The Pliocene Paradox

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ABSTRACT

During the early Pliocene, 5 to 3 million years ago (Ma), globally averaged temperatures were significantly higher than they are today even though the external factors that determine climate were essentially the same. The appearance of northern continental glaciers, and of cold surface waters in oceanic upwelling zones in low latitudes signaled the termination of those warm conditions. This introduced feedbacks involving ice-albedo and tropical ocean-atmosphere interactions that amplified obliquity (but not precession) cycles in equatorial sea surface temperatures, and in global ice volume, with the former leading the latter by several thousand years. A future melting of northern glaciers and a deepening of the thermocline could restore the warm conditions of the early Pliocene.

Introduction

The early Pliocene was both similar to, and also very different from the world of today. The intensity of sunlight incident on the Earth, the global geography, and the atmospheric concentration of carbon dioxide (1), were close to what they are today, but surface temperatures in polar regions were so much higher that continental glaciers were absent from the northern hemisphere, and sea level was approximately 25m higher than today (2,3,4,5). To resolve this apparent paradox, that conditions today, and those during the early Pliocene, are two different climate states in response to practically the same external forcing, a wealth of information is available, from the Pliocene Research Interpretations and Synoptic Mapping Project (PRISM) for example (2). Although there are some inconsistencies in the information concerning tropical ocean temperatures (6), most of the discrepancies can be eliminated by considering Pliocene observations, not in isolation, but in the broader context of earlier and subsequent climate changes.

Over the past 65 Ma, since the beginning of the Cenozoic when temperatures in polar regions were in the neighborhood of 10^0C, the Earth experienced the erratic global cooling shown in fig. 1(a). This was a consequence primarily of the drifting of the continents with which is associated changes in ocean-basin geometry, mountain-building, volcanic eruptions and other phenomena that affect the two factors that mainly determine globally averaged surface temperatures: the albedo of the planet, and the atmospheric concentration of greenhouse gases. Superimposed on the global cooling were periodic climate cycles in response to Milankovitch forcing. This term refers to modest, perfectly periodic variations in the distribution of sunlight because of periodic variations in orbital parameters such as the tilt (obliquity) of the Earth's axis. Although this forcing has been relatively constant over the past several million years, the amplitude of the climatic response has changed because, at different times, the long-term global cooling introduced different climate feedbacks.

The Pliocene is of special interest because, during that epoch, around 3 Ma, the amplitude of the response of climate to Milankovitch forcing increased, culminating in dramatic oscillations, over the last ~1Ma, between prolonged glaciations, or ice ages, and brief, warm interglacials (fig. 1). During the warm interglacial periods - the current one started some ten thousand years ago - conditions approach those of the early Pliocene. Will the present warm conditions terminate soon, to be followed by the next ice age? Or will the onset of the next ice age be inhibited by the current rise in the atmospheric concentration of greenhouse gases induced by humans? (See fig. 1b.) Will that rise restore the warm conditions of the early Pliocene? Answers to these questions require identification of the processes that maintained warm conditions in the early Pliocene and then terminated them. Initial studies focused on high latitude processes associated with the appearance of northern continental glaciers at ~3Ma. More recently, attention has turned to the tropics after the discovery that El Niño was a perennial (rather than intermittent) phenomenon up to ~3Ma.

A Polar Perspective

The global cooling that started some 50 Ma in due course led to the appearance of large ice sheets, first on Antarctica around 35 Ma, and subsequently on northern continents at ~3 Ma (7). In sediments from a site in the northern Pacific Ocean ice rafted debris appear abruptly at ~2.7 Ma, indicating that the warm conditions of the early Pliocene had come to an end (8). Did this happen because of the increase in the Earth's albedo when the glaciers appeared?

This hypothesis has been tested by using general circulation models of the atmosphere. The results indicate that removal of the northern glaciers increases temperatures only in the regions where the glaciers had been (4). If, in addition, sea surface temperatures are specified to be high in high latitudes, but unchanged in low latitudes, then the atmosphere is warmer over much larger regions, with some effects extending into the tropics (9).

