The Role of the Tropical Oceans on Global Climate During a Warm Period and a Major Climate Transition

BY ANA CHRISTINA RAVELO AND MICHAEL WILLIAM WARA

Paleoceanographic records extracted from a global array of sediment cores obtained by the Ocean Drilling Program (ODP) can be used to elucidate differences between oceanographic conditions during the early Pliocene warm period (~4.5 to 3.0 million years ago [Ma]) and the late Pliocene and Pleistocene cool ice age period (3.0 Ma to present). Oxygen isotope gradients derived by laboratory analysis of calcareous microfossil shells from low-latitude sites are used to reconstruct tropical surface hydrographic (i.e., temperature and/or salinity) gradients and to examine the role of tropical oceans on global climate over the last 5 million years, including the factors that caused the warm to cold climate transition, commonly referred to as the onset of significant Northern Hemisphere Glaciation (NHG). We find that a small west-east temperature gradient across the Pacific Ocean, similar to El Niño conditions, accompanied and perhaps played a critical role in determining early Pliocene global warmth; steeper temperature gradients, more typical of the modern ocean, were established during the cool, ice-age climatic state by ~1.5 Ma.

What caused the end of El Niño-like conditions and the onset of the ice ages? We show that changes in the oxygen isotope gradient between the Indian and Pacific Oceans occurred between ~3.0 and 1.5 Ma, indicating that gradual tectonic influences on flow through the Indonesian seaways may have caused changes in tropical sea surface temperature patterns that forced NHG. The marked increase in the salinity gradient between the Pacific and Atlantic Oceans, possibly related to restriction of flow through the Panamanian seaway, occurred ~4.2 Ma, too early to be responsible for the onset of NHG. Other sources of gradual global climate cooling through the Pliocene are also discussed.

The Ocean Drilling Program (ODP) drillship *JOIDES Resolution* sailing through the Panama Canal. Photo courtesy of ODP.

BACKGROUND

One of the main objectives of the ODP was to gain insight into fundamental processes regulating Earth's ocean and climate systems. Although the instrumental record provides direct measurements of seasonal to interannual climate cycles, paleoclimatological proxy-climate data derived from ocean sediment samples can be used to map oceanic and climatic variability that occurs on time scales longer than a few decades. Paleoclimatological studies are relevant to future global warming predictions because they provide evidence for the response of global climate to high levels of greenhouse gases, the relationships between high-frequency variability and mean climate states, and the causes of abrupt, nonlinear climate responses to perturbations and persistent forcings. Paleoclimate records are also critical for the study of climate components that respond on relatively long time scales (e.g., ice sheets and deep ocean circulation). To this end, the ODP focussed much of it drilling efforts on recovering ocean sediments (from a range of latitudes and water depths in the Pacific, Atlantic, and Indian Oceans) that represent past extreme warm and cold periods, past abrupt changes and transitions, and past variability or cyclicity of climate under different global boundary conditions. In particular, a unique opportunity exists to study climate change over the last 5 million years using ODP cores. Earth's climate regime was dramatically different in the early Pliocene warm period (5 to 3 Ma) as compared to the Pleistocene ice age (~1.5 million years to present) (Figure 1), and it is relatively easy to obtain pristine material from this time period. As outlined below, the first goal of our study was to understand the role of tropical oceans in determining global warmth during the early Pliocene (~5 to 3 Ma). The second goal was to understand the role of the tropical oceans in the transition to cooler climatic conditions (from ~3 to ~1.5 Ma).

Ana Christina Ravelo *(acr@es.ucsc.edu) is Associate Professor, Ocean Sciences Department, University of California, Santa Cruz, CA.* **Michael William Wara**, *Ph.D., is currently a student at Stanford Law School, Stanford, CA.*

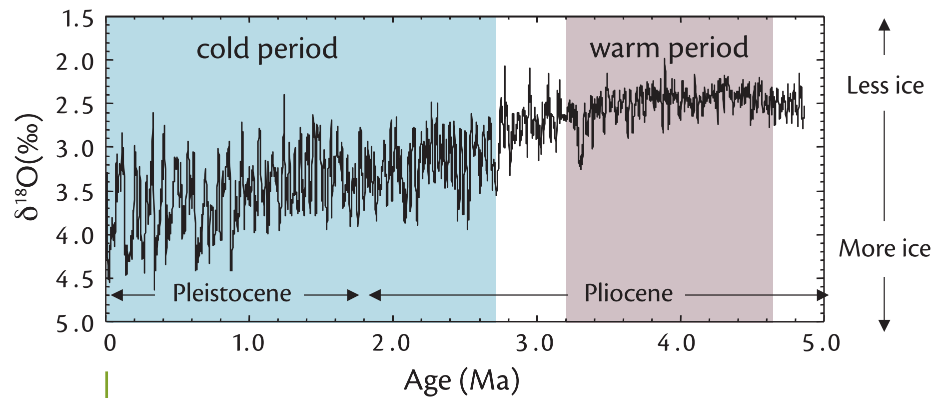

Figure 1. Benthic foraminifera $\delta^{18}O$ record of high-latitude climate (primarily ice volume) from ODP Site 677 (Shackleton et al., 1990) and ODP Site 1085 (Andreasen, 2001). The end of the Pliocene warm period and the onset of Northern Hemisphere glaciation occurred at ~3.0 Ma.

