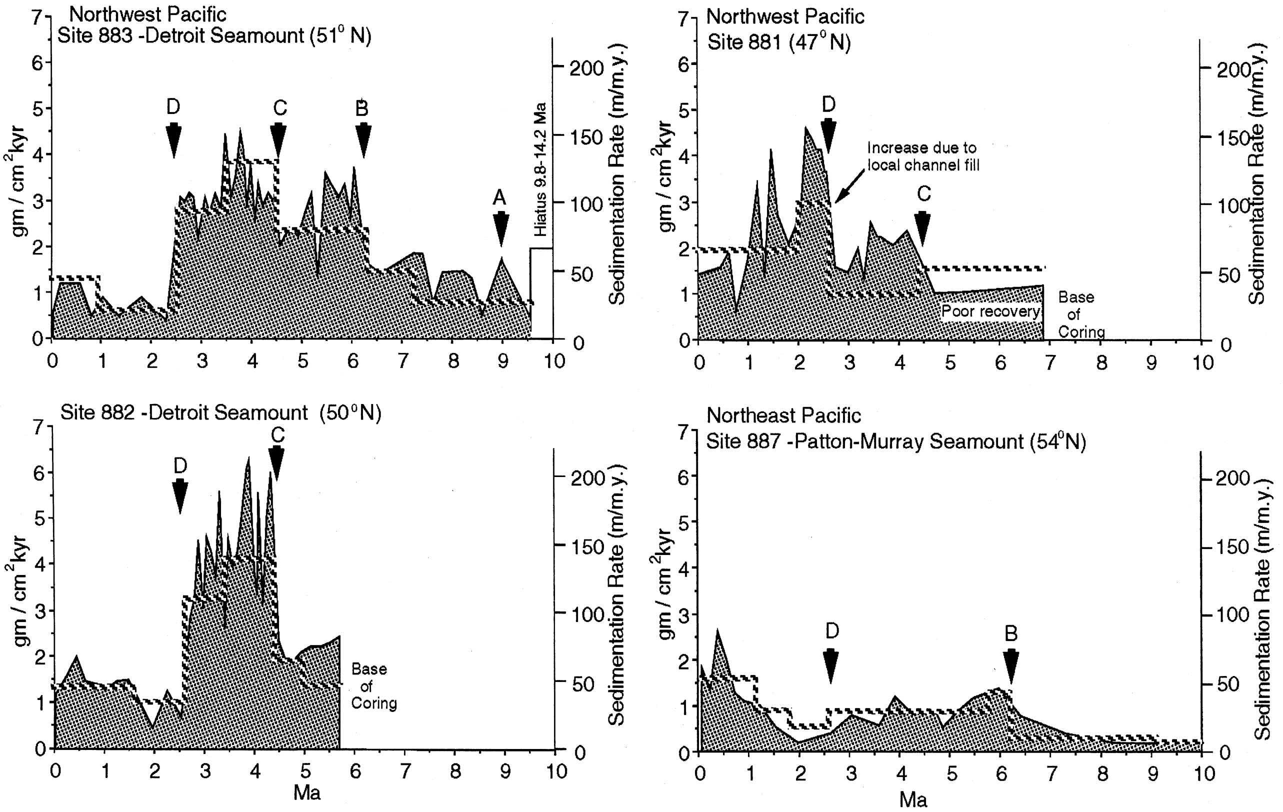

Fig. 3. Diatom mass accumulation rates for the past 10 million years compared with the sedimention rates (diagonally striped line) at Sites 881, 882, 883 and 887.

updated diatom datum level ages of Barron and Gladenkov (1995); additional microfossil data for Site 438 (Lazarus *et al.* 1995) provide further age control.

Application of this age model to shipboard data on percent diatoms and sediment density allows the calculation of the sedimentation rate and diatom MARs for the past 10 m.y. at Site 438 (Fig. 4). Smear slide estimates of the percent diatoms and GRAPE measurements of the wet bulk density have been taken from Shipboard Scientific Party (1980). Dry bulk density was estimated from GRAPE wet bulk density by using the formula of Haug *et al.* (1995).

Diatom MARs at Site 438 are generally low (< 1.5 g/cm^2 k.y.) with the exception of two intervals of markedly increased values. Between about 5.7 and 4.5 Ma (event C), MAR roughly doubles to ca. 3 g/cm^2 k.y., whereas MARs of between 3 and 4 g/cm^2 k y. are suggested between 3.1 and 2.6 Ma (event D). Both of these MAR increases are associated with major increases in the sedimentation rate. The intervening 4.5–3.1 Ma interval of decreased diatom MAR approximates the interval of greatest diatom MARs and fastest sedimentation rates at Detroit Seamount Site 882 (Fig. 3). Minor increases in sedimentation rate are observable at 9.0 Ma (event A) and 6.2 Ma (event B), but diatom MAR does not change significantly at these times.

The sedimentation rate at Site 438 declines sharply at 2.7 Ma from over 200 m/m.y. to about 20 m/m.y., resulting in a greater than three-fold decrease in diatom MAR (from ca. 3 to < 1 g/cm^2 k.y.) (Fig. 4). Although it is tempting to correlate this major MAR decrease with 2.7 Ma (event D) decreases at Sites 882, 883 and 887 (Fig. 3), Arthur *et al.* (1980) argue that regional uplift at Site 438 during the latest Pliocene and Pleistocene shifted the main locus of hemipelagic deposition seaward and was responsible for this sedimentation change.

California

On the bottom part of Fig. 4, estimates of biogenic silica MAR for the Monterey and Sisquoc Formations of the Santa Barbara coastal area of southern California as determined by Isaacs (1983, 1985) have been plotted for comparison with the Site 438 record (the ages and biochronology on the Monterey and Sisquoc Formations have been updated). Various studies including the author's observations show that practically all of the biogenic silica in Monterey rocks was contributed by diatoms, so biogenic silica MAR is essentially equivalent to diatom MAR. Note the MAR scale has been doubled over that shown in the top half of Fig. 4.