These results suggest that high sea surface temperatures in high latitudes, more than low albedo, helped maintain the warm conditions of the early Pliocene, but what processes maintained those high surface temperatures? One possibility is a larger poleward transport of heat by oceanic currents during the early Pliocene. Presumably the closure of the Panamanian Seaway in the early Pliocene reorganized the oceanic circulation and in due course altered the poleward transport of heat. However, an open Seaway could not explain the early Pliocene high latitude warmth because the closure at the end of the warm period would have intensified the deep, thermohaline circulation in the Atlantic sector (10,11), thus transporting more, not less heat northward. It is conceivable that what mattered most was the northward transport, not of heat, but of moisture that promoted the growth of the glaciers and hence a cooling trend (12). One problem with this argument is that the timing of the closure between 4.5 and 4.0 Ma (13) was much earlier than the onset of cold high latitude conditions. Furthermore, this still leaves the warm conditions of the early Pliocene unexplained.

Another puzzle concerns the amplification of the Milankovitch cycles that started around 3 Ma. At high latitudes, the appearance of northern glaciers at that time introduced ice-albedo feedbacks that could have caused the amplification. Those feedbacks depend on variations in the intensity of solar radiation at high latitudes in summer (7). The precession of the equinox, and changes in the angle of tilt of the Earth's axis (obliquity), make comparable contributions to high latitude summer solar radiation. Why then, in the global ice volume records, is the obliquity signal dominant over the precession signal?

Obliquity is dominant in variations of the equator-to-pole gradient in sunlight. This gradient becomes a key parameter if the focus is, not on the poleward transport of heat which involves a negative feedback, but the transport of moisture which can amplify the waxing and waning of ice sheets (14). However, this gradient, by bringing conditions in the tropics into play, introduces new difficulties. Recent findings indicate that, around 3 Ma, Milankovitch cycles started amplifying not only in ice-volume but also in equatorial sea surface temperatures. Once again obliquity is the dominant signal, except that the signal in sea surface temperatures leads the one in ice volume by several thousand years (15, 16). How the fluctuations in ice-albedo could have caused the variations in the tropics is a puzzle, even if variations in ice-albedo, which matter most to climate, should lead variations in global ice volume. What are the tropical processes that could have contributed to climate variations associated with ice ages?

Tropical Perspective

At first, the paucity of the available data suggested that tropical and subtropical sea surface temperatures during the early Pliocene were essentially the same as those of today (2). Recent data from regions not covered by PRISM indicate otherwise. Apparently the salient features of sea surface temperature patterns in low latitudes, the cold surface waters off the western coasts of Africa and the Americas in fig. 2, were absent until approximately 3 Ma. This is evident in fig. 3

which shows that, up to about 3 Ma, the sea surface temperature difference between the eastern and western equatorial Pacific was very small and cold surface waters were absent from the coastal upwelling zones off the western coasts of Africa and the Americas (17,18,19,20,21). Today, a large reduction in the east-west temperature gradient along the equator in the Pacific occurs only briefly during El Niño which in effect was perennial rather than intermittent up to 3 Ma. Corroborating evidence for a permanent El Niño is available in land-records (22) that document the distinctive regional climate signatures associated with El Niño. Up to 3 Ma there was a persistence of mild winters in central Canada and the northeastern United States, droughts in Indonesia, and torrential rains along the coasts of California and Peru, and in eastern equatorial Africa. The onset of dry conditions in the latter region around 3 Ma favored the evolution of African hominids (23).

Persistent El Niño conditions would have had a huge impact on the global climate given that, today, even brief El Niño episodes can have a large influence. The reasons are evident in fig. 2 which shows a remarkably high correlation between tropical sea surface temperature and rainfall patterns. Tall, rain-bearing, convective clouds cover the warmest waters but highly reflective stratus decks that produce little rain cover the cold waters. During El Niño, the warming of the eastern equatorial Pacific reduces the area covered by stratus clouds thus decreasing the albedo of the planet, while the atmospheric concentration of the powerful greenhouse gas, water vapor, increases. Calculations with a General Circulation Model of the atmosphