ELSEVIER

Palaeogeography, Palaeoclimatology, Palaeoecology 108 (1994) 537-555

PALAEO

The middle Miocene climatic transition: East Antarctic ice sheet development, deep ocean circulation and global carbon cycling

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(Received July 2, 1993; revised and accepted July 28, 1993)

Abstract

The middle Miocene represents a major change in state in Cenozoic climatic evolution, following the climax of Neogene warmth in the late early Miocene at ~16 Ma. The early stage of this climatic transition from ~16 to 14.8 Ma was marked by major short term variations in global climates, East Antarctic Ice Sheet (EAIS) volume, sea level, and deep ocean circulation. In the later stage from ~14.8 to 12.9 Ma, climatic developments included major growth of the EAIS and associated Antarctic cooling, a distinct increase in the meridional temperature gradient, large fluctuations in sea level followed by a global sea level fall, and important changes in deep water circulation, including increased production of Southern Component Water. East Antarctic ice sheet growth and polar cooling also had large effects on global carbon cycling and on the terrestrial biosphere, including aridification of mid-latitude continental regions. Increased stability of the EAIS after 14.8 Ma represents a crucial step in the establishment of late Neogene global climate systems.

What controlled these changes in polar climates and the East Antarctic ice sheet? Deep ocean circulation changes probably played a major role in the evolution and variation in polar climates, as they have throughout the Cenozoic. Oxygen and carbon isotopic evidence for warm, saline deep water production in the eastern Tethyan/northern Indian Ocean indicates that meridional heat transport to the Antarctic inhibited Cenozoic polar cooling and EAIS growth during the early middle Miocene from ~16 to ~14.8 Ma. Inferred competition between warm low-latitude sources (derived from the eastern Tethyan-northern Indian Ocean) and a cold high-latitude source (Southern Component Water) from ~16 to 14.8 Ma may have been associated with instability in the Antarctic climate and cryosphere. Reduction of warm, saline deep water flow to the Southern Ocean at ~14.8 Ma may have decreased meridional heat transport to the Antarctic, cooling the region and leading to increased production of Southern Component Water.

These middle Miocene climatic and cryospheric changes in the Antarctic had profound effects on marine and terrestrial climates. As the meridional surface temperature gradient increased, boundaries between climatic zones strengthened, leading to increased aridification of mid-latitude continental regions in Australia, Africa and North and South America, enhancing the development of grasslands and stimulating the evolution of grazing mammals.

1. Introduction

The Antarctic ice sheets are a major component of the Earth's climate system, strongly influencing ocean and atmospheric circulation. The long-term evolution of the Antarctic cryosphere is currently under much debate, partly because of the hypothesis of Webb and Harwood (1991) and Barrett et al. (1992) that Antarctic ice sheets were very unstable during the late Neogene. Better understanding of the Cenozoic history of the Antarctic cryosphere and its relations to global climates and deep ocean circulation is needed to evaluate the long-term stability of the ocean-climate system and its effects

SSDI 0031-0182(93)E0165-P

on biotic evolution. Recent papers have discussed the relation between Antarctic cryospheric development and deep ocean circulation during a major growth phase of the East Antarctic ice sheet (EAIS) in the middle Miocene from 16 to 12 Ma (Woodruff and Savin, 1989, 1991; Wright et al., 1992; Flower and Kennett, 1993a; in review).

The middle Miocene represents a major change in state in the climatic evolution of the Cenozoic. From cool climates of the early Oligocene, global climates warmed during the late Oligocene and reached a climatic optimum in the late early Miocene (the warmest interval of the Neogene and the warmest since the late Eocene), then underwent rapid cooling during the middle Miocene (Shackleton and Kennett, 1975; Savin et al., 1975; Miller et al., 1987; Hornibrook, 1992). Further late Neogene global cooling included important steps in the Miocene and the late Pliocene and Quaternary; the latter step involving the onset of Northern Hemisphere ice sheets. Some of the major climatic and paleoceanographic events of the middle Miocene are summarized below. The time scale of Berggren et al. (1985) is used throughout the discussion.

(1) A large increase in benthic foraminiferal $\delta^{18}O$ of $\sim$1.0–1.3‰ occurred from $\sim$16 to 12.5 Ma (Shackleton and Kennett, 1975; Miller et al., 1987). This global benthic increase has been inferred to reflect a combination of EAIS growth and deep water cooling (Shackleton and Kennett, 1975; Savin, 1977; Woodruff et al., 1981; Kennett, 1986; Woodruff and Savin, 1989, 1991; Wright et al., 1992), although some workers have attributed the increase entirely to deep water cooling (Matthews and Poore, 1980; Prentice and Matthews, 1988; Prentice and Matthews, 1991).

(2) Middle Miocene sea levels are marked by high-amplitude variations from $\sim$16 to 14 Ma followed by a two-step, semi-permanent sea level fall between $\sim$14 and 12.5 Ma (Haq, 1987).

(3) Middle Miocene climatic changes were associated with a pulse of evolutionary turnover in terrestrial and marine biota. Planktonic foraminifera exhibit increased turnover from the tropics to the high latitudes in the late early to early middle Miocene (Wei and Kennett, 1986).

(4) Benthic foraminiferal assemblages underwent a significant evolutionary turnover from $\sim$17 to 14 Ma (Woodruff, 1985; Thomas, 1985, 1987; Miller and Katz, 1987). In the southwest Pacific, the establishment of Neogene benthic assemblages at middle bathyal depths coincided with the largest middle Miocene $\delta^{18}O$ increase from 14.5 to 14.1 Ma (Kurihara and Kennett, 1986; Kurihara and Kennett, 1992).

(5) Surface ocean circulation systems were invigorated; changes included intensification of gyral circulation and consequent increases in the strength of oceanographic fronts (Thunell and Belyea, 1982; Kennett et al., 1985).

(6) Early to middle Miocene mean ocean $\delta^{13}C$ records show a broad increase from $\sim$17 to 13.5 Ma, the so-called "Monterey Carbon Excursion." This positive excursion is composed of several distinct $\delta^{13}C$ maxima which reflect episodic large-scale changes in organic carbon deposition relative to carbonate sedimentation (Vincent and Berggren, 1985; Woodruff and Savin, 1991; Flower and Kennett, 1993b,c).

(7) Large sedimentary deposits of organic carbon such as the Monterey Formation of California (Vincent and Berger, 1985) and the phosphatic deposits of the southeastern U.S. (Compton et al., 1990, 1993) may have influenced global climates through the sequestering of organic carbon and consequent drawdown of atmospheric partial CO_2, via a series of positive-feedback mechanisms.

(8) A redistribution of silica and carbonate deposition in the global ocean occurred, involving the transfer of the locus of biogenic silica deposition from the Atlantic Ocean to the North Pacific and the Antarctic Oceans (Keller and Barron, 1983).

(9) Plate tectonic developments included the middle Miocene closure of the eastern portal of the Tethys Ocean in the present day eastern Mediterranean (Hsu and Bemouth, 1978) and restriction of the Indonesian Seaway (Kennett et al., 1985).

(10) Deep water circulation underwent major changes throughout the middle Miocene. A Tethyan-Indian Saline Water (TISW) deep water source present in the early Miocene probably ended during the middle Miocene (Woodruff and

Savin, 1989). Southern Component Water production increased during the middle Miocene, while Northern Component Water production was low between ~16 and 12.5 Ma (Woodruff and Savin, 1989, 1991; Wright et al., 1992).

(11) Continental climates underwent major changes in the middle Miocene. Increased aridity is inferred at this time for mid-latitude continental regions including Australia (Robert et al., 1986; Stein and Robert, 1986), Africa (Retallack, 1992), North America (Webb, 1977; Wolfe, 1985) and South America (Pascual and Jaureguizar, 1990), and may have fostered the development of grasses and the consequent evolution of grassland-adapted biota.