The MAR of biosilica in the upper Monterey Formation undergoes stepwise increases at about 8.5 and 7.6 Ma, corresponding, respectively, to the bases of Isaacs' upper calcareous–siliceous and clayey–siliceous members (Isaacs 1983). The precise timing of these increases is rather uncertain, because Isaacs (1983) averaged her MAR data for the stratigraphic units as a whole. Maximum biosilica MARs in the upper Monterey Formation (ca. 6 g/cm^2 k.y.) are comparable with the maximum diatom MARs that occur later in the middle part of the Pliocene between 4.4 and 2.7 Ma at Detroit Seamount Site 882 (Fig. 3). However, the overlying Sisquoc Formation, which ranges in age from 6.5 Ma at its base to about 4.0 Ma

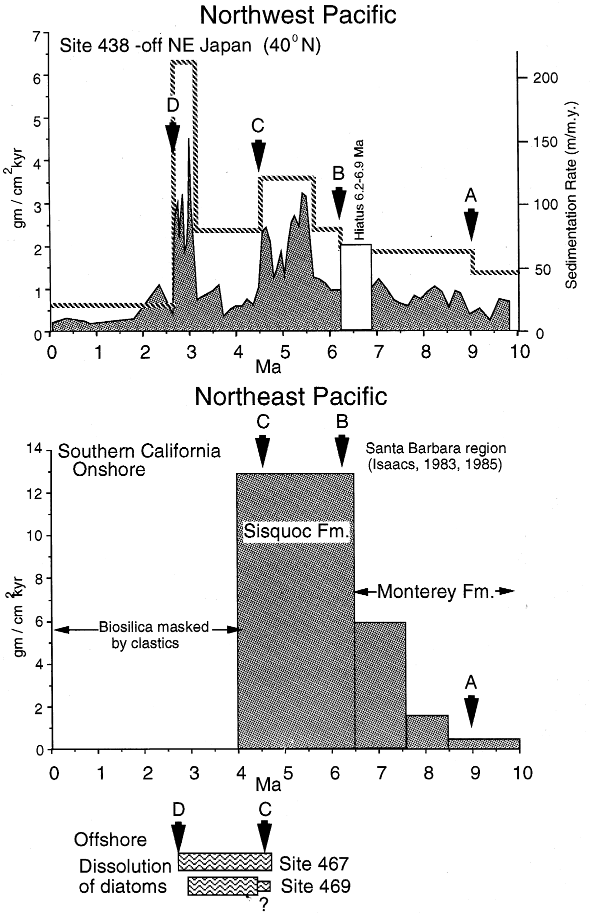

Fig. 4. Diatom mass accumulation rates at Site 438 off northeast Japan (upper part) compared with Isaacs' (Isaacs, 1983, 1985) estimates biosilica accumulation rates in the Monterey and Sisquoc formations in the Santa Barbara area of southern California for the past 10 million years. Sedimentation rate at Site 438 is shown by the diagonally striped line.

at its top (Dumont and Barron 1995), has an average biosilica MAR that is considerably higher (ca. 13 g/cm^2 k.y.) than those of the Monterey. Above the Sisquoc Formation, biosilica is almost totally masked by clastic debris in coastal California rocks and was not measured by Isaacs (Isaacs, 1983, 1985). Examination by the author of numerous smear slides of these post-Sisquoc sediments during the past 25 years has shown them to be very poor in diatoms.

The onset of the Sisquoc maximum in biosilica MAR slightly precedes a major increase in diatom MAR (event B) at Detroit Seamount Site 883 and Patton-Murray Seamount Site 887 (Fig. 3). The top of the Sisquoc Formation at about 4.0 Ma approximates an increase in diatom MAR at Site 887, but it postdates the major increase in diatom MAR that occurred at Site 882 at 4.4 Ma (event C) (Fig. 3). It should be pointed out, however, that the Sisquoc Formation is quite variable in thickness and its top may be locally as old as 5.0 Ma (Barron 1992). At offshore DSDP Sites 467 (33°50.97′N, 120°45.47′W, water depth 2128 m) and 469 (32°37.00′N, 120°32.90′W, water depth 3790 m)(Fig. 1), which are more isolated from the influx of clastic material, sediments deposited between ca. 4.5 Ma (event C) and 2.7 Ma (event D) contain rare, poorly preserved diatoms, suggesting a major decline in diatom productivity occurred during the middle part of the Pliocene (Barron, 1981 1992) (Fig. 4).

Discussion

It may be tempting to follow Barron and Baldauf (1990) and attribute late Neogene changes in diatom sedimentation in the North Pacific to high latitude cooling or warming and its effect on upwelling and diatom productivity. It is worth noting, however, that during the late Miocene and Pliocene, intermediate and surface water flow across the Central American Seaway were progressively cut off (Keller *et al.* 1989; Farrell *et al.* 1995), most likely leading to enhanced basin–basin fractionation between the North Pacific and the North Atlantic and increased nutrient availability in the North Pacific. It is therefore appropriate to discuss each of the four late Neogene events (A–D) that signal major change in North Pacific diatom sedimentation (Figs 2–4) both in terms of long-term changes in paleoclimatology and basin–basin fractionation. Because the closure of the Central American Seaway would result in enhanced basin–basin fractionation (Broecker and Peng 1982; Maier-Reimer *et al.* 1990), it would also be worthwhile to review important paleoceanographic changes that have been linked to the shoaling of the sill across the Isthmus of Panama. Figure 5 compares the late Neogene changes in diatom sedimentation (events A–D on Figs 2–4) with a spliced record of high latitude cooling as revealed by benthic foraminiferal oxygen isotope data from DSDP Sites 588 and 590 (Kennett 1986) and to paleoceanographic events that are tabulated by Farrell *et al.* (1995) which signal the progressive closing of the Central American Seaway. This isotope curve was used, because it is directly tied to paleomagnetic stratigraphy back through the entire late Miocene (Barton and Bloemendal 1986).

Event A

Was the 9 Ma (event A) increase in diatom MAR that is observed at Sites 883, 884 and 438 caused by high latitude cooling or in part by the uplift of Central America and intensification of basin–basin fractionation between the North Pacific and the North Atlantic? According to the benthic foraminiferal oxygen isotope curve of Sites 588–590, this event coincides with a negative shift or high latitude warming, rather than a positive shift (Fig. 5). The event postdates the Mi7 cooling event of Wright *et al.* (1992), which is correlated with the lowermost part of magnetic polarity Chron C4A, by about 0.5 m.y. In terms of diatom zones, event A falls at the boundary between the *Denticulopsis dimorpha* Range Zone and the overlying *D. katayamae* Partial Range Zone, an interval characterized by a warming trend in diatom and planktonic foraminiferal assemblages according to Barron and Keller (1983). Thus, it is not entirely clear that the event A increase was due to high latitude cooling.