Warm early Pliocene. The Pliocene warm period was the most recent period of sustained global warmth relative to today with ~3°C higher global surface temperature (Sloan et al., 1996; Dowsett et al., 1996; Haywood et al., 2000). Northern Hemisphere ice sheets were small (Jansen and Sjøholm, 1991), and atmospheric carbon dioxide concentration (pCO_2) was probably slightly higher (Van der Burgh et al., 1993; Raymo et al., 1996) compared to today. Initial compilations of sea surface temperatures (SSTs) from the Pliocene warm period indicated that high-latitude oceans, but not tropical oceans, were significantly warmer than today (Dowsett et al., 1996). This observation led paleoclimatologists to speculate that the Pliocene warm period could be explained by enhanced meridional heat advection, which would cause high-latitude warming without affecting tropical SSTs significantly (Sloan et al., 1996; Chandler et al., 1994). The difficulty in understanding the role for enhanced advection versus greenhouse-gas forcing was summarized by Crowley (1991, 1996), who highlighted the need for additional low-latitude SST data in order to resolve the causes of Pliocene warmth. Contrary to this earlier work, a recent modeling study predicts that in the Pliocene warm period tropical and sub-tropical, SSTs were warmer than modern SSTs due to differences in the cryosphere and cloud feedbacks (Haywood and Valdes, 2004). This new and different modeling result is independently supported by observations based on alkenone-derived SSTs (Haywood et al., in review) and suggests that the Pliocene warm period could actually be related to higher-than-modern pCO_2 concentrations (Crowley, 1991; Crowley, 1996). More importantly, the model suggests that the causes of Pliocene warmth need to be understood in the context of warmer tropical and sub-tropical oceans. Thus, to advance understanding of the causes of early Pliocene global warmth, our first goal was to further characterize major differences in tropical climate conditions during the warm period compared to today.

The transition to colder climate. A comprehensive view of high-latitude climate changes demonstrates that the end of the Pliocene warm period at ~3.0 Ma and the onset of significant NHG were generally not "abrupt" (Raymo, 1994), although in some locations in the North Pacific and North Atlantic, the transition appears to have occurred specifically at 2.75 Ma (with errors of < ~50 ka) (Haug et al., 1999) once specific regional thresholds were reached. The cause or causes of NHG has been the subject of much debate. NHG may have been a result of long-term global cooling caused by a small decrease in carbon dioxide concentration coupled with positive feedbacks (e.g., Saltzman and Verbitsky, 1993; Berger and Wefer, 1996; Crowley, 1996; Raymo, 1998), as suggested for the onset of significant Antarctic glaciation around 34 Ma (DeConto, 2003). Alternatively, other studies propose that a high-latitude threshold, such as atmospheric cooling from repeated volcanic eruptions (Prueher and Rea, 2001), or diversion of Arctic air masses to mid-latitudes due to mountain building (Ruddiman and Raymo, 1988; Ruddiman and Kutzbach, 1990), was reached by ~3.0 Ma, which allowed for the onset of significant NHG and subsequent global cooling. As a third alternative, recent studies have proposed that low-latitude tectonic events, such as the uplift of the Isthmus of Panama (Driscoll and Haug, 1998) or the uplift and movement of islands within the Indonesian seaways (Cane and Molnar, 2001), may have forced tropical climate reorganization, which in turn may have caused high-latitude cooling and NHG. These hypotheses for the end of the warm period can be tested by examining the timing of the transition to cooler Pleistocene climate in different regions using ODP records. A recent paper (Ravelo et al., 2004) showed that major climate events around the globe occurred at significantly different times, implying that no abrupt regional event followed by a cascade of climate changes could have led to the Pleistocene ice ages. This result implies that the end of the warm period was caused by a gradual and persistent forcing, with different regions responding to the (unidentified) forcing at different times due to regionally specific thresholds. Thus, the second goal of our study was to investigate the potential role of the tropical oceans in gradually putting an end to the Pliocene warm period.

Reorganization of tropical SST patterns can strongly influence global climate, as occurs interannually with the El Niño-Southern Oscillation phenomenon (Cane and Evans, 2000; Philander and Fedorov, 2003). Even small changes in tropical SST patterns can profoundly affect extratropical climate on geological time scales (Yin and Battisti, 2001). Although it has been proposed that low-latitude tectonic events (uplift of the Isthmus of Panama or restriction of flow through the Indonesian seaway) may have changed the distribution of heat between basins, causing reorganization of global climate patterns, tropical ODP records have never been systematically synthesized and interpreted in this light. This article will present ODP records of sea surface changes from the tropical Indian, Pacific, and Atlantic Oceans with the intention of characterizing tropical conditions during the warm Pliocene, resolving the timing of tropical SST changes at the end of the warm period, and exploring the implications of this timing on proposed mechanisms for the Pliocene-Pleistocene climate transition.

MODERN VERSUS EARLY PLIOCENE SST DISTRIBUTION

The modern tropical SST distribution in the Indian, Pacific, and Atlantic Oceans is predominantly determined by the trade winds and the structure and depth of the subsurface thermocline. Although wind-driven equatorial upwelling occurs across the three basins, the thermocline is sufficiently deep in the Indian Ocean and in the western Pacific and Atlantic Oceans that SSTs remain relatively warm in those regions. Cool SSTs are found where upwelling-favorable winds and a shallow thermocline coincide, mainly in eastern Pacific and Atlantic tropical oceans (Figure 2). Thus, the tropical Pacific, whose temperature distribution is thought to influence climate patterns on a global scale, is characterized by strong west-east gradients in temperature and pressure. The easterly trade winds, strengthened by these gradients, further reinforce cool upwelling in the east, thereby augmenting the temperature and pressure gradients. The strong zonal, or Walker, circulation is maintained by these positive air -sea feedbacks (Bjerknes, 1969). The average conditions in the tropical Pacific today include strong zonal gradients and Walker circulation. However, the same air-sea feedbacks that maintain strong Walker circulation also amplify small perturbations that weaken Walker circulation, causing the thermocline to deepen in the