2. Discussion

The main purposes of this paper are (1) to review middle Miocene climate history in the context of Cenozoic climatic evolution, (2) to summarize evidence for middle Miocene East Antarctic ice sheet variations, (3) to discuss deep ocean circulation changes as a potential control of Antarctic climates and East Antarctic ice sheet variations, (4) to examine relations between global cooling and carbon cycling, (5) to review ocean silica and carbonate sedimentation patterns in the context of changes in deep water circulation patterns, and (6) to summarize the influence of middle Miocene climate change on the evolution of Neogene terrestrial environments and biota.

2.1. Middle Miocene climate change: Cenozoic context

The middle Miocene climatic transition represents a major step in Cenozoic climatic evolution. This step is one of a number of threshold events which punctuated the general trend of climatic cooling and increased cryospheric development during much of the Cenozoic. A summary of oxygen isotopic data on benthic foraminifera provides a clear record of the major deep-sea cooling steps and continental ice sheet growth (Fig. 1; Miller et al., 1987). Benthic foraminiferal $\delta^{18}O$ records demonstrate that the global Cenozoic trend toward colder deepwater temperatures and greater polar ice volumes was punctuated by several rapid steps (Fig. 1). From a climatic optimum in the early to early middle Eocene (the warmest interval of the Cenozoic) climates cooled through the remainder of the Eocene and underwent a major cooling in the late Oligocene ~36 Ma. This interval witnessed the initiation of thermohaline circulation dominated by cold deep waters (Shackleton and Kennett, 1975; Benson, 1975). Antarctic climates fluctuated through the Oligocene, but did not approach the global warmth of the early to early middle Eocene (Kennett and Barker, 1990).

The second major, permanent step in Cenozoic cooling occurred in the middle Miocene from ~14.8 to 14.1 Ma. This event began just after the early/middle Miocene boundary, and divides the Neogene into two distinct intervals, an early Neogene warm interval and a late Neogene cool interval. Middle Miocene cooling was associated with increased production of cold Antarctic deep

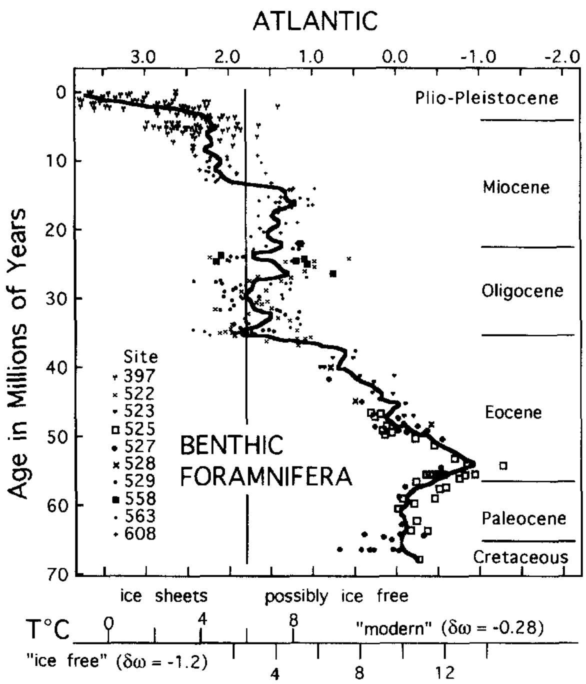

Fig. 1. Summary of Cenozoic benthic foraminiferal oxygen isotopic records from the Atlantic Ocean (from Miller et al., 1987).

waters and major growth of the East Antarctic ice sheet, as discussed below. Further polar cooling, especially in the Antarctic region, is reflected in a late Miocene $\delta^{18}O$ increase. The last major Cenozoic climatic transition occurred during the late Pliocene with the first development of Northern Hemisphere ice sheets. The middle Miocene event thus represents an important step between the initiation of global cooling following the early middle Eocene and the establishment of modern ocean–climate systems of the Quaternary.

2.1.1. Biotic evidence for middle Miocene climatic change

A large body of paleontological evidence documents the evolution of Cenozoic climates in the middle Miocene. Both terrestrial and marine biotic records indicate a major transition from a climatic optimum near the early/middle Miocene boundary marked by broad tropical and warm subtropical biotic provinces, which contracted to lower latitudes during the middle Miocene.

Some of the best land-based marine records documenting widespread climatic change are from the Cenozoic marine sequences of New Zealand, Japan, and eastern Russia. Excellent marine sequences in New Zealand have provided a detailed history for much of the Cenozoic. A recent summary of Cenozoic marine paleoclimates in New Zealand (Hornibrook, 1992) demonstrates that, from cool temperatures of the early Oligocene, climates warmed during the late Oligocene to early Miocene, reaching a climatic optimum in the late early Miocene ~16 Ma. These trends are evidenced by the southward migration of tropical and subtropical fauna in the late early Miocene. Warm-water, larger benthic foraminifera and molluscs reached their southernmost distribution limits at this time. The presence of these forms in the southern South Island suggest that a warm subtropical sea, the warmest since the middle Eocene, surrounded most of New Zealand in the late early Miocene (Hornibrook, 1992). Reef-building corals which live in the present day in the tropical Great Barrier Reef within a temperature range of 19-28°C became well-established and diverse on the northern North Island at this time (Hornibrook, 1992).

Neogene fossil sequences from Japan demonstrate that temperature changes occurred contemporaneously in the Northern Hemisphere, indicating inter-hemispheric climatic changes. Marine and terrestrial fossil assemblages indicate early Miocene temperatures warmed to a mid-Neogene climatic optimum in Japan near the early–middle Miocene boundary at ~16 Ma. The mid-Neogene optimum is one of the most pronounced Neogene events in Japan (Tsuchi, 1990). Early middle Miocene biotic trends document the singularity of this Neogene climatic optimum. A distinct fossil horizon found throughout Honshu contains many marine faunas of tropical affinity. These include mangrove swamp molluscs, reef-building corals, pteropods, nautiloids and nektonic turtles. Terrestrial biota of tropical character include pollen from mangrove trees, oxidized red soils, and a fossil dung beetle from the Noto Peninsula (Itoigawa and Yamanoi, 1990). Paleotemperature estimates based on molluscan assemblages for the Mizunami Group suggest paleotemperatures ~6°C higher than the present day. Distinctive red beds from the middle Miocene Tsugara Formation, Niigata Prefecture, suggest laterization, or similar warm-climate weathering processes during the early middle Miocene. A tropical to subtropical marine fauna migrated into northern Japan. The fauna includes gastropods that originated in the Tethys Sea, as well as larger foraminifera, reef-building corals and mangroves (Itoigawa and Yamanoi, 1990). Subtropical palynofloras extended as far north as eastern Siberia and Kamchatka (Volkova et al., 1986). The occurrence of subtropical molluscan species (~60% of the assemblage) and temperate flora (33% *Fagus*) in boreal Kamchatka (Gladenkov, 1992) further illustrates the extent of early middle Miocene warmth in the northwest Pacific region.

Fossil terrestrial floral assemblages from different continents also confirm an early middle Miocene climatic optimum, followed by rapid climatic cooling (Wolfe, 1985). Along the Pacific coast of North America, broad-leaved evergreen forest reached at least to 45°N (Wolfe, 1981). Fossil leaf and pollen data from the Nenana coal field in the Alaska Range, Alaska, suggest an interval of particular warmth from ~18 to 14 Ma (Leopold, pers. comm., 1993). In the Great Plains region, the Kilgore flora near the Nebraska–South Dakota border (of Barstovian age, ~14 Ma) features occasional subtropical elements that accompany large tortoises and crocodiles (Axelrod, 1985). Eurasian and African climates also experienced a climatic optimum during the late early to early middle Miocene (Bernor, 1984).