On the other hand, beginning at about 8.8 Ma (age updated) the $\delta^{13}C$ of North Atlantic and Pacific deep waters began a steady divergence which Wright *et al.*

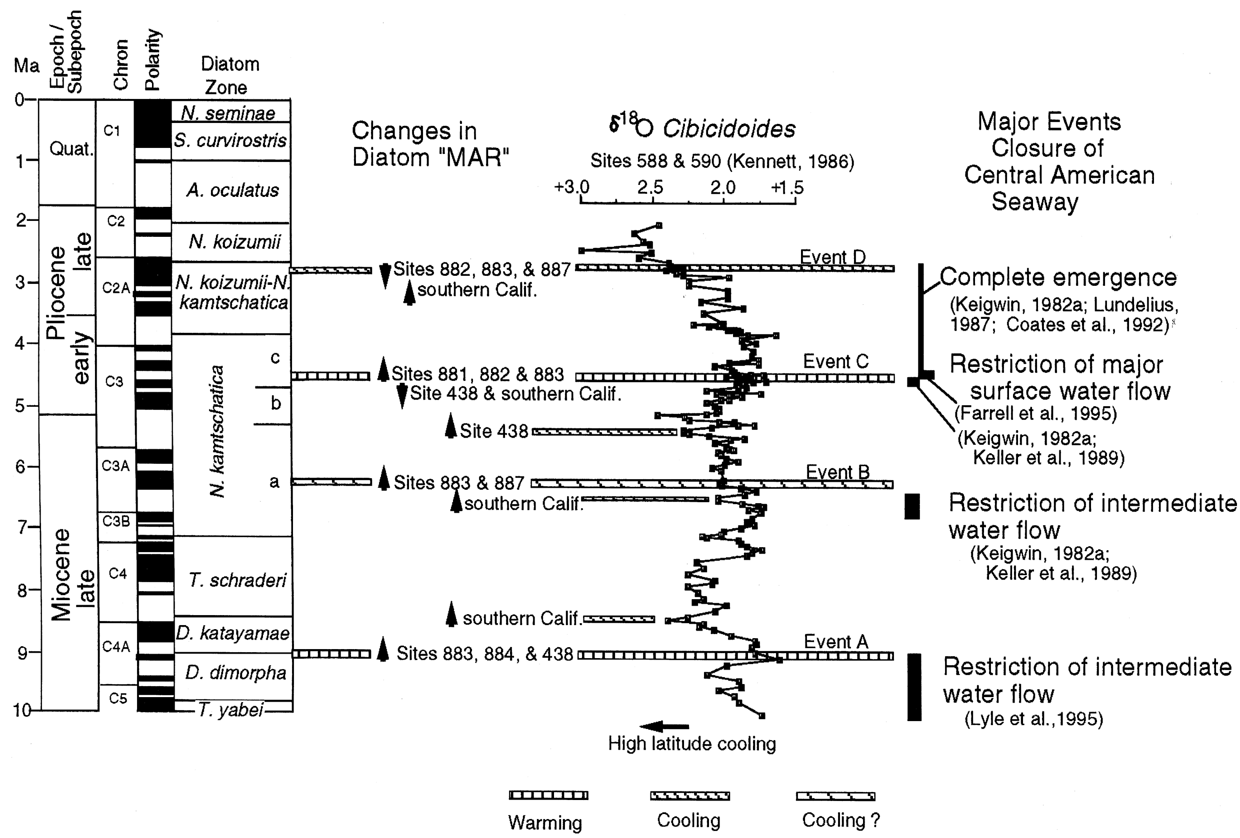

Fig. 5. Comparison of major late Neogene changes in North Pacific diatom mass accumulation rates (MAR) with the spliced oxygen isotope record from Sites 588 and 590 and with major events in the emergence of the Isthmus of Panama. The diatom zonation is that of Akiba (1986) and Barron and Gladenkov (1995) and the geomagnetic polarity time scale is that of Cande and Kent (1992).

(1991) attributed to an increase in the production of Northern Component Water (the ancestor to NADW). This would have lead to increased basin–basin fractionation and enhanced nutrient levels in North Pacific deep waters, which may have, in turn, lead to increased diatom productivity in surface waters of the North Pacific. Between 10 and 9 Ma a widespread carbonate dissolution event occurred throughout the equatorial Pacific (the "carbonate crash"), which Lyle *et al.* (1995) attributed to the cessation of the flow of less corrosive Caribbean intermediate and deep water into the Pacific, caused by the shoaling of the sill across the Central American Seaway. Some interchange of North and South American mammals began between 9.9 and 8.7 Ma, suggesting the presence of a chain of closely spaced islands in the Central America region according to Marshall (1988). It would thus seem possible that the increased diatom productivity in the North Pacific and the equatorial Pacific at 9 Ma (event A) was triggered by partial closure of the Central American Seaway.

Event B

Event B at 6.2 Ma marks a major increase in diatom MARs at the higher latitudes of the North Pacific (Figs 3 and 4), but it appears to postdate by 300 k.y. a major high latitude cooling event at about 6.5 Ma according to the Site 588–590 benthic foraminiferal oxygen isotope curve (Fig. 5). Increases in diatom MARs in southern California (6.5 Ma) and at Site 438 off northeast Japan (5.5 Ma) (Fig. 4) appear to be more closely tied to positive shifts in the Site 588–590 isotope curve (Fig. 5), suggesting that high latitude cooling may have stimulated upwelling and diatom productivity at these middle latitude locations. Hodell *et al.* (1994) recently compiled a high resolution benthic foraminiferal oxygen isotope curve for the uppermost and lowermost Miocene of Morocco which displays the same trend as the Site 588–590 curve. Event B (6.2 Ma) again lies in an interval of fluctuating (at 40 k.y. intervals) isotope values, falling between major oxygen isotope increases that are dated at 6.8 (stage C3Ar.180.12) and 5.63 Ma (stage TG22).