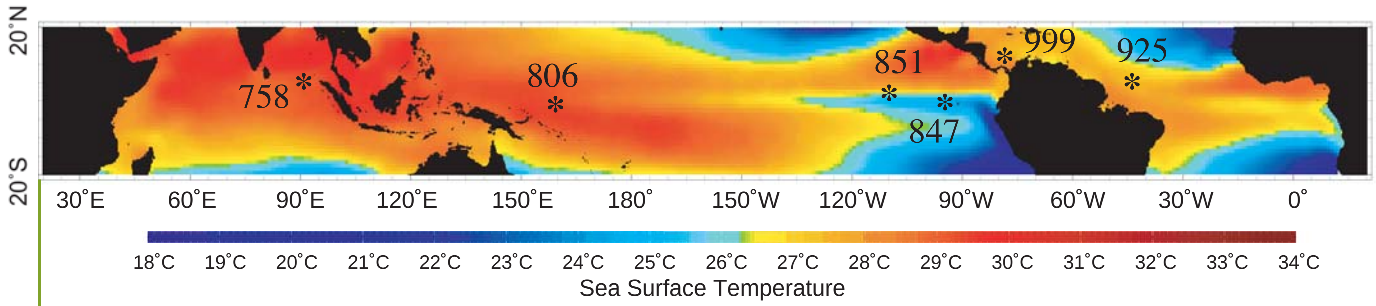

Figure 2. Sea surface temperature (SST) map of the world's tropical oceans (Reynolds and Smith, 1994). Change in the strong SST gradient across the equatorial Pacific is monitored using ODP Sites 806, 847, and 851. Changes in the strong eastern Pacific to Caribbean and Caribbean to western Atlantic salinity gradients (not shown) are monitored using ODP Sites 851 and 999, and ODP Sites 999 and 925, respectively. Changes in the weak SST gradient from the western Pacific to the eastern Indian Ocean are monitored using ODP Sites 806 and 758.

east and El Niño conditions to develop every few years.

There is strong evidence that tropical Pacific conditions in the early Pliocene resembled a permanent El Niño. Specifically, the west-east SST gradient was greatly reduced, and the thermocline in the eastern tropical Pacific was deep compared with modern normal conditions (Cannariato and Ravelo, 1997; Chaisson and Ravelo, 2000; Wara, 2003; see discussion of Figure 6 below). Furthermore, extratropical climate anomalies relative to today were similar to those manifested during a modern El Niño (Molnar and Cane, 2002). The reasons for an El Niño-like pattern in the early Pliocene are unclear, but could be related to differences in the source of thermocline waters compared to today. The thermocline in the tropical Pacific is influenced by conditions in the regions where it is ventilated at mid-latitudes (Gu and Philander, 1997; Harper, 2000) by the amount of exchange with the Indian Ocean through Indonesian passages (Rodgers et al., 2000; Cane and Molnar, 2001), and by whole-ocean stratification (Philander and Fedorov, 2003). Changes in any of these factors could have potentially caused the thermocline to be deeper and/or warmer in the warm Pliocene compared to today, thereby resulting in warmer SSTs in upwelling regions, reduced west-east SST gradient in the Pacific, and weaker Walker circulation.

ESTABLISHMENT OF MODERN SST GRADIENTS

How and when did the modern tropical SST patterns become established? How did the restriction of flow through the Indonesian and Panama seaways affect tropical ocean conditions at the end of the warm Pliocene, and specifically, at the onset of significant NHG? These questions can be answered using records from tropical ODP sites that monitor changes in hydrographic gradients (Figure 2, Table 1). Although relatively new paleo-proxy measurements, such as magnesium to calcium ratios (Mg/Ca) and alkenone unsaturation ratios ($U^{k'}37$) are being widely applied to reconstruct SSTs in Pleistocene and recent times, there are no published records spanning the last 5 million years using these techniques at tropical locations. There are, however, oxygen isotope ($\delta^{18}O$) records from a number of tropical locations (Figure 2). The records presented in this article were all generated by measuring $\delta^{18}O$ of the calcite shells of *Globigerinoides sacculifer*, a planktonic (surface-dwelling) foraminifer species that has been widely applied in studies of surface paleoceanography.

The $\delta^{18}O$ composition of a foraminiferal shell primarily reflects changes in the $\delta^{18}O$ of seawater, and the temperature of seawater, in which the shell calcified. Evaporation/precipitation processes are responsible for changes in both the whole-ocean "global" and the "local" $\delta^{18}O$ of seawater. When evapora-

Table 1. Site Locations

Site	Latitude	Longitude	Water Depth (m)
999 (Caribbean Sea)	13°N	79°W	2828
925 (Western Tropical Atlantic Ocean)	4°N	43°W	3042
806 (Western Tropical Pacific Ocean)	0	159°E	2520
851 (Eastern Tropical Pacific Ocean)	3°N	111°W	3761
847 (Eastern Tropical Pacific Ocean)	0	95°W	3334
758 (Eastern Tropical Indian Ocean)	5°N	90°E	2924

tion occurs, water vapor is enriched in the lighter isotope (^{16}O), and therefore precipitation (rainfall or snow) has low δ^{18}O compared to seawater. Ice sheets act as a reservoir of water (precipitation) with low δ^{18}O values, and thus, during cold periods when large ice sheets store abundant water, the δ^{18}O of the whole ocean is relatively high. Much of the long-term variability in a δ^{18}O record is due to changes in the amount of ice stored on land. Over the last 5 million years, the long-term trend to higher δ^{18}O values (Figure 1) is mostly due to the effect of increasing Northern Hemisphere ice-sheet size on the whole-ocean δ^{18}O composition. Evaporation and precipitation also affects the local δ^{18}O composition of surface seawater, just as it does the salinity of surface seawater. Because rainfall has low δ^{18}O values compared to seawater, regions with high rainfall relative to evaporation will have lower seawater δ^{18}O values than regions with lower rainfall relative to evaporation. As a result, the δ^{18}O and salinity of seawater are highly correlated, and past local variations in salinity can sometimes be derived using the δ^{18}O measured on planktonic foraminifera. In addition, there is a strong temperature-dependent oxygen isotopic fractionation during the precipitation of calcite with more ^{18}O being incorporated into calcite as temperature decreases; thus, local temperature changes also influence the δ^{18}O composition of foraminiferal calcite shells. Overall, changes in foraminiferal δ^{18}O at any one locality reflects three factors: whole-ocean "global" changes in the δ^{18}O of seawater due to changes in ice-sheet size, local changes in the δ^{18}O of seawater due to changes in the hydrologic cycle, and local changes in seawater temperature.