The early middle Miocene climatic optimum was immediately followed by an abrupt cooling in mid- to high-latitudes. Forests throughout the Northern Hemisphere underwent a major transition during the middle Miocene from a mixed northern hardwood forest to a mixed conifer forest at about 65°N (Wolfe, 1981; 1985; Askin and Spicer, in press). Thermophilous hardwood forest abundances in Alaska shifted in pulses about 15.5 Ma, became rare after 15.5 Ma, then disappeared (Leopold, pers. comm., 1993). This cooling is evidenced in Japan by the replacement of the tropical Kadonosawa Fauna by the Shiobara-Yama Fauna in northern Japan toward the upper part of tropical foraminiferal zone N9 (Itoigawa and Yamanoi, 1990) at ~14 Ma. Climatic cooling in New Zealand was marked by the disappearance of the larger tropical foraminifera (Hornibrook, 1992). Tropical to subtropical conditions in Europe marked by mangrove communities in the western Mediterranean became much cooler and drier (Steininger et al., 1985) and contracted to lower latitudes (Zachariasse, 1983).

Oceanic faunal data provide continuous records with widespread geographic coverage of large-scale climatic changes, and illustrate increased zonality of climatic belts in association with middle Miocene cooling (Kennett, 1977; Kennett, 1978). Changing distribution patterns of oceanic surface plankton provinces clearly reflect middle Miocene strengthening of climatic belts, including a contraction of the tropical and subtropical provinces. The Neogene climatic optimum near the early/middle Miocene boundary was marked by peak planktonic foraminiferal diversity in zone N8 at about 16 Ma (Cifelli, 1969; Jenkins, 1973). Subtropical to tropical planktonic foraminiferal (Thunell and Belyea, 1982; Kennett et al., 1985) and coccolithophorid (Haq, 1980) provinces in the Atlantic Ocean were widely distributed at this time and underwent a marked contraction towards lower latitudes during the middle Miocene. From analysis of changes in the depth distribution of planktonic foraminifera, Keller (1985) suggested an increased division of habitat niches within the water column accompanied significant surface water warming in the tropics (Savin et al., 1985) in the middle Miocene. Tropical coral reef communities in the Gulf of Papua (10°S paleolatitude) appeared in the late early Miocene and continued through the middle Miocene (Davies et al., 1989; Feary et al., 1991), supporting suggestions that warming of tropical surface waters accompanied mid- to high-latitude cooling and growth of the EAIS.

Marine biotic records also provide evidence for middle Miocene invigoration of surface circulation. Polar/subpolar faunas developed in the North Atlantic during the middle Miocene, and provide evidence for a cool eastern boundary current as far south as ~30°N (Thunell and Belyea, 1982). Hodell and Kennett (1985) provide evidence for the provincialization of planktonic foraminiferal faunas in the South Atlantic, and for the intensification of the Benguela Current after 16 Ma. In the eastern equatorial Pacific, nearly 50% of the diatom assemblages were replaced in conjunction with major middle Miocene cooling from 14.9 to 12.4 Ma, reflecting increased upwelling and cooling in surface waters (Barron and Baldauf, 1990). The cool water form *Denticulopsis hustedtii* increased abruptly in the equatorial Pacific at 13.5 Ma (Barron and Baldauf, 1990) and in the northwest Pacific at 13.8 Ma (Koizumi, 1990). Warm-water diatoms in the northwest Pacific were replaced by cold-water species from 14.0 to 13.3 Ma, with the greatest turnover occurring at 13.8 Ma (Koizumi, 1990), probably in response to increased strength of the Kuroshio Current.

In short, the middle Miocene climatic transition represents an important change in state in the climatic evolution of the Cenozoic. Warm climates of the early Miocene were the warmest since the middle Eocene; middle Miocene cooling from ~15 to 12.5 Ma led to colder climates in mid- to high-latitudes, greater climatic zonality and more invigorated surface ocean circulation than at any previous interval in the Cenozoic.

2.2. Evolution of the Antarctic cryosphere

Some of the best evidence for the cryospheric development of Antarctica comes from deep-sea flux records of glacially-derived ice-rafted detritus (IRD) in Antarctic and Subantarctic cores. IRD can be transported great distances by icebergs derived from large ice sheets. It is common in modern Southern Ocean sediments as far north as the Antarctic Polar Front Zone (~10–50°S), which acts as a northern barrier for icebergs.

IRD is very rare through the Oligocene to the middle Miocene in the Weddell Sea off of East Antarctica (Kennett and Barker, 1990). Earlier occurrences are documented from the early Oligocene (Zachos et al., 1991) and from the late Eocene to early Oligocene (Barrett et al., 1989; Barron et al., 1991). Continental ice sheets were present during certain intervals of the Oligocene (e.g. Miller et al., 1987), but almost certainly were less extensive than in the present day. It is possible that the East Antarctic ice sheet was of temperate rather than polar climatic character (Barrett et al., 1989). Indeed, paleobotanical evidence of Antarctic forest development is incompatible with extensive Oligocene ice sheets on East Antarctica (Mildenhall, 1989). IRD abundances increased in the middle Miocene in the Antarctic (Margolis, 1975; Kennett and Barker, 1990) and Subantarctic (Warnke et al., 1992), although major increases did not occur until the late Miocene. The available IRD data is consistent with the suggestion that the middle Miocene represents a major step in the development of a large, "cold" ice sheet. The glacigenic sedimentation history in the southern sector of the Indian Ocean supports a major increase in East Antarctic ice volume during the middle Miocene (Ehrmann and Mackensen, 1991).

Oxygen isotopic records are critical in providing further perspective on EAIS development during the middle Miocene. A large increase in benthic foraminiferal $\delta^{18}O$ of ~1.0–1.3‰ occurred from ~16 to 12.5 Ma (Shackleton and Kennett, 1975; Miller et al., 1987). The proportion of the $\delta^{18}O$ increase due to EAIS growth, versus that due to deep water cooling during the middle Miocene, is a long-standing question. This global benthic increase has been inferred to reflect some combination of EAIS growth and deep water cooling (Savin, 1975; Woodruff and Savin, 1989; Miller et al., 1991; Zachos et al., 1992; Wright et al., 1992), although some workers have ascribed the increase entirely to deep water cooling (Matthews and Poore, 1980; Prentice and Matthews, 1988, 1991).

Covariance between benthic and planktonic oxygen isotopic records from non-upwelling, low-latitude locations is generally accepted as the best indication of $\delta^{18}O$ variations due to ice volume because of the relative stability of surface water temperatures in these regions (Miller et al., 1987, 1991). Three successive middle Miocene increases in benthic foraminiferal $\delta^{18}O$ from ~16.1 to 15.5, 14.5–13.6, and 13.2–12 Ma are accompanied by planktonic $\delta^{18}O$ increases (Miller et al., 1991; Wright et al., 1992), indicating three growth phases of the East Antarctic ice sheet. High-resolution benthic and planktonic oxygen isotopic records from Site 588A in the low-latitude, western Pacific (Fig. 2; Flower and Kennett, 1993c) provide detailed isotopic records relevant to EAIS variations from ~16 to 12 Ma, including the latter two growth phases between 14.5 and 12.4 Ma. Coincident $\delta^{18}O$ increases of from 0.7‰ and 0.3‰, respectively, in benthic and planktonic records

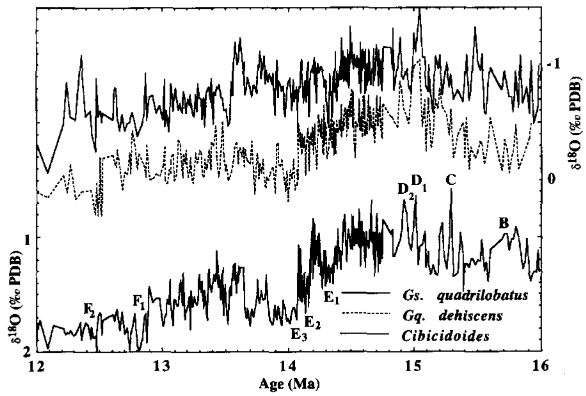

Fig. 2. Oxygen isotopic records vs. age for DSDP Site 588A from the southwest Pacific based on the benthic foraminifera *Cibicidoides wuellerstorfi* or *C. kullenbergi* and the planktonic foraminifera *Globigerinoides quadrilobatus* and *Globoquadrina dehiscens*. Oxygen isotopic events in the nomenclature of Woodruff and Savin (1991) are labelled. The age model is derived from an integrated isotopic biochronology (Flower and Kennett, 1993c).

from 14.5 to 14.1 Ma and from 12.9 to 12.4 Ma (Fig. 2), suggests EAIS growth at these times.