Event B slightly postdates the 6.7 Ma onset of "latest Miocene–earliest Pliocene biogenic bloom" of Farrell *et al.* (1995), which is characterized by significant increases in both $CaCO_3$ and opal MARs in the eastern equatorial Pacific. Farrell *et al.* (1995) point out the widespread nature of this "latest Miocene–earliest Pliocene biogenic bloom" in the low-latitude Indo-Pacific region. Biogenic carbonate sedimentation in the western equatorial Pacific peaks at about 6.6 Ma and declines steadily after 5 Ma according to (Berger *et al.* 1993). Sharp maxima in carbonate MARs are recorded by Peterson *et al.* (1992) between about 6 and 4 Ma along the Arabian margin and at ODP Site 707 in the low-latitude Indian Ocean. Berger *et al.* (1993) suggest that this latest Miocene-earliest Pliocene biogenic bloom was caused by greater nutrient concentrations in Pacific waters due to increased production of NADW (increased basin–basin fractionation), but

Farrell *et al.* (1995) argue for regional fractionation of nutrients, rather than a global increase in the supply of nutrients to the ocean, because of their observation of sharp intra-basinal differences of opal and $CaCO_3$ in the eastern equatorial Pacific.

Keigwin (1982a) noted increased contrast between Pacific and Caribbean benthic foraminifer $\delta^{13}C$ beginning at 6.8–6.6 Ma (equivalent to the end of the late Miocene carbon shift), which he interpreted to be evidence of termination of deep-to-intermediate-water exchange across the Central American Seaway caused by shoaling of the sill. Wright *et al.* (1991), however, did not record increased contrast between North Atlantic and Pacific benthic foraminifer $\delta^{13}C$ at this time, so it is unclear whether there was an overall increase in basin–basin fractionation. Yet, Keller *et al.* (1989) argued that intermediate water flow from the Pacific into the Caribbean became restricted at 6.8–6.6 Ma (ages updated) based on a planktonic foraminiferal assemblage change that they related to the onset of major upwelling in the western Caribbean.

The latest Miocene is characterized by highly fluctuating oxygen isotope values and rapidly changing paleoceanographic conditions (Hodell *et al.* 1994). A combination of high-latitude cooling and gradual reorganization of North Pacific water masses, possibly in response to partial closure of the Central American Seaway, may have been responsible for the seemingly progressive nature of latest Miocene increases in diatom sedimentation in different regions of the North Pacific (Fig. 5).

Event C

At about 4.5 Ma (event C), diatom MARs increased at higher latitudes of the northwest Pacific, while they declined at Site 438 off northeast Japan and were apparently waning off southern California (Figs 3 and 4). The Site 588–590 oxygen isotope curve (Fig. 5) suggests that 4.5 Ma marked the onset of a period of sustained high latitude warming that lasted at least 1 m.y. In terms of high resolution Pliocene isotope curves, Shackleton *et al.* (1995) note that the interval from 4.5 to 3.5 Ma was characterized by $\delta^{18}O$ values that are almost entirely lower than present-day values. Other evidence for climatic warming in the North Pacific at this time comes from the study by Lagoe *et al.* (1993) of the Yakataga Formation in the Gulf of Alaska, which revealed that glaciomarine sedimentation was terminated at 4.4 Ma and did not resume until about 3 Ma (ages updated).

Haug *et al.* (1995) showed that intervals of increased opal and carbonate flux at Detroit Seamount Site 882 during the middle part of the Pliocene coincided with global interglacial periods. They argued that enhanced circulation of North Atlantic Deep Water during these interglacial intervals supplied higher levels of nutrients to the surface waters of the high latitude Pacific and fueled diatom and coccolith productivity. There appears, however, to be little evidence of a major increase in the $\delta^{13}C$ contrast between benthic foraminiferal records of the Atlantic and Pacific at this time (Whitman and Berger 1993), so it is not clear that basin–basin fractionation was enhanced significantly. Rather, during this climatically warm period, large quantities of relatively warm, saline surface waters penetrated further to the north (Barron 1995). Sancetta and Silvestri (1986) argue that the modern subarctic water mass did not exist in the western Pacific, but a broad transition zone between warmer and cooler waters was present in the North Pacific, with a northern margin north of 48°N and a southern margin south of 41°N. Such a broad transition zone would have resulted in a northward spread of relatively warm, salty surface waters, which would lead to enhanced evaporation at the surface and an increased upward diffusion of nutrient-rich deep waters and fueled diatom productivity (Haug *et al.* 1995). The apparent decline in diatom productivity in middle latitude regions, such as Site 438 off Japan and southern California and increase in higher latitude regions of the northwest Pacific at about 4.5 Ma, is further indication of a circulation change at this time.

Additional evidence of changing paleoceanography in the Pacific at about 4.4 Ma comes from the eastern equatorial Pacific, where Farrell *et al.* (1995) document a permanent eastward shift in the locus of maximum opal MAR from the eastern equatorial Pacific Basin (near 0°N, 107°W) to the Galapagos region (near 3°S, 92°W). At this time there was also a marked decline in the level of $CaCO_3$ MAR throughout the eastern equatorial Pacific, signaling an end of the latest Miocene-earliest Pliocene "biogenic bloom" (Farrell *et al.* 1995). Farrell *et al.* (1995) suggest that shoaling of the Isthmus of Panama at 4.4 Ma caused the cessation of the outflow of nutrient rich waters of the North Equatorial Countercurrent into the Caribbean, resulting in a backup of nutrients in the Galapagos region and leading to increased diatom sedimentation there. Cessation of the minor westward flow of nutrient-depleted Caribbean waters into the eastern equatorial Pacific region east of the Galapagos at this time might also have aided increased diatom production in that region. On the other hand, Hovan's (Hovan 1995) eolian sediment data show that trade-wind strength decreased in the eastern equatorial Pacific between 5.0 and 4.0 a, resulting in a southward shift in the position of the intertropical convergence zone which may have led to a narrowing of the zone of maximum upwelling along the equator.

Keller *et al.* (1989) argued for increasingly restricted surface current flow across the Central American Seaway beginning at 4.4 Ma (age updated) based on comparison of planktonic foraminiferal assemblages at DSDP Sites 502 and 503, on either side of the Isthmus of Panama. Similarly, Keigwin (1982a,b) reported that planktonic foraminiferal $\delta^{18}O$ values at Caribbean Site 502 became significantly heavier than those of eastern equatorial Pacific Site 503 beginning at 4.4 Ma (age updated), implying that surface waters in the Caribbean suddenly became much more saltier than those in the eastern equatorial Pacific due to the restriction of flow of less saline Pacific surface waters into the Caribbean. Thus, it would appear that the onset of a period of climatically warmer high-latitude paleotemperatures and paleoceanographic changes linked with a shoaling of the Isthmus of Panama may have combined to cause major changes in diatom sedimentation in the Pacific at about 4.5 Ma.