Because changes in the whole-ocean δ^{18}O are embedded in all the tropical δ^{18}O records of *G. sacculifer*, it is difficult to use single records to predict absolute SST in past times. Rather, we can use the differences among the tropical records to reconstruct temperature and/or salinity gradients between sites. In other words, the shared variability (trends and cycles) in δ^{18}O records from different locations are primarily due to global ice-volume changes, but variations in one record relative to another reflect changes in local surface-water properties. The records can be examined to resolve the timing of the evolution from sustained El-Niño-like to modern-like hydrographic gradients.

Restriction of the Panamanian Seaway. Today, hydrographic conditions on either side of the Isthmus of Panama are different, with the eastern Pacific having lower temperature and salinity than the western Atlantic (Caribbean Sea). If the uplift of the Isthmus of Panama, which restricted flow between the two basins and reorganized the surface circulation in the Atlantic, played a key role in the end of the Pliocene warm period (Driscoll and Haug, 1998), then the δ^{18}O records on either side of the Isthmus should provide evidence for this reorganization at ~3 Ma. However, Haug et al. (2001) demonstrate that there is a clear divergence between δ^{18}O records from the eastern Pacific and the Caribbean at ~4.2 Ma, and attribute the establishment of the modern salinity gradient to the effective restriction of the Panama seaway at that time (Figure 3).

Models show that the Panama seaway restriction could have had a significant influence on Atlantic circulation (Maier-Reimer et al., 1990; Mikolajewicz and Crowley, 1997), including enhanced meridional ocean heat advection to high latitudes and thermohaline circulation. However, the idea that the increase in

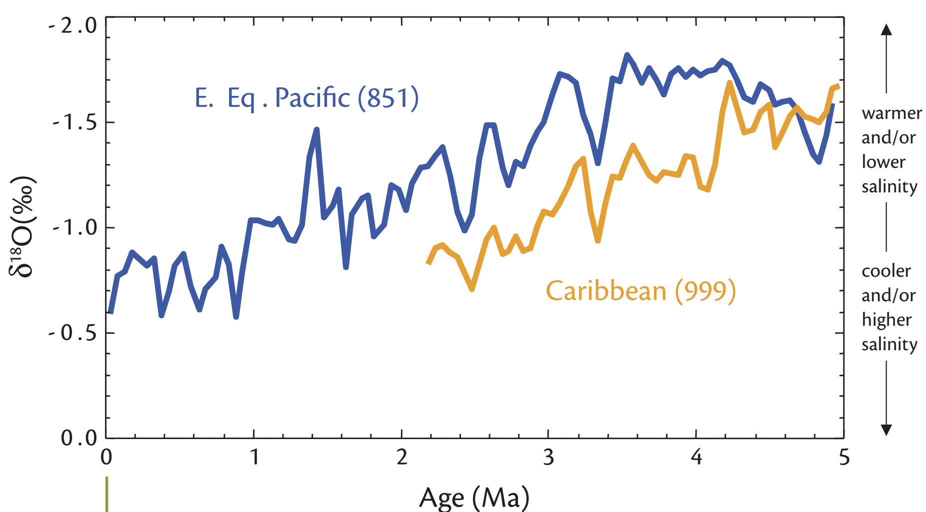

Figure 3. Planktonic (*Globigerinoides sacculifer* without sac-like final chamber) foraminifera δ^{18}O record from ODP Sites 851 in the eastern equatorial Pacific Ocean (Cannariato and Ravelo, 1997) (blue) and ODP Site 999 in the Caribbean Sea (Haug et al., 2001) (orange), from both sides of the Panama seaway. Only the smoothed (using weighted squared error method) record through the higher (~4 kyr) resolution data is shown. Note divergence of records between 4.5 and 4.0 Ma, indicating the establishment of a salinity gradient between the two sites.

warm water advected to the North Atlantic supplied the heat and moisture needed to build large ice sheets and establish permanent sea-ice coverage in the Arctic (Driscoll and Haug, 2001) needs to be further tested considering the nearly 1 million year lead time between the seaway closing and the onset of significant NHG after ~3 Ma. In fact, rather than being related to the end of the Pliocene warm period, the Panama Seaway closing event at 4.2 Ma may have been related to the beginning of the warm period (Berger and Wefer, 1996), which was superimposed on the long-term cooling that started tens of millions of years earlier.

While the divergence of the eastern Pacific and the Caribbean $\delta^{18}O$ records at 4.2 Ma has been interpreted as reflecting the restriction of flow through the Panama seaway, an alternative explanation is that the higher $\delta^{18}O$ values of the Caribbean after 4.2 Ma are due to a decrease in precipitation (increase in salinity) resulting from a migration of the Intertropical Convergence Zone (ITCZ) south relative to the Caribbean. This migration of the ITCZ, a zone of high precipitation, is supported by the decrease in $\delta^{18}O$ values (decrease in salinity) in the equatorial Atlantic (Figure 4) that occurred at the same time (Billups et al., 1998). Thus, the increase in the Caribbean $\delta^{18}O$ record between 4.5 and 4.0 Ma may be related to an increase in local salinity due to shifts in tropical atmospheric patterns, rather than the restriction of surface water exchange with the Pacific through the seaway. In either case, the records indicate that modern-like (steeper) hydrographic gradients between the eastern Pacific and Caribbean, and between the Caribbean and the tropical Atlantic, were established within the warm early Pliocene, and that the "event" was probably not directly responsible for the onset of NHG after ~3.0 Ma.