The $\delta^{18}O$ increase from 14.5 to 14.1 Ma is the largest benthic $\delta^{18}O$ increase and exhibits the greatest benthic-planktonic $\delta^{18}O$ covariance in the middle Miocene. We infer that the largest semi-permanent EAIS increase occurred from 14.5 to 14.1 Ma. Estimates of the magnitude of sea level fall related to EAIS growth range from ~50 m (Haq et al., 1987) to a maximum of ~65–130 m (Miller et al., 1987). The latter estimate is based on an increase in the $\delta^{18}O$ of seawater of 0.7‰ from 14.5 to 14.1 Ma and a calibration range of 0.055–0.11‰ per 10 m sea level change (Miller et al., 1987). Because benthic foraminiferal $\delta^{18}O$ increases were up to 1.3‰ (e.g., Woodruff and Savin, 1991), at least 0.6‰ of the benthic $\delta^{18}O$ increase must therefore be due to simultaneous deep water cooling of ~2.5°C.

2.2.1. (In)stability history of the EAIS

Recent high-resolution benthic oxygen isotopic records have refined knowledge of early to middle Miocene $\delta^{18}O$ history. Compilations can be found in Woodruff and Savin (1989, 1991) and Wright et al. (1991), plus our own high-resolution records from Site 588A (Flower and Kennett, 1993c). These records suggest a significant change in variability about 14.8 Ma (Fig. 2). From a Neogene minimum of ~0.5‰ in the late early Miocene at 16.5 Ma (Kennett, 1986; Miller et al., 1991), $\delta^{18}O$ values fluctuated strongly by 0.5–1.0‰ from ~16 to 14.8 Ma, including distinct minima at 16.6, 15.7, 15.3 and 14.9 Ma. This interval of large fluctuations in $\delta^{18}O$ ended with a permanent increase from 14.5 to 14.1 Ma. The $\delta^{18}O$ minimum at 14.9 Ma ($\delta^{18}O$ event D of Woodruff and Savin, 1991) was the last of the early to middle Miocene minima, and may represent a pivotal point in middle Miocene climate change.

Decreased variability in high-resolution oxygen isotopic records from 14.5 to 14.1 Ma suggest Antarctic cryospheric development resulted in increased stability of the EAIS. To the extent that the strong variations in the $\delta^{18}O$ signal in the early to middle Miocene reflect ice volume, ice sheet fluctuations must have been greater during the interval from ~16.6 to 14.9 Ma compared with from 14.9 to 12 Ma. Instability of the Antarctic cryosphere during the late early to early middle Miocene may have preceded major EAIS growth during the early middle Miocene (Miller et al., 1987). EAIS ice growth from 14.5 to 14.1 Ma may have resulted in a more stable EAIS ice sheet, and a corresponding decrease in variability of the climate system.

Support for a semi-permanent increase in East Antarctic ice sheet volume during the middle Miocene comes from geomorphological information from the Transantarctic Mountains. Ashfalls dated to 14 Ma are preserved above the polar desert pavements in the Dry Valleys (Sugden, 1992; Marchant, 1992; Marchant et al., 1993). The remarkable preservation of these ashfalls in a region susceptible to erosion during deglaciation suggests that Antarctic temperatures have remained close to their present levels and that the East Antarctic ice sheet has been a semi-permanent feature during the last ~14.5–14.1 m.y.

2.2.2. Middle Miocene sea levels

The early to middle Miocene global sea level curve supports the scenario of major EAIS variations near the late early/middle Miocene boundary followed by semi-permanent increases from 14.5 to 14.1 Ma and 12.9 to 12.4 Ma. Good correspondence exists between the sea level record for the middle Miocene (Haq et al., 1987) and benthic $\delta^{18}O$ records. Middle Miocene sea levels are marked by large, short-term variations from ~16 to 14 Ma followed by a permanent global sea level fall at 14.2 Ma (Fig. 3; Haq et al., 1987). Early middle Miocene sea levels from ~17 to 14 Ma feature a series of large fluctuations of ~50 m, including three highstands separated by two sea level falls. Because of the poor stratigraphic resolution of the sea level curve, it is not yet possible to correlate these fluctuations with confidence to the oxygen isotopic record. Clearly there was a period of instability in sea level during an interval of generally high sea levels from ~17 to 14 Ma, just prior to polar cooling and global sea level fall at ~14.2 Ma. Increased stability of the EAIS during a major growth phase involving global sea level fall near 14.2 Ma may have resulted in a more stable state of the ocean/climate system. The

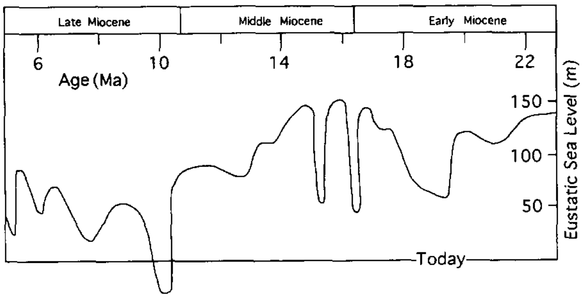

Fig. 3. Miocene sea level record relative to the present day (from Haq et al., 1987). Note the large fluctuations of >50 m from ~16.5 to 14.5 Ma, and the global eustatic sea level fall in two steps at 14.2 and 12.9 Ma. A major fall of >50 m also occurred ~10.5 Ma.

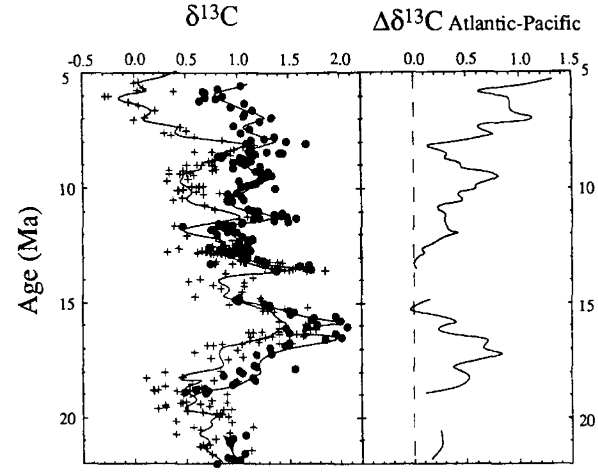

Fig. 4. Comparison of carbon isotopic composition of Atlantic (●) and Pacific (+) deep waters through the Miocene (from Wright et al., 1992). Also shown is the Atlantic–Pacific difference in $\delta^{13}C$. Note the interval of similar $\delta^{13}C$ values during the middle Miocene from ~16 to 12.5 Ma.

timing, magnitude and semi-permanence of this sea level fall suggest it corresponds to the rapid benthic foraminiferal $\delta^{18}O$ increase between 14.5 and 14.1 Ma. Middle Miocene sea levels reached a low at about 12.5 Ma, near the middle Miocene maximum in $\delta^{18}O$ at 12.8 Ma.

2.3. Role of deep water circulation in Antarctic climatic evolution

In this section we discuss relations between middle Miocene climatic and EAIS variations and deep water circulation history. Miocene global deep water circulation has been examined in several recent contributions (Woodruff and Savin, 1989, 1991; Wright et al., 1992). Woodruff and Savin (1989) presented carbon isotopic evidence for the presence of a low-latitude, high-salinity deep water source in the early to early middle Miocene Indian Ocean: Tethyan Indian Saline Water (TISW). This warm deep water source was terminated by the closure of the eastern portal of the Tethys Ocean in the eastern Mediterranean region during the middle Miocene (Woodruff and Savin, 1989). Northern Component Water flux was low during the early middle Miocene from ~16 to 12.5 Ma and increased at about 12.5 Ma (Schnitker, 1980; Woodruff and Savin, 1989, 1991; Wright et al., 1992). However, it is likely that Northern Component Water existed during the early Miocene (from ~20 to 16 Ma; Wright et al., 1992), based on the carbon isotopic difference between the Atlantic and Pacific Oceans (Fig. 4), and on the presence of early Miocene sediment reflectors in the North Atlantic inferred to have been produced by strong bottom water currents (Miller et al., 1987).

Southern Component Water production clearly increased in strength during the middle Miocene from ~16 to 12.8 Ma (Woodruff and Savin, 1989, 1991; Wright et al., 1992). Southern Component Water sources increased as Northern Component Water and Tethyan sources decreased, perhaps accounting, in part, for the disappearance of strong inter-basinal $\delta^{13}C$ gradients between the Atlantic and Pacific (Fig. 4). Coincident increases during the middle Miocene in the strength of Southern Component Water and in growth of the EAIS suggest a complementary relationship and illustrate the expansion of Antarctic influence on global climates through deep water circulation.

2.3.1. Deep ocean circulation models for EAIS growth

One group of hypotheses for the cause of middle Miocene polar cooling and Antarctic ice sheet growth suggests control by deep water circulation. These hypotheses must explain both the substantial increase in the planetary thermal gradient and increased growth of the EAIS. Deep water circulation patterns may have caused increased middle Miocene EAIS growth, through increased moisture supply to the Antarctic or through decreased poleward heat transport. Some workers have suggested that increased supply of moisture by warm, deep waters upwelling in the Antarctic directly caused an increase in the Antarctic ice sheet. Schnitker (1980) suggested that the initiation during the middle Miocene of important upwelling of Northern Component Water at the Antarctic Divergence zone may have contributed moisture to accelerate EAIS growth through increased precipitation. Prentice and Matthews (1988, 1991) suggested that supply of warm, saline deep water to the Antarctic fostered EAIS growth throughout the Tertiary.

Alternatively, changes in the meridional supply of heat to the high latitudes may have controlled EAIS variations through its effect on polar surface temperatures. Based on reconstructions of the carbon isotopic distribution in the world ocean, Woodruff and Savin (1989) inferred that Tethyan Indian Saline Water (TISW) upwelled in the Antarctic region throughout the early Miocene. Meridional transport of heat by this deep water source may have prevented large-scale growth of the EAIS through the early Miocene until the middle Miocene ~14 Ma (Woodruff and Savin, 1989). The termination, during the middle Miocene, of Tethyan Indian Saline Water (TISW) may have led to further cooling of Antarctic surface waters and, in conjunction with increased Southern Component Water production, fostered middle Miocene EAIS growth (Woodruff and Savin, 1989). The termination of a Tethyan source of warm deep water to the Antarctic ~14 Ma corresponds more closely to the time of major EAIS growth from 14.5 to 14.0 Ma (Flower and Kennett, 1993a; in review) than increased Northern Component Water production ~12.5 Ma. Therefore, it seems that decreased meridional heat transport by TISW was a more important control of EAIS growth than moisture supply by Northern Component Water during the middle Miocene.

2.3.2. *Evidence for warm, saline deep water in the southwest Pacific*

Oxygen isotopic records from a depth transect of Deep Sea Drilling Project sites at 35°S provide independent evidence for warm, saline deep water influx into the southwest Pacific in the early middle Miocene (Flower and Kennett, 1993a; in review). Benthic foraminiferal oxygen isotopic data from a depth transect of drilled cores from the Lord Howe Rise to the New Caledonia Basin exhibit an interval of reduced and highly variable vertical $\delta^{18}O$ gradients during the early middle Miocene from ~16 to 14.9 Ma (Fig. 5). During this interval, a reversal in the vertical $\delta^{18}O$ gradient with depth between Sites 590B and 591B (1200 and 2100 m paleodepth, respectively) occurred from 15.9 to 15.6 Ma (Fig. 5). Deep waters are inferred to have been warmer and more saline than intermediate waters in the southwest Pacific during this relatively warm interval ~15.7 Ma (Flower and Kennett, 1993a; in review).

The reduced vertical $\delta^{18}O$ gradient and inversion suggest that warm, saline deep water derived from the Tethyan region was competing with cold Southern Component Water sources in the southwest Pacific during the interval from ~16 to 14.9 Ma, just prior to major EAIS growth from 14.5 to 14.1 Ma. Vertical deep water temperature gradi-

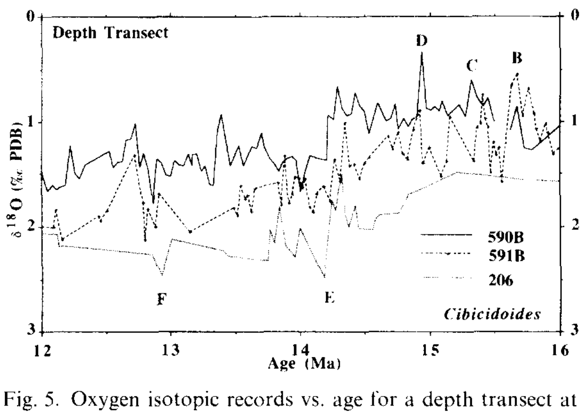

Fig. 5. Oxygen isotopic records vs. age for a depth transect at 1200, 2100, and 3100 m at 35°S paleolatitude formed by DSDP Sites 590B, 591B and 206 based on the benthic foraminifera *Cibicidoides*. Oxygen isotopic events B to F of Woodruff and Savin (1991) are labelled. The age model for each site is derived from an integrated isotopic biochronology (Flower and Kennett, in press). A brief reversal in the oxygen isotopic gradient from 15.9 to 15.6 Ma may indicate the influence of warm, saline deep water at Site 591B. An interval of large variations in strength of high- and low-latitude deep water sources is inferred from ~16 to 14.9 Ma, which precedes the rapid increase at ~14.2 Ma (Event E) considered to reflect East Antarctic ice growth.

ents probably lessened during warm intervals ~15.7, 15.3 and 14.9 Ma, due to episodic increases in strength of warm, low-latitude sources relative to Southern Component Water sources. Competition between these deep water sources may also account for the variability in benthic $\delta^{18}O$ records observed. Increased meridional heat transport by deep waters upwelling in the Antarctic during these intervals led to instability of the EAIS and inhibited EAIS growth. Correspondingly, decreased meridional heat transport by TISW and increased strength of Southern Component Water fostered EAIS growth after 14.9 Ma. In short, oxygen isotopic data from a depth transect in the southwest Pacific supports the suggestion (Woodruff and Savin, 1989) that decreased supply of warm saline deep water derived from the Tethyan region fostered EAIS growth from 14.5 to 14.1 Ma.

The presence of Tethyan deep water in the southwest Pacific in the early middle Miocene suggests that Tethyan deep water in the southwest Pacific in the early middle Miocene suggests that Tethyan sources continued to introduce warm, saline deep waters to the world ocean in the early Neogene, long after their inferred maxima during portions of the middle Eocene and Oligocene (Kennett and Stott, 1990). The interval from ~16 to 14.9 Ma may represent the last strong influence of warm saline deep water derived from the eastern Tethys Ocean.

Cenozoic climatic and cryospheric developments are ultimately tied to plate tectonic changes, in particular, through the influence of the isolation of Antarctica upon Southern Ocean circulation and perhaps the episodic closure of the eastern portal of the Tethys Ocean upon low-latitude warm saline deep water formation. Early middle Miocene warm, saline deep water production may have been strongly influenced by the collision of Africa with Eurasia. Separation of the eastern Mediterranean and Paratethys Seas from the northern Indian Ocean would have decreased the contribution of warm, saline deep waters to the world ocean. Paleogeographic reconstructions (Hsu, 1978; Hsu and Bernoulli, 1978) and African–Eurasian mammal dispersal patterns (Steininger et al., 1985) and plankton distributions (Adams et al., 1983; Zachariasse, 1983) indicate the establishment of land bridges linking Arabia and Eurasia during two intervals in the late early and early middle Miocene, at 18 ± 1 Ma and at 15 ± 1 Ma (Thomas, 1985). In the intervening interval represented by the Langhian Stage, the widespread evaporative eastern Mediterranean and Paratethys Seas may have contributed to increased production of warm, saline deep waters through the eastern Mediterranean/Indo-Pacific seaway. Plate tectonic developments in the eastern Tethyan region controlling the formation of evaporative basins and the final closure of the eastern Mediterranean may have produced a final pulse of warm, saline deep water supply to the Indo-Pacific during the early middle Miocene.

2.3.4. *Benthic foraminiferal faunal change*

Faunal changes in benthic foraminifera in the early to middle Miocene are relevant to the history of paleoenvironmental change in deep water. One of three Cenozoic intervals of evolutionary turnover in benthic foraminifera occurred during the late early to early middle Miocene from ~17 and 13.5 Ma (Thomas, 1985; Woodruff, 1985; Miller and Katz, 1987; Thomas and Vincent, 1987). Turnover of approximately 20% of the benthic fauna began prior to, and continued through, middle Miocene deep ocean cooling from 14.5 to 14.1 Ma. Benthic foraminiferal faunas changed first at lower bathyal to abyssal depths, and later at middle bathyal depths as species migrated to shallower depths (Woodruff, 1985; Woodruff and Savin, 1989). These changes coincided with the early to middle Miocene carbon isotopic excursion, and may have resulted from changes in bottom water nutrient concentrations and corrosivity in association with circulation changes (Woodruff, 1985; Miller and Katz, 1987; Woodruff and Savin, 1989).

Middle bathyal benthic foraminiferal faunas in the southwest Pacific changed at the end of this interval of deep-water benthic turnover. Kurihara and Kennett (1992) noted that the establishment of Neogene deep water assemblages dominated by *Epistominella exigua* coincided with the middle Miocene $\delta^{18}O$ increase from 14.5 to 14.1 Ma and suggested that those faunas responded to deep water cooling in association with EAIS growth during this interval. The establishment of this cold, deep water assemblage, which persisted through the late Neogene, reinforces the inference of moderate deep water cooling from 14.5 to 14.1 Ma, associated with major EAIS growth.

2.4. Global carbon cycling in the early middle Miocene

Middle Miocene carbon isotopic records provide evidence for large-scale transfers of carbon between major carbon reservoirs, which potentially had an important effect on global climates. Large changes in the mean ocean $\delta^{13}C$ composition as recorded by benthic foraminiferal calcite occurred from ~17 to 13.5 Ma. Six discrete $\delta^{13}C$ maxima reaching ~2.0‰ occurred during the late early to early middle Miocene (Fig. 6). Strong covariance between benthic and planktonic $\delta^{13}C$ records confirms that these variations represent reservoir changes in global carbon cycling (Flower and Kennett, 1993c).

Mean ocean $\delta^{13}C$ is primarily controlled by the proportion of carbon being stored as organic carbon in terrestrial and marine marginal basin reservoirs relative to calcium carbonate sediments in the deep sea (Vincent and Berger, 1985; Miller

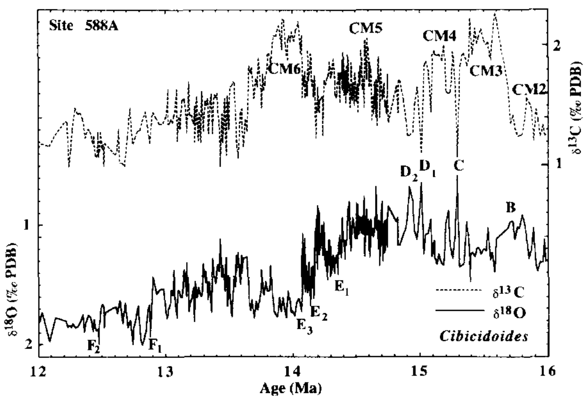

Fig. 6. Oxygen and carbon isotopic records vs. age for DSDP Site 588A from the southwest Pacific based on the benthic foraminifera *Cibicidoides wuellerstorfi* or *C. kullenbergi*. Events of Woodruff and Savin (1991) are labelled *2–6*, *B–F*. The age model is derived from an integrated isotopic biochronology (Flower and Kennett, 1993c). Note the covariance between $\delta^{18}O$ and $\delta^{13}C$ records from ~16 to 13.7 Ma.

and Fairbanks, 1985). Deposition of organic carbon enriched in ^{12}C raises mean ocean $\delta^{13}C$ and lowers ΣCO_2, increasing the carbonate ion content of the deep ocean. Intervals of high organic carbon accumulation in marginal basin sediments, therefore, are marked by $\delta^{13}C$ maxima in the deep sea record, and by increased carbonate preservation. Conversely, transfer of organic carbon to the deep ocean lowers mean ocean $\delta^{13}C$ and raises ΣCO_2, resulting in increased dissolution and reduced carbonate preservation. Mean ocean $\delta^{13}C$ maxima have been shown to coincide closely with increases in carbonate preservation (Woodruff and Savin, 1991).

The discrete maxima observed in benthic $\delta^{13}C$ records may reflect episodic increases in organic carbon deposition in marginal marine basins such as the Monterey Formation (Vincent and Berger, 1985; Flower and Kennett, 1993b) and the phosphatic deposits of the southeastern United States (Compton et al., 1990, 1993). Inferred episodic increases in coastal productivity may be related to periodic insolation changes, as the six known $\delta^{13}C$ maxima have a period of ~400 kyr (Woodruff and Savin, 1991). Of primary importance in assessing the impact on the carbon cycle is the relation of such periodic organic carbon deposition to global cooling through positive feedback mechanisms involving atmospheric partial CO_2 drawdown.

2.4.1. Relationship between $\delta^{18}O$ and $\delta^{13}C$

A positive correlation between $\delta^{18}O$ and $\delta^{13}C$ has been observed in the early to middle Miocene from 16 to 13.7 Ma (Fig. 6), ending at the termination of the Monterey Carbon Excursion (Woodruff and Savin, 1991; Flower and Kennett, 1993c). Individual $\delta^{13}C$ maxima within the early to middle Miocene global $\delta^{13}C$ excursion were coincident with $\delta^{13}O$ increases inferred to involve deep water cooling and EAIS growth. Each of the $\delta^{18}O$ minima also coincides with a $\delta^{13}C$ minimum.

Covariance between benthic $\delta^{18}O$ and $\delta^{13}C$ indicates that global climate was closely linked to marine organic carbon burial from ~16 to 13.7 Ma. Episodic invigoration of ocean circulation in conjunction with polar cooling may have increased global ocean productivity and burial of organic carbon-rich sediment, thereby drawing down atmospheric partial CO_2. Although generally high sea levels in the early to early middle Miocene may have increased the area for organic carbon deposition, according to the original formulation of the Monterey Hypothesis, enhanced carbon burial producing $\delta^{13}C$ maxima must have occurred during relative lowstands of sea level marked by higher $\delta^{18}O$ values. This hypothesis suggests that global cooling from 15.6 to 15.4 Ma and from 14.5 to 14.0 Ma was accelerated by episodic organic carbon burial.

2.4.2. Polar cooling accelerated by pCO_2 drawdown

The 14.5–14.1 Ma oxygen isotopic increase inferred to reflect major EAIS growth was superimposed on a general global cooling trend from ~16 Ma which clearly reached a threshold at 14.2 Ma. Drawdown of atmospheric pCO_2 may have contributed to an acceleration of polar cooling and EAIS growth at this time. Attempts to assess the magnitude of potential middle Miocene global cooling by estimating changing concentrations of greenhouse gases such as CO_2 have been promising but inconclusive (Arthur et al., 1991).

The broad early to middle Miocene positive $\delta^{13}C$ excursion coincides with the deposition of large quantities of organic carbon in marginal basin sediments around the Pacific Rim, including the Monterey Formation of California (Vincent and Berger, 1985) and the phosphate deposits of the southeastern United States (Compton et al., 1990, 1993). Sequestering of carbon in the Monterey Formation and correlative sediments may have led to global cooling through drawdown of atmospheric partial CO_2 by a series of positive feedback mechanism, termed the "Monterey Hypothesis". This hypothesis suggested that initiation of atmospheric partial CO_2 drawdown at ~17 Ma would have reached a threshold necessary for global cooling ~2–3 m.y. later, accounting for the lag before major cooling at ~14.5–14.1 Ma.

On the other hand, recent oxygen and carbon isotopic data from Monterey Formation sediments at Naples Beach, California (Flower and Kennett, 1993b) demonstrates that enhanced organic carbon deposition coincided with the $\delta^{18}O$ increase dated in the deep-sea record from 14.5 to 14.1 Ma. Because this increase coincided with the initiation of $\delta^{13}C$ maximum 6 (14.2–14.0 Ma), organic carbon-rich deposition in the Monterey Formation contributed both to the generation of $\delta^{13}C$ maxima and to synchronous global cooling. This result suggests that episodic organic carbon burial accelerated global cooling from 14.5 to 14.1 Ma through atmospheric partial CO_2 drawdown.

2.5. Global ocean sediment depositional patterns

Biosiliceous and carbonate sediment deposition patterns were closely related to deep ocean circulation changes in the early to middle Miocene. A redistribution of biogenic silica and carbonate deposition in the global ocean between ~18 and 12 Ma involved transfer of the locus of biosiliceous deposition from the North Atlantic Ocean to the North Pacific and Indian Oceans (termed the "silica switch"; Fig. 7; Keller and Barron, 1983; Woodruff and Savin, 1989; Barron and Baldauf, 1990). Significantly, the transition included an interval of comparable biosiliceous deposition in each of these oceans from ~14 to 11 Ma (Woodruff and Savin, 1989).

The "silica switch" must be related to large-scale changes in deep water circulation, which controlled deep water aging and nutrient distribution in the world ocean and, thereby, the locus of

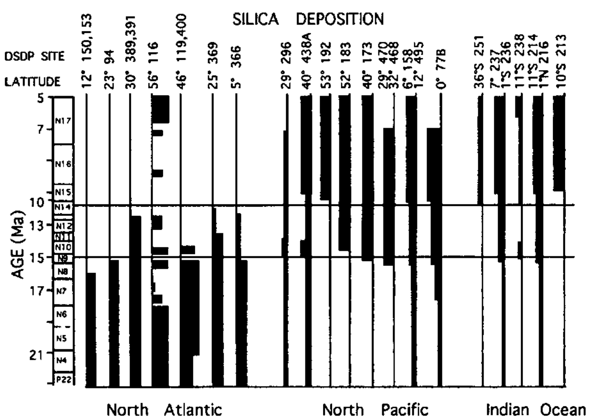

Fig. 7. The global stratigraphic and geographic distribution of biosiliceous sediments in the Miocene. A shift in the locus of deposition from the Atlantic Ocean to the Pacific and Indian Oceans occurred from ~18 to 12 Ma, the so-called "silica shift."

carbonate and silica deposition. In a lagoonal-style circulation typified by the modern North Atlantic, young deep water with low silica and high carbonate ion concentrations (enhancing carbonate preservation) flows south at depth and is resupplied by surface waters. Conversely, the North Pacific exhibits an estuarine style, where old deep waters rich in nutrients including silica (thus corrosive to carbonate) flow north and upwell (fostering biosiliceous productivity), then generally flow south. The "silica switch" may be generally explained by transitions in the North Atlantic and North Pacific from one type of circulation to the other (Keller and Barron, 1983).

The "silica switch" broadly coincides with the initiation of a strong Northern Component Water source in the North Atlantic (Barron and Baldauf, 1990; Wright et al., 1992). Initiation of a strong source of young deep water enriched in the carbonate ion may have allowed the North Atlantic to become a carbonate ocean, as it remained throughout the Neogene. The consequent establishment of a conveyor-style circulation system for the global ocean would have aged Pacific deep waters, fostering increased silica deposition and preservation. Evidence from the $\delta^{13}C$ contrast between the Atlantic and Pacific oceans through the Cenozoic suggests deep waters aged towards the Pacific through the Miocene, except between ~16 and 12.5 Ma (Fig. 4; Wright et al., 1992). During the middle Miocene transition interval from ~14 to 11 Ma, silica deposition was distributed throughout the Atlantic, Pacific and Indian Ocean basins (Fig. 7). Deep waters must have aged in a northward direction during this interval, reflecting an increased contribution from Southern Ocean sources (Woodruff and Savin, 1989). Early Miocene silica distribution patterns are consistent with a strong Tethyan/Indian deep water source, except that greater biosiliceous sedimentation is expected than was observed in the North Pacific (Woodruff and Savin, 1989), and a high $\delta^{13}C$ deep water source was present in the North Atlantic (Wright et al., 1992).

2.6. Global effects of Antarctic cooling and ice sheet growth

Middle Miocene polar cooling and East Antarctic ice growth had major effects on global terrestrial environments. The increased meridional thermal gradient, fostered by high-latitude cooling, had major effects on atmospheric circulation, probably including contraction and intensification of the Hadley circulation (Flohn, 1978). Meridional restriction and invigoration of ocean-atmospheric circulation led to the development of more sharpely defined climatic belts and faunal provinces, reflected in both open-ocean and continental records.

2.6.1. Increased aridity in the mid-latitudes

Evidence for the intensification of the arid mid-latitude climatic zone related to restricted Hadley cell atmospheric circulation comes from enhanced development of grasslands and associated biota on all of the major continents (Berggren and Van Couvering, 1974). For example, the early to middle Miocene record of Patagonian mammals provides a clear view of sharp paleoenvironmental changes in South America. Humid subtropical forest environments, which had persisted to the southern tip of South America through the Paleogene and early Miocene, abruptly shifted to colder, drier grassland environments in the middle Miocene. Middle Miocene mammals at the beginning of the Panaraucanian Faunistic Cycle (middle to early Pliocene, ~14–4 Ma) indicate that subtropical to tropical environments shifted from southern to northern Patagonia (~30°S) (Pascual and Jaureguizar, 1990). Many taxa related to subtropical woodlands (including platyrrhine monkeys, anteater-like armadillos and certain marsupials) disappeared from Patagonia. The extinction of primitive tree sloths coincided with an increase in ground sloths (Webb, 1985). The development of drier, more widespread plains and pampas favored high-crowned ungulates, which for the first time became the most abundant mammals in South America (Pascual and Ordreman Rivas, 1971; Pascual, 1984).

The early Cenozoic tropical forests of East Africa appear to have been partly replaced by woodland–grassland habitats during the middle Miocene (Andrews and Van Couvering, 1975; Bonnefille, 1984). The middle Miocene development of drier grassland habitats at the Fort Ternan site, Kenya, is the earliest yet documented in Africa during the Cenozoic (Retallack, 1990, 1992). The appearance of grasses (Dugas and Retallack, 1993) was also associated with that of a related mammalian fauna exhibiting toothwear patterns characteristic of grassland grazers (Shipman et al., 1981). However, the nature of the vegetation and paleoenvironment at Fort Ternan is currently debated in light of carbon isotopic results on pedogenic soil carbonate suggesting a contribution either from woodland vegetation or from primitive grasses similar to modern African grasslands at 2000–3000 m (Cerling et al., 1991).