Event D

After 2.7 Ma (event D), diatom accumulation rates abruptly decline at the higher latitudes of the North Pacific (Fig. 3) (Sites 882, 883, and 887), coincident with a major increase in ice rafted detritus (Haug *et al.* 1995) (Fig. 5) and the onset of Northern Hemisphere glaciation. Cooling and freshening of surface waters would have increased vertical stratification in the high latitude North Pacific and would have hindered the upward diffusion of nutrient-rich deep waters, leading to decreased opal sedimentation. At the same time, Raymo *et al.* (1992) show that after 2.75 Ma, NADW production was gradually suppressed by global cooling, implying at least a temporary decrease in basin–basin fractionation.

As stated earlier, the MAR and sedimentation rate increase at 2.7 Ma at Site 881 (Fig. 3) appears to be due to localized channel fill and is not representative of regional trends in diatom accumulation. A true increase in diatom accumulation off California at about 2.7 Ma, on the other hand, is implied by the replacement of dissolved diatoms by well-preserved diatom assemblages at this time at DSDP Sites 467 and 469 off southern California (Fig. 4). Presumably, this preservational change reflects enhanced diatom productivity due to an increase in wind driven upwelling off southern California (Barron 1981, Barron 1992).

The Isthmus of Panama apparently had completely emerged by 2.7 Ma (age updated) according to most workers (Keigwin, 1982a; Lundelius 1987; Coates *et al.* 1992), so it is unlikely that the 2.7 Ma decline in diatom sedimentation (event D) in the high-latitude North Pacific was directly related to the closure of the Central American Seaway. Rather, major reorganization of surface (Sancetta and Silvestri 1986) and intermediate (Haug *et al.* 1995) water masses at about 2.7 Ma are the likely causes of contemporaneous changes in diatom sedimentation patterns that are documented in the North Pacific.

Comparison with laboratory data

In Section 2, it was argued that the smear slide estimates of the percentage diatoms that were used to calculate diatom MAR are proportional to laboratory measurements of the weight percent opal, with the differences generally being within a factor of 2. In Fig. 6 estimates of diatom MAR at Sites 887 and 882 are compared directly with the respective laboratory measurements of opal MAR made at these sites by Rea and Snoeckx (1995) and Haug *et al.* (1995). In the two intervals that are compared, 1–10 Ma at Site 887 and 2.4–3.2 Ma at Site 882, diatom MAR appears to approximate the opal MAR within a factor of 2 or less, considering that the diatom MAR values are averaged for each core. The high resolution nature of the opal MAR data allows for a more accurate estimation of the MAR events. The event B MAR increase event occurs at 6.1 Ma at Site 887, whereas the event D MAR decrease is dated at 2.73 Ma at Site 882. The D event is not so evident in the opal MAR record of Rea and Snoeckx (1995) at Site 887. Clearly, laboratory determinations of opal weight percent are preferable to

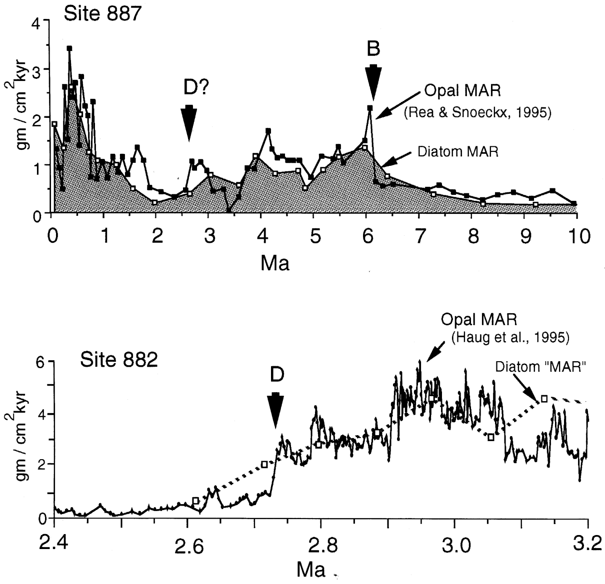

Fig. 6. Comparison of diatom MAR estimates (empty squares) with published opal MAR data (solid squares) in selected intervals of Sites 882 and 887. Diatom MAR is based on visual estimates of percent diatoms and is averaged for each core.

visual estimates of percent diatoms, but the latter estimates are routinely completed by shipboard scientists on all sediment collected by the Ocean Drilling Program (and its predecessor, the Deep Sea Drilling Project) and offer a means for making low resolution, regional comparisons.

Conclusions

Diatom mass accumulation (MARs) rates for North Pacific sections collected by Ocean Drilling Program Leg 145 were estimated based on shipboard smear slide and density data using age models that benefited from improved magnetostratigraphic calibration of Miocene diatom biostratigraphy. These records show that diatom productivity in the high latitude North Pacific was relatively weak until the later part of the early Miocene (ca. 18–19 Ma), when modest increases in diatom MAR (to <0.5 g/cm^2 k.y.) occurred at Sites 884 and 887, coincident with an expansion of biosiliceous sedimentation throughout the North Pacific (Fig. 2) (Barron 1992). These increases are part of the low latitude Atlantic-to-North Pacific "silica shift" of Keller and Barron (1983), and they appear to signal an enhancement of North Atlantic–Pacific basin–basin fractionation (Barron and Baldauf 1990) as measured by increasing contrast between the benthic foraminiferal $\delta^{13}C$ of the two ocean basins (Wright *et al.* 1992).

The greatest changes in biosiliceous sedimentation in the North Pacific, however, took place during the late Neogene. Detailed comparison of the diatom MAR records at North Pacific Sites 881, 882, 883 and 887 with records at DSDP Site 438 off northeast Japan and in coastal outcrops of the Monterey and Sisquoc formations of southern California reveals that at least four major changes in diatom sedimentation patterns occurred in the North Pacific during the last 10 m.y.

At 9 Ma (event A), diatom accumulation rates increased in the northwest Pacific (Sites 883 and 884) and off northeast Japan (Site 438). A slightly younger (8.5 Ma) biogenic silica MAR increase is suggested for the Monterey Formation in southern California by Isaacs (Isaacs, 1983, 1985), but it is poorly constrained by biostratigraphy. Event A appears to fall within an interval of global warming or relatively stable paleotemperatures, so it does not seem likely that the diatom MAR increase was due to enhanced upwelling caused by high latitude cooling. A major divergence of benthic foraminiferal $\delta^{13}C$ data that began at about 8.8 Ma (Wright *et al.* 1991) is taken as evidence of enhanced North Atlantic–Pacific, basin–basin fractionation which would have increased nutrient levels in the North Pacific and fueled increased diatom productivity. Because there is both paleoceanographic (Lyle *et al.* 1995) and faunal (Marshall 1988) evidence for shoaling of the Isthmus of Panama immediately prior to 9 Ma, it is possible that reduced flow of intermediate waters from the Caribbean to the Pacific may have caused an enhancement of nutrient levels in North Pacific deep waters that ultimately led to increased diatom productivity.

A second expansion of diatom sedimentation in the North Pacific occurred progressively during the latest Miocene, seemingly beginning at 6.5 Ma in coastal California, spreading to the high latitudes at 6.2 Ma (event B), and finally affecting Site 438 off northeast Japan at 5.5 Ma (Figs 3 and 4). This stepwise increase follows slightly the 6.7 Ma onset of a latest Miocene-earliest Pliocene biogenic bloom that is documented in the eastern equatorial Pacific and it mostly falls between major high latitude cooling steps that are identified by oxygen isotope studies (Fig. 5). Although it does not seem to coincide with an increase in basin–basin fractionation according to $\delta^{13}C$ benthic foraminiferal records of Wright *et al.* (1991), it does immediately follow a time (6.8–6.6 Ma) when both planktonic foraminiferal and isotope data both suggest that intermediate water flow was further restricted across the Isthmus of Panama (Keigwin, 1982a; Keller *et al.* 1989). Progressive high latitude cooling, coupled with a gradual reorganization of North Pacific water masses, appear to be responsible for these latest Miocene changes in diatom sedimentation in the North Pacific.

At about 4.5 Ma (event C) diatom accumulation rates appear to have declined at middle latitudes of the North Pacific (Site 438 and offshore southern California) while they increased at higher latitudes of the northwest Pacific (Sites 882 and 883) (Figs 3 and 4). No change is observable at northeast Pacific Site 887. This event appears to coincide with the onset of generally warmer high latitude paleotemperatures that persisted for at least another million years (Fig. 5) (Shackleton *et al.* 1995). It is reasoned that transport of warmer, more saline waters to higher latitudes of the North Pacific may have lead to a decrease in the vertical stratification of the water mass and permitted increased upwelling of nutrient rich deep waters that fueled diatom production (Haug *et al.* 1995). At the same time, a reduced pole-to-equator thermal gradient may caused a slackening of offshore winds and reduced upwelling along the coasts of Japan and southern California at this time, leading to a decline in diatom productivity there.

In the eastern equatorial Pacific, the locus of maximum opal accumulation permanently shifted eastward to the region east of the Galapagos Islands at 4.4 Ma, possibly as a result of the shoaling of the Isthmus of Panama and a cessation of major surface water flow from the Pacific into the Caribbean (Farrell *et al.* 1995). Published comparisons of North Atlantic–Pacific $\delta^{13}C$ isotope records do not suggest increased basin–basin fractionation at this time, so it would be premature to suggest that shoaling of the Isthmus lead to enhanced nutrient levels in North Pacific deep waters at this time. Nevertheless, Haug *et al.* (1995) convincingly argue that increased opal and carbonate production at northwest Pacific Site 882 during interglacial periods of the middle Pliocene are due to strengthening of the NADW-driven, "conveyor belt" of deep water circulation.

At 2.7 Ma (event A) major cooling of surface waters and the southward spread of icebergs caused increased vertical stratification of water masses in the North Pacific, resulting in a slackening in the upwelling of nutrient-rich deep waters in the northwest Pacific (Haug *et al.* 1995). Diatom MARs at high latitude Sites 882, 883 and 887 sharply declined, and high diatom productivity probably shifted more toward the coasts of Asia and North America under the influence of increased zonal winds. Increased offshore winds appear to have enhanced upwelling and renewed diatom productivity off the coast of southern California at about the same time (Barron 1981 Barron 1992). Modern patterns of diatom sedimentation were largely established over broad regions of the North Pacific at this time, although diatom productivity increased markedly during the 0.8 m.y. at northeast Site 887, where it likely reflects a greater focusing of the subpolar upwelling system associated with the Alaskan gyre (Rea and Snoeckx 1995).

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REFERENCES

Akiba, F. (1986) Middle Miocene to Quaternary diatom biostratigraphy in the Nankai Trough and Japan Trench, and modified lower Miocene through Quaternary diatom zones for the middle-to-high latitudes of the North Pacific. In *Initial Reports, Deep Sea Drilling Program* (Edited by Kagami, H., Karig, D.E. and Coulbourn, W.T.), Vol. 87, pp. 393–481. US Government Printing Office, Washington, DC.

Arthur, M.A., von Huene, R. and Adelseck, C.G. Jr (1980) Sedimentary evolution of the Japan fore-arc region off northern Honshu, Legs 56 and 57, Deep Sea Drilling Project. In *Initial Reports, Deep Sea Drilling Program* (Edited by Scientific Party), Vols 56, 57, pp. 521–568. US Government Printing Office, Washington, DC.

Barron, J.A. (1980) Lower Miocene to Quaternary diatom biostratigraphy of DSDP Leg 57, off northeastern Japan. In *Initial Reports, Deep Sea Drilling Program* (Edited by Scientific Party), Vols 56, 57, pp. 641–693. US Government Printing Office, Washington, DC.

Barron, J.A. (1981) Late Cenozoic diatom biostratigraphy and paleoceanography of the middle-latitude North Pacific, Deep Sea Drilling Project Leg 63. In *Initial Reports, Deep Sea Drilling Program* (Edited by Yeats, R.L., and Haq, B.U.), Vol. 63, pp. 507–538. US Government Printing Office, Washington, DC.

Barron, J.A. (1992) Paleoceanographic and tectonic controls on the Pliocene diatom record of the northeastern Pacific. In *Pacific Neogene-Environment, Evolution, and Events:* (Edited by Tsuchi, R. and Ingle, J.C., Jr), pp. 25–41. Tokyo University Press, Tokyo.

Barron, J.A. (1995) High resolution diatom paleoclimatology of the middle part of the Pliocene of the northwest Pacific. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Rea, D.K., Basov, I.A., Scholl, D.W, and Allan, J.F.), Vol. 145, pp. 43–53. Ocean Drilling Program, College Station, TX.

Barron, J. A. and Baldauf, J. G. (1990) Development of biosiliceous sedimentation in the North Pacific during the Miocene and early Pliocene. In *Pacific Neogene Events— Their Timing Nature and Interrelationship* (Edited by Tsuchi, R.), pp. 43–63. University of Tokyo Press, Tokyo.

Barron, J.A. and Gladenkov, A.Y. (1995) Early Miocene to Pleistocene diatom stratigraphy of Leg 145. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Rea, D.K., Basov, I.A., Scholl, D.W, and Allan, J.F.), Vol. 145, pp. 21–41. Ocean Drilling Program, College Station, TX.

Barron, J. A. and Keller, G. (1983) Paleotemperature oscillations in the middle and late Miocene of the northeastern Pacific. *Micropaleontology* **29,** 150–181.

Barron, J.A., Basov, I.A., Beaufort, L., Dubuisson, G., Andrey Y. Gladenkov. A.Y., Morley, J.J., Okada, M., Olafsson, G., Pak, D.K., Roberts, A.P., Shilov, V.V. and Weeks, R.J. (1995) Biostratigraphic and magnetostratigraphic summary of ODP Leg 145. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Rea, D.K., Basov, I.A., Scholl, D.W. and Allan, J.F.), Vol. 145, pp. 559–575. Ocean Drilling Program, College Station, TX.

Barton, C.E. and Bloemendal, J. (1986) Paleomagnetism of sediments collected during Leg 90, southwest Pacific. In *Initial Reports, Deep Sea Drilling Program* (Edited by Kennett, J.P. and von der Borch, C.C.), Vol. 90, pp. 1273–1316. US Government Printing Office, Washington, DC.

Berger, W. H. (1970) Biogenous deep-sea sediments: fractionation by deep-sea circulation. *Deep Sea Research* **81,** 31–43.

Berger, W.H., Leckie, R.M., Janecek, T.R., Stax, R. and Takayama, T. (1993) Neogene carbonate sedimentation on Ontong Java Plateau: highlights and open questions. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Berger, W.H., Kroenke, L.W. and Mayer, L.A.), Vol. 130, pp. 711–744. Ocean Drilling Program, College Station, TX.

Broecker, W.S. and Peng, T.-H. (1982) *Tracers in the Sea*. Lamont–Doherty Geological Observatory, Palisades, NY.

Cande, S. C. and Kent, D. V. (1992) A new geomagnetic polarity time scale for the Late Cretaceous and Cenozoic. *Journal of Geophysical Research* **97**(B10), 13917–13951.

Coates, A. G., Jackson, J. B. C., Collins, L. S., Cronin, T. M., Dowsett, H. J., Bybell, L. M., Jung, P. and Obando, J. A. (1992) Closure of the Isthmus of Panama: the near-short marine record of Coasta Rica and western Panama. *Geological Society of America Bulletin* **104,** 814–828.

Craig, H, Broecker, W.S. and Spencer, D. (1981) *GEOSECS Pacific Expedition* (Vol. 4J: Sections and Profiles). National Science Foundation, Washington, DC.

Dumont, M.P. and Barron, J.A. (1995) Diatom biochronology of the Sisquoc Formation in the Santa Maria Basin, California, and its paleoceanographic and tectonic implications. In *U.S. Geological Survey Bulletin 1995* (Edited by Keller, M.A.), pp. K1–K-16.

Farrell, J., Raffi, I., Janecek, T., Murray, D., Levitan, M., Dadey, K., Emeis, K.-C., Lyle, M., Flores, J.-A. and Hovan, S. (1995) Late Neogene sedimentation patterns in the eastern equatorial Pacific Ocean. In *Proceedings of the Ocean Drilling Program, Science Results*, (Edited by Pisias, N.G., Mayer, L.A., Janecek, T.R., Palmer-Julson, A. and van Andel, T.H.), Vol. 138, pp. 717–756. Ocean Drilling Program, College Station, TX.

Gladenkov, A.Y. and Barron, J.A. (1995) Oligocene and early middle Miocene diatom biostratigraphy of Hole 884B. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Rea, D.K., Basov, I.A., Scholl, D.W. and Allan, J.F.), Vol. 145, pp. 21–51. Ocean Drilling Program, College Station, TX.

Haug, G.H., Maslin, M.A., Sarnthein, M., Stax, R. and Tiedemann, R. (1995) Evolution of northwest Pacific sedimentation patterns since 6 Ma: Site 882. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Rea, D.K., Basov, I.A., Scholl, D.W. and Allan, J.F.), Vol, 145, pp. 293–314. Ocean Drilling Program, College Station, TX.

Hamilton, T.S. (1995) Seismic stratigraphy of Cenozoic sediments from North Pacific seamount platforms and deep-sea sites, Leg 145. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Rea, D.K., Basov, I.A., Scholl, D.W, and Allan, J.F.), Vol. 145, pp. 437–453. Ocean Drilling Program, College Station, TX.

Hodell, D. A., Benson, R. H., Kent, D. V., Boersma, A. and Rakic-El Bied, K. (1994) Magnetostratigraphic, biostratigraphic, and stable isotope stratigraphy of an upper Miocene drill core from the Sale Biqueterie (northwestern Morocco); a high-resolution chronology for the Messinian stage. *Paleoceanography* **9,** 835–855.

Hovan, S.A. (1995) Late Cenozoic atmospheric circulation intensity and climatic history recorded by eolian deposition in the eastern equatorial Pacific Ocean, Leg 138. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Pisias, N.G., Mayer, LA., Janecek, T.R., Palmer-Julson, A. and van Andel, T.H.), Vol. 138, pp. 615–625. Ocean Drilling Program, College Station, TX.

Isaacs, C.M. (1983) Compositional variation and sequence in the Miocene Monterey Formation, Santa Barbara coastal area, California. In *Cenozoic Marine Sedimentation, Pacific Margin, U.S.A.* (Edited by Larue, D.K. and Steel, D.K.), pp. 117–132. Pacific Section, Society of Economic Paleontologists and Mineralogists, Los Angeles, CA.

Isaacs, C. M. (1985) Abundance versus rates of accumulation in fine-grained strata of the Miocene Santa Barbara basin, California. *Geo-Marine Letters* **5,** 25–30.

Keigwin, L. D. (1982a) Isotopic paleoceanography of the Caribbean and east Pacific: Role of Panama uplift in late Neogene time. *Science* **217,** 350–353.

Keigwin, L.D. (1982b) Stable isotope stratigraphy and paleoceanography of Sites 502 and 503. In *Initial Reports of the Deep Sea Drilling Project* (Edited by Prell, W.L.and Gardner, J.V.), Vol. 68, pp. 445–453. US Government Printing Office, Washington, DC.

Keller, G. and Barron, J. A. (1983) Paleoceanographic implications of Miocene deep-sea hiatuses. *Geological Society of America Bulletin* **94,** 590–613.

Keller, G., Zenker, C. E. and Stone, S. M. (1989) Late Neogene history of the Pacific–Caribbean gateway. *Journal of South American Earth Sciences* **2,** 73–108.

Kennett, J.P. (1986) Miocene to early Pliocene oxygen and carbon isotope stratigraphy in the southwest Pacific, Deep Sea Drilling Project Leg 90. In *Initial Reports, Deep Sea Drilling Program* (Edited by Kennett, J.P. and von der Borch, C.C.), Vol. 90, pp. 1383–1411. US Government Printing Office, Washington, DC.

Lagoe, M. B., Eyles, C. H., Eyles, N. and Hale, C. (1993) Timing of late Cenozoic tidewater glaciation in the far North Pacific. *Geological Society of America Bulletin* **105,** 1542–1560.

Lazarus, D., Spencer-Cervato, C., Pianka-Biolzi, M., Beckmann, J.P., von Salis, K., Hilbrecht, H. and Thierstein, H. (1995) Revised Chronology of Neogene DSDP Holes from the World Ocean, *Ocean Drilling Program, Texas A and M University Technical Note*, 24. Ocean Drilling Program, College Station, TX.

Lundelius, E.L. (1987) The North American Quaternary sequence. In *Cenozoic Mammals of North America*. (Edited by Woodburne, M.O.), pp. 211–235. University of California Press, Berkeley, CA.

Lyle, M., Dadey, K.A. and Farrell, J.W. (1995) The late Miocene (11–8 Ma) eastern Pacific carbonate crash: evidence for reorganization of deep-water circulation by closure of the Panama gateway. In *Proceedings of the Ocean Drilling Program, Science Results*

(Edited by Pisias, N.G., Mayer, L.A., Janecek, T.R., Palmer-Julson, A., and van Andel, T.H.), Vol. 138, pp. 821–838. Ocean Drilling Program, College Station, TX.

Maier-Reimer, E., Mikolajewicz, U. and Crowley, T. (1990) Ocean general circulation model sensitivity experiment with an open central American Isthmus. *Paleoceanography* **5,** 349–366.

Marshall, L. G. (1988) Land mammals and the Great American Interchange. *American Scientist* **76,** 380–388.

Peterson, L.C., Murray, D.W., Ehrmann, W.U. and Hempel, P. (1992) Cenozoic carbonate accumulation and compensation depth changes in the Indian Ocean. In *Synthesis of Results from Scientific Drilling in the Indian Ocean, American Geophysical Union Monograph* (Edited by Duncan, R. A., Rea, D. K., Kidd, R. B., von Rad, U. and Weissel, J.K.), Vol. 70, pp. 311–333. American Geophysical Union, Washington, DC.

Raymo, M. E., Hodell, D. and Jansen, E. (1992) Response of deep ocean circulation to initiation of Northern Hemisphere glaciation (3–2 Ma). *Paleoceanography* **7,** 645–672.

Rea, D.K. and Snoeckx, H. (1995) Sediment fluxes in the Gulf of Alaska: paleoceanographic record from ODP Site 887 on the Patton-Murray Seamount platform. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Rea, D.K., Basov, I.A., Scholl, D.W. and Allan, J.F.), Vol. 145, pp. 247–256. Ocean Drilling Program, College Station, TX.

Rea, D.K., Basov, I.A., Janecek, T.R., Palmer-Julson, A. *et al.* (1993) In *Proceedings of the Ocean Drilling Program, Science Results* Vol. 145. Ocean Drilling Program, College Station, TX.

Sancetta, C. and Silvestri, S. (1986) Pliocene–Pleistocene evolution of the North Pacific Ocean-atmosphere system, interpreted from fossil diatoms. *Paleoceanography* **1,** 163–180.

Shipboard Scientific Party (1980) Sites 438 and 439: Japan Deep Sea Terrace. In *Initial Reports of the Deep Sea Drilling Project* (Edited by Scientific Party), Vol. 56, 57. pp. 23–191. US Government Printing Office, Washington, DC.

Shipboard Scientific Party (1990) Site 799. In *Proceedings of the Ocean Drilling Program, Initial Reports* (Edited by Ingle, J.C., Jr., Suyehiro, K. and von Breymann, M.T.), Vol. 128, pp. 237–402. Ocean Drilling Program, College Station, TX.

Shipboard Scientific Party (1993) Site 881. In *Proceedings of the Ocean Drilling Program, Initial Reports* (Edited by Rea, D.K., Basov, I.A., Janecek, T.R. and Palmer-Julson, A.), Vol. 145, pp. 37–83. Ocean Drilling Program, College Station, TX.

Shackleton, N.J., Hall, M.A. and Pate, D. (1995) Pliocene stable isotope stratigraphy of Site 846. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Pisias, N.G., Mayer, L.A., Janecek, T.R., Palmer-Julson, A. and van Andel, T.H.), Vol. 138, pp. 337–355. Ocean Drilling Program, College Station, TX.

Whitman, J.M. and Berger, W.H. (1993) Pliocene–Pleistocene carbon isotope record, Site 586, Ontong Java Plateau. In *Proceedings of the Ocean Drilling Program, Science Results* (Edited by Berger, W.H., Kroenke, L.W. and Mayer, L.A.), Vol. 130, pp. 333–348. Ocean Drilling Program, College Station, TX.

Woodruff, F. and Savin, S. M (1989) Miocene deep-water oceanography. *Paleoceanography* **4,** 87–140.

Wright, J. D., Miller, K. G. and Fairbanks, R. G. (1991) Evolution of modern deepwater circulation: Evidence from the late Miocene Southern Ocean. *Paleoceanography* **6,** 275–290.

Wright, J. D., Miller, K. G. and Fairbanks, R. G. (1992) Early and middle Miocene stable isotopes: Implications for deepwater circulation and climate. *Paleoceanography* **7,** 357–389.