Restriction of the Indonesian Seaway. In the modern ocean, the tropical Indian and western tropical Pacific Oceans are part of the tropical "warm pool" and have similar surface-water conditions (Figure 2). Pliocene events, such as the uplift and drift of islands (Cane and Molnar, 2001, and references therein), may have changed the amount and character of the flow through the shallow straits and passageways connecting the Indian and Pacific Oceans. Thermocline water in the South Pacific is warm compared to that from the North Pacific, and as New Guinea drifted north across the equator, the flow of Pacific thermocline water into the Indian Ocean changed from having a dominantly southern warm source to a dominantly northern cold source (Rodgers et al., 2000). Cooler thermocline water in the Indian Ocean could have influenced SSTs in upwelling regions that in turn, through teleconnections, could have caused North America to cool allowing for the onset of NHG (Cane and Molnar, 2001).

Currently, there is no firm paleoceanographic evidence of thermocline temperature changes supporting this intriguing hypothesis. However, the records from either side of the Indonesian seaway indicate that the $\delta^{18}O$ gradient between the eastern Indian Ocean and the western Indian Ocean was different in the Pliocene (until ~1.5 Ma) than today. In the warm Pliocene, a $\delta^{18}O$ gradient was present (Figure 5), in contrast to the lack of a SST gradient that characterizes conditions today (Figure 2). If the $\delta^{18}O$ records are strictly interpreted as indicating change in the SST gradient, the data suggest that at the end of the warm Pliocene period, SST in the east Indian Ocean decreased, with near-modern SSTs established by ~1.5 Ma. Cooling of the Indian Ocean surface water relative to the west Pacific Ocean

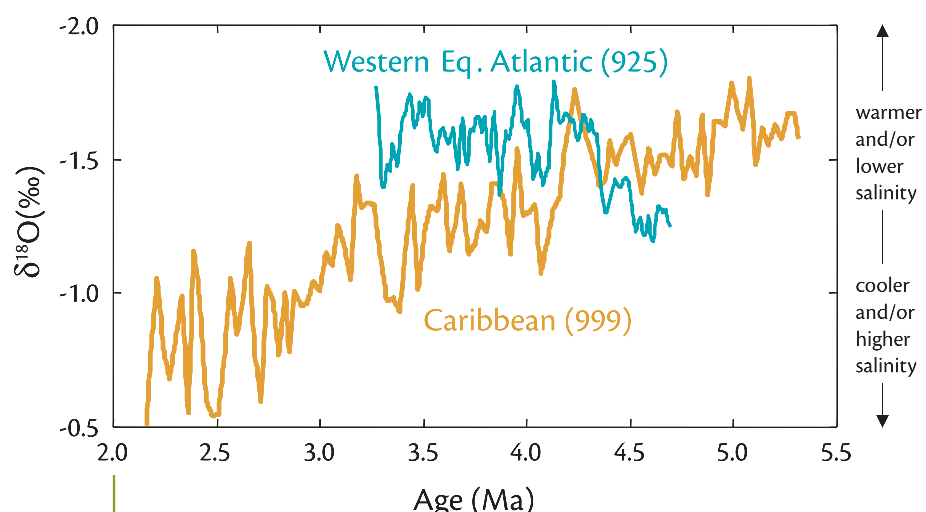

Figure 4. Planktonic (*G. sacculifer* without sac-like final chamber) foraminifera $\delta^{18}O$ record from ODP Sites 925 in the western tropical Atlantic Ocean (Billups et al., 1998) (aqua) and ODP Site 999 in the Caribbean Sea (Haug et al., 2001) (orange). Only the smoothed (using weighted squared error method) record through the higher (~4 kyr) resolution data is shown. Note crossover of records between 4.5 and 4.0 Ma, indicating a reversal of the salinity gradient between the two sites.

is consistent with the Cane and Molnar (2001) hypothesis. However, significant changes in the monsoon system and associated oceanographic conditions are known to have also occurred at around the same time (e.g., Clemens et al., 1996; Gupta and Thomas, 2003), and whether δ18O gradient changes are related to tectonically driven changes in the exchange of thermocline water between basins still needs to be tested. While it is hard to draw any firm conclusions in the absence of paleo-proxy data that more directly reflects sea surface and thermocline temperatures, certainly the gradual decrease in the δ18O gradient from ~3.0 Ma to ~1.5 Ma is intriguing. Whether it is related to a change in the Indonesian throughflow, which may have triggered global cooling and glaciation, still needs further investigation.

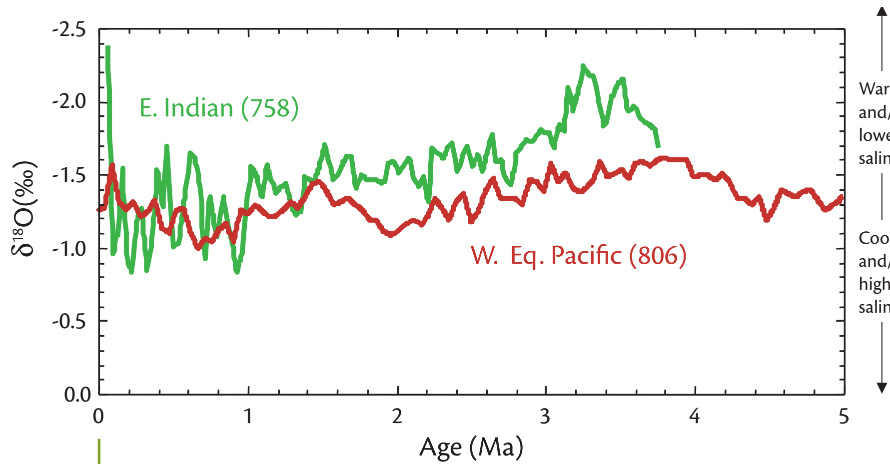

Figure 5. Planktonic (G. *sacculifer* without sac-like final chamber) foraminifera δ18O record from ODP Sites 758 in the eastern tropical Indian Ocean (Chen et al., 1995) (green) and ODP Site 806 in the western equatorial Pacific Ocean (Berger et al., 1993; Jansen et al., 1993) (red), from both sides of the Indonesian seaways. Only the smoothed (using weighted squared error method) record through the higher (~4 kyr) resolution data is shown. Note convergence of records by ~1.5 Ma, indicating the establishment of the modern (lack of) gradient between the two sites.

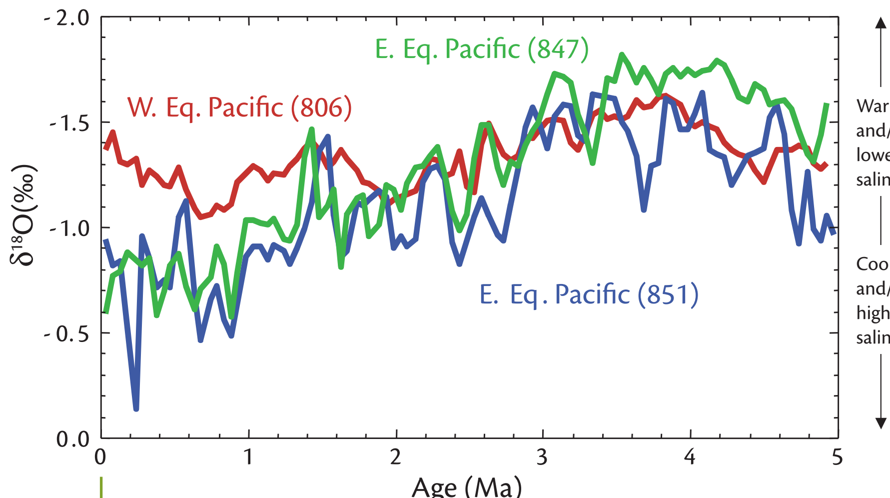

Figure 6.: Planktonic (G. *sacculifer* without sac-like final chamber) foraminifera δ18O record from ODP Sites 847 (Wara, 2003) (green) and 851 (Cannariato and Ravelo, 1997) in the eastern equatorial Pacific Ocean (blue), and ODP Site 806 in the western equatorial Pacific Ocean (Berger et al., 1993; Jansen et al., 1993) (red), from both sides of the equatorial Pacific Ocean. Only the smoothed (using weighted squared error method) record through the higher (~4 kyr for 851 and 806, and ~15 kyr for 847) resolution data is shown. Note divergence of records by ~1.5 Ma, indicating the establishment of the modern gradient and strong Walker circulation.

Establishment of Walker Circulation. The tropical Pacific west-east surface δ18O gradient, as recorded at ODP sites on the equator (Figures 2 and 6), serves as a paleo-proxy for the west-east temperature and pressure gradients and the strength of Walker Circulation. The gradient was small in the Pliocene warm period, indicating relatively weak Walker circulation. The Pacific cross-basinal gradient developed at ~1.5 Ma (Figure 6), which is the same time (within age errors) when the eastern Indian to western Pacific gradient disappeared (Figure 5), suggesting that the two could have been influenced by the same factors. One possibility is that both could be explained by the tectonic drift of New Guinea (discussed above) and the resulting reduction of warm thermocline water flow into the Indian Ocean, which could have caused the thermocline to cool in the Indian Ocean and to warm in the Pacific Ocean. However, how an increase in thermocline temperature in the western Pacific could have led to enhanced Walker Circulation is unclear and needs to be explored. Alternatively, it is possible that both observations (the decrease in SST in the Indian Ocean, and the increase in Walker circulation) were instead related to a gradual shoaling of the global thermocline. Philander and Fedorov (2003) argue that as global climate cooled through the Pliocene, whole-ocean stratification increased and the thermocline shoaled, causing SSTs in upwelling regions around the globe to cool. Cooling of SSTs in the east equatorial Pacific upwelling region could have caused the enhancement of Walker circulation. Although a gradually shoaling global thermocline, responding to global cooling, could explain many of our observations of changes in tropical surface ocean conditions, the ultimate cause of global cooling has not yet been identified.

CONCLUSIONS

Sustained El Niño-like conditions in the early Pliocene warm period. The average conditions in the eastern Indian, western Pacific, and eastern Pacific Oceans can be summarized using the average oxygen isotopic values within different time periods (Figure 7). The lack of a $\delta^{18}O$ gradient across the tropical Pacific Ocean (Figures 6 and 7) is strong support for the fact that on average, El Niño-like conditions (weak Walker circulation) were a persistent feature of the Pliocene warm period. However, it is unlikely that the modern dynamics responsible for the El Niño phenomenon are directly relevant to understanding early Pliocene climate because the average "background" state of the modern tropical Pacific is different than that of the warm Pliocene, and because there are large differences in the processes that operate on geological versus El Niño's interannual time scales. Regardless of the cause, weak tropical Walker circulation is potentially a permanent characteristic of a warm period, and should have influenced large-scale atmospheric circulation patterns and climate (Rind, 2000; Yin and Battisti, 2001). The factors that contributed to El Niño-like tropical conditions need to be studied if we are to understand fundamental controls of global climate change, especially in the context of future global warming. In particular, we should pursue further investigation of the factors that control the character of the ventilated thermocline and its effect on tropical SST patterns and Walker circulation.

Tropical changes at the end of the early Pliocene warm period. Changes in tropical surface conditions indicate that modern-like (steep) SST gradients were established in steps, first at ~4.2 Ma between the eastern Pacific and western Atlantic (Figure 3), and then at ~1.5 Ma when strong zonal, or Walker Circulation, developed (Figure 6). Modern-like conditions (0 to 1 Ma) (Figure 7) were not established until well after the onset of NHG at ~3.0 Ma. In other words, global climate change after the end of the Pliocene warm period was asynchronous, and the onset of NHG was not directly triggered by sudden tropical reorganization. Rather, a gradual source of global cooling must be identified.

Currently, there are two possibilities. The first is that the gradual forcing of global cooling was related to decreasing greenhouse gas composition due to changes in mineral weathering rates or other processes that influence the carbon cycle. Although the cause of a decrease is unknown, it could provide an explanation for global cooling, the onset of NHG, the theoretical increase in ocean stratification, and the shoaling of the ventilated thermocline with its accompanying effect on tropical SSTs, as documented in this study. The second possibility is that gradual tectonic forcings could have influenced ocean/atmosphere circulation, triggering global climate change. In fact, gradual changes in the $\delta^{18}O$ gradient between the western Pacific and the eastern Indian Ocean occurs during the transition from the warm Pliocene to cool Pleistocene (from about 3.0 to 1.5 Ma), and may be related to the gradual tectonic uplift and drift of

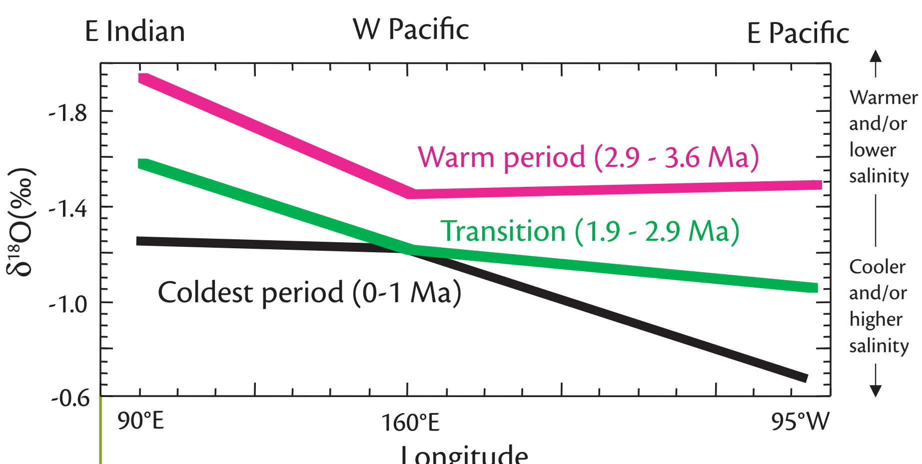

Figure 7. Average planktonic $\delta^{18}O$ values for three different time intervals at the eastern Indian (758), western Pacific (806), and eastern Pacific (847) sites. The gradients in the late Pleistocene (black lines) reflect "modern-like" gradients. In the Pliocene warm period (~2.9 to 3.6 Ma) (pink lines), the gradients indicate that the Pacific Walker circulation was greatly reduced. During the transition (~1.9 to 2.9 Ma) (green line) after the onset of Northern Hemisphere glaciation, the tropical gradients were not significantly different than they were in the warm period. This implies that Northern Hemisphere glaciation had little influence on tropical conditions. Note that the constant offset between the warm period (~2.9 to 3.6 Ma) and the transition period (~2.0 to 2.9 Ma) reflects higher global ice volume in the later period.

islands in the Indonesian seaway. Future studies should focus on reconstructing changes in the Pacific and Indian Ocean thermoclines, and understanding the effects of those changes on tropical and global climate.

ACKNOWLEDGEMENTS

This study was improved by suggestions made by H. Dowsett and an anonymous reviewer. All paleoceanographic data were generated using material obtained by the Ocean Drilling Program. This work was supported by the National Science Foundation. Any opinions, findings, and conclusions expressed here are those of the authors and do not necessarily reflect the views of the National Science Foundation.

REFERENCES

Andreasen, D.H. 2001. Isolating Sources of Plio-Pleistocene Paleoclimate Variability. Thesis, University of California, Santa Cruz.

Berger, W.H., T. Bickert, H. Schmidt, and G. Wefer. 1993. Quaternary oxygen isotope record of pelagic foraminifers; Site 806, Ontong Java Plateau. *Proceedings of the Ocean Drilling Program Scientific Research* 130:381-395.

Berger, W.H., and G. Wefer. 1996. Expeditions into the Past: Paleoceanographic studies in the South Atlantic. Pp. 363-410 in *The South Atlantic: Present and Past Circulation*, G. Wefer, W.H. Berger, G. Sidler, and D.J. Webb, eds. Springer-Verlag, Berlin.

Billups, K., A.C. Ravelo, and J.C. Zachos. 1998. Early Pliocene climate: A perspective from the western equatorial Atlantic warm pool. *Paleoceanography* 13:459-470.

Bjerknes, J. 1969. Atmospheric teleconnections from the tropical Pacific. *Monthly Weather Review* 97:163-172.

Cane, M.A., and M. Evans. 2000. Do the tropics rule? *Science* 290:1,107 -1,108.

Cane, M.A., and P. Molnar. 2001. Closing of the Indonesian Seaway as a precursor to East African aridification around 3-4 million years ago. *Nature* 411:157-162.

Cannariato, K., and A.C. Ravelo. 1997. Plio-Pleistocene evolution of eastern tropical Pacific surface water circulation and thermocline depth. *Paleoceanography* 12:805-820.

Chaisson, W., and A. Ravelo. 2000. Pliocene development of the East-West hydrographic gradient in the Equatorial Pacific. *Paleoceanography* 15:497-505.

Chandler, M., D. Rind, and R. Thompson. 1994. Joint investigations of the middle Pliocene climate: II. GISS GCM Northern Hemisphere results. *Global Planet Change* 9:197-219.

Chen, J., J.W. Farrell, D.W. Murray, and W.L. Prell. 1995. Timescale and paleoceanographic implications of a 3.6 Ma oxygen isotope record from the northeast Indian Ocean (Ocean Drilling Program Site 758). *Paleoceanography* 10:21-48.

Clemens, S., D.W. Murray, and W.L. Prell. 1996. Nonstationary phase of the Plio-Pleistocene Asian monsoon. *Science* 274:943-948.

Crowley, T.J. 1991. Modeling Pliocene Warmth. *Quaternary Science Reviews* 10:275-282.

Crowley, T.J. 1996. Pliocene climates: The nature of the problem. *Marine Micropaleontology* 27:3-12.

DeConto, R.M., and D. Pollard. 2003. Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. *Nature* 421:245-249.

Dowsett, H., J. Barron, and R.Z. Poore. 1996. Middle Pliocene sea surface temperatures: A global reconstruction. *Marine Micropaleontology* 27:13-25.

Driscoll, N.W., and G.H. Haug. 1998. A short circuit in thermohaline circulation; a cause for Northern Hemisphere glaciation? *Science* 282:436-438.

Gu, D., and G.S. Philander. 1997. Interdecadal climate fluctuations that depend on exchanges between the tropics and extratropics. *Science* 275:805-807.

Gupta, A.K., and E. Thomas. 2003. Initiation of Northern Hemisphere glaciation and strengthening of the northeast Indian monsoon: Ocean Drilling Program Site 758, eastern equatorial Indian Ocean. *Geology* 31: 47-50.

Harper, S. 2000. Thermocline ventilation and pathways of tropical-subtropical water mass exchange. *Tellus A* 52:330-345.

Haug, G.H., R. Tiedemann, R. Zahn, and A.C. Ravelo. 2001. Role of Panama uplift on oceanic freshwater balance. *Geology* 29(3):207-210.

Haug, G.H., D.M. Sigman, R. Tiedemann, T.F. Pedersen, and M. Sarnthein. 1999. Onset of permanent stratification in the subarctic Pacific Ocean. *Nature* 401:779-782.

Haywood, A.M., P.J. Valdes, and B.W. Sellwood. 2000. Global scale palaeoclimate reconstruction of the middle Pliocene climate using the UKMO GCM: Initial results. *Global Planet Change* 25:239-256.

Haywood, A.M., and P.J. Valdes. 2004. Modelling Middle Pliocene Warmth: Contribution of Atmosphere, Oceans and Cryosphere. *Earth and Planetary Science Letters* 218:363-377.

Haywood, A.M., P. Dekens, A.C. Ravelo, and M. Williams. In review. Warmer tropics during the mid-Pliocene: Evidence from alkenone paleothermometry and a fully coupled mid Pliocene ocean-atmosphere model.

Jansen, E., L.A. Mayer, J. Backman, R.M. Leckie, and T. Takayama. 1993. Evolution of Pliocene climate cyclicity at Hole 806 B (5-2 Ma); oxygen isotope record. *Proceedings of the Ocean Drilling Program Scientific Research* 130:349-362.

Jansen, E., and J. Sjøholm. 1991. Reconstruction of glaciation over the past 6 Myr from ice-borne deposits in the Norwegian Sea. *Nature* 349:600-603.

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

Mikolajewicz, U., and T.J. Crowley. 1997. Response of a coupled ocean/energy balance model to restricted flow through the central American Isthmus. *Paleoceanography* 12:429-441.

Molnar, P., and M.A. Cane. 2002. El Niño's tropical climate and teleconnections as a blueprint for pre-Ice Age climates. *Paleoceanography* 17:11-1–11-13.

Philander, S.G., and A.V. Fedorov. 2003. Role of tropics in changing the response to Milankovitch forcing some three million years ago. *Paleoceanography* 18 (2):1045.

Prueher, L.M., and D.K. Rea. 2001. Volcanic triggering of late Pliocene glaciation: Evidence from the flux of volcanic glass and ice-rafted debris to the North Pacific Ocean. *Palaeogeography, Palaeoclimatology, Palaeoecology* 173:215-230.

Ravelo, A.C., D. Andreasen, M. Lyle, A. Olivarez Lyle, and M.W. Wara. 2004. Regional climate shifts caused by gradual global cooling in the Pliocene epoch. *Nature* 429:263-267.

Raymo, M.E. 1994. The Initiation of Northern Hemisphere Glaciation. *Annual Review of Earth And Planetary Sciences* 22:353-383.

Raymo, M.E., B. Grant, M. Horowitz, and G.H. Rau. 1996. Mid-Pliocene warmth: Stronger greenhouse and stronger conveyor. *Marine Micropaleontology* 27:313-326.

Raymo, M.E. 1998. Glacial puzzles. *Science* 281:1,467-1,468.

Reynolds, R.W., and T.M. Smith. 1994. Improved global sea surface temperature analyses. *Journal of Climate* 7:929-948.

Rind, D. 2000. Relating paleoclimate data and past temperature gradients: Some suggestive rules. *Quaternary Science Review* 19:381-390.

Rodgers, K.B., M. Latif, and S. Lugetke. 2000. Sensitivity of equatorial Pacific and Indian Ocean watermasses to the position of the Indonesian throughflow. *Geophysical Research Letters* 27:2,941-2,945.

Ruddiman, W.F., and J.E. Kutzbach. 1990. Late Cenozoic plateau uplift and climate change. *Transactions of the Royal Society of Edinburgh: Earth Sciences* 81:301-314.

Ruddiman, W.F., and M.E. Raymo. 1988. Northern Hemisphere climate regimes during the past 3 Ma: Possible tectonic connections. *Philosophical Transactions of the Royal Society of London* B318:411-430.

Saltzman, B., and M.Y. Verbitsky. 1993. Multiple instabilities and modes of glacial rhythmicity in the Plio-Pleistocene: A general theory of late Cenozoic climatic change. *Climate Dynamics* 9:1-15.

Shackleton, N.J., A. Berger, and R. Peltier. 1990. An alternative astronomical calibration of the lower Pleistocene timescale based on ODP Site 677. *Transactions of the Royal Society of. Edinburgh: Earth Sciences* 81:251-261.

Sloan, L.C., T.J. Crowley, and D. Pollard. 1996. Modeling of middle Pliocene climate with the NCAR GENESIS general circulation model. *Marine Micropaleontology* 27:51-61.

Van der Burgh, J., H. Visscher, D.L. Dilcher, and W.M. Kürschner. 1993. Paleoatmospheric signatures in Neogene fossil leaves. *Science* 260:1,788-1,790.

Wara, M. 2003. On the Interaction Between Tropical and High Latitude Climate Over the Past 5 Million Years. Thesis, University of California, Santa Cruz.

Yin, J.H., and D.S. Battisti. 2001. The importance of tropical sea surface temperature patterns in simulations of last glacial maximum climate. *Journal of Climate* 14:565-581.