The development of grasslands in East Africa fostered a major faunal turnover in the middle Miocene (Retallack, 1991). The mammalian assemblages abruptly became antelope- and giraffe-dominated in East Africa at about 15 Ma, whereas snail assemblages retained many woodland elements (Pickford, 1985). A more pervasive grassland-adapted mammalian fauna marked by hipparionine horses appeared in the late Miocene, from 9 to 6 Ma in East Africa (Van Couvering, 1980), reflecting the continued development of savannah-like conditions.

In North America, the Great Plains became mixed woodland–grassland in the middle Miocene, due to decreasing precipitation (Webb, 1977; Axelrod, 1985; Wolfe, 1985; Leopold and Denton, 1987). Precipitation decreased to $<35\%$ of the annual total at ~ 15 Ma, as broad-leaved deciduous hardwood floras were eliminated from western Nevada, then in Oregon and Washington (Axelrod, 1992). Following a maximum in equid diversity near the early/middle Miocene boundary (MacFadden and Hulbert, 1988; 1991), mesodont (moderate-height teeth) horses were largely replaced by more hyposdont (high-crowned teeth) horses between 15 and 12 Ma reflecting increased grassland vegetation (Hulbert, 1993). Mirroring the pattern seen in South American mammals, grazers diversified and thrived as browsers declined in conjunction with the increased development of grasslands. Similarly, the aridification of parts of Asia was accompanied by major mammalian faunal turnover in the early to middle Miocene from ~ 20 to 13.5 Ma. Increased diversity of bovid and rodent faunas is known from the Siwalik formations of Pakistan (Barry et al., 1985).

Southern South African climates also underwent increased aridification toward the present-day hyperarid conditions of the Namib and Kalahari Deserts. The best record of these changes is provided by eolian dust flux records from deep-sea cores in the Indian Ocean (Hovan and Rea, 1992). Stepwise changes in continental climates toward hyperarid conditions are indicated by a *decrease* in eolian dust flux. A step-wise reduction in eolian flux is seen through the Cenozoic, with major decreases observed during the latest Eocene, during the Oligocene and in the middle Miocene. Extremely low dust flux to the Indian Ocean since the middle Miocene is inferred to reflect the development of hyperarid conditions in the Namib Desert of western South Africa about 13 Ma (Hovan and Rea, 1992).

Although Archer (1993) has shown no change in the middle Miocene toward grazers in marsupial assemblages, abundant evidence for the contemporaneous aridification of Australia comes from the development of grassland floras (Kemp, 1978) and from accumulation in deep sea sediments of biogenic material derived from upwind grassland areas. A detailed evolution of middle Miocene Australian climate history is provided by the influx rates and mineralogy of terrestrial clays in southwest Pacific Deep Sea Drilling Project sites (Robert et al., 1986; Stein and Robert, 1986). The aridification of Australia was controlled by both the northward drift of the Indo-Australian plate and Antarctic cooling and ice sheet growth. Aridification advanced southward, as Australia drifted northward into an intensifying mid-latitude high pressure belt. The first occurrence of phytoliths (small particles of opaline silica secreted by grasses in their epidermal cells) in the same Tasman Sea cores in the middle Miocene at about 14 Ma (Locker and Martini, 1989) provides strong evidence for grassland development. The high open-ocean accumulation rate of phytoliths may also relate to increased intensity of wind systems conducive to phytolith transport (Locker and Martini, 1989).

Antarctic cryospheric development and associated deep water circulation changes may also have fostered aridification by supplying cool intermediate waters to mid-latitude surface waters in upwelling regions, resulting in decreased moisture levels in the atmosphere. The origin of the Namib Desert in the middle Miocene may be related to increased production of Antarctic intermediate waters that upwelled along the west African coast (Seisser, 1978; Van Zinderen Bakker, 1975). The general increase in upwelling strength that accompanied middle Miocene paleoceanographic development may have fostered increased diversity of Alcids (diving seabirds) (Warheit, 1992), whose Neogene evolution seems to be closely tied to the development of productive upwelling systems. Thus the increased development of the Antarctic cryosphere and associated deep water production had a multitude of effects on middle Miocene environmental and biotic evolution, through its strong influence on ocean–atmosphere circulation.

2.7. *Asynchroneity of Cenozoic polar cryospheric development*

The long-term history of Antarctic cryospheric development since the early Oligocene or perhaps the late Eocene (Kennett and Barker, 1990; Barrett et al., 1989; Barron et al., 1991) contrasts sharply with the late Pliocene to Quaternary development of Northern Hemisphere glaciations (Shackleton et al., 1984; Raymo et al., 1989). This asynchroneity fundamentally results from the absence of a polar continent in the Northern Hemisphere, on which a permanent continental ice sheet can be supported and preserved (Thiede et al., 1989). It is important to note that the only permanent Northern Hemisphere ice sheet during the Quaternary is the relatively small Greenland ice sheet; the Northern Hemisphere Quaternary glaciations have otherwise seen the accumulation of large expansive, but temporary continental ice sheets. Nevertheless, the history of Antarctic ice sheet growth can provide insights into the variable state of the Northern Hemisphere ice sheets during the Quaternary.

The early middle Miocene evolution of the Antarctic cryosphere from ~14.8 to 14.0 Ma provides an useful comparison with the late Pliocene/Quaternary onset of Northern Hemisphere ice sheets from ~3 to 2.5 Ma. The late early to early middle Miocene (~16–14.8 Ma) variability in the climate system was accompanied by large fluctuations in sea level and was followed by a series of cooling steps leading to a more stable state. Similarly, the onset of Northern Hemisphere ice sheets was marked by an increase in the amplitude of climatic variations, which have continued through the Quaternary. This instability may be a Quaternary analog for the early middle Miocene history of the East Antarctic ice sheet. The instability of the Northern Hemisphere ice sheets even during the late Quaternary may be related to their location at lower latitudes, and the absence of large land masses in the polar region.

The middle Miocene and late Pliocene/early Quaternary intervals of polar ice sheet growth are also marked by relatively low fluxes of relatively warm and saline Northern Component Water. Decreased Northern Component Water flux was associated with bipolar cooling and cryospheric development during the late Pliocene/early Quaternary, including the onset of Northern Hemisphere ice sheets (Raymo et al., 1992). Similarly, Northern Component Water and Tethyan Indian Saline Water flux decreased and Southern Component Water production increased during the middle Miocene (Woodruff and Savin, 1989, 1991; Wright et al., 1992), in conjunction with a major growth phase of the East Antarctic ice sheet. To the extent that the deep ocean thermohaline circulation driven by Northern Component Water has acted as a heat pump for the high latitudes as in the Quaternary (Broecker and Denton, 1990), intervals of low Northern Component Water and high Southern Component Water flux may have fostered cooler polar climates and increased ice sheet development during at least two Neogene intervals of major cryospheric development.

These speculations are consistent with suggestions that deep water circulation has played a major role in the evolution of Neogene climates. Deep water circulation patterns, in combination with tectonic changes (including those that isolated the Antarctic continent and controlled the Tethyan/Indo Pacific seaway), help explain both the marked asynchroneity between Northern and Southern Hemisphere ice sheet development, and the rapid steps that punctuate Cenozoic climatic evolution.

Acknowledgements

This contribution was supported by National Science Foundation grants DPP92-18720 and EAR92-18738 to JPK, and by a Geological Society of America Travel Grant to the 1992 International Geological Congress in Kyoto, Japan to BPF. We thank T. Cronin, J. Barron and B. Tiffney for constructive reviews.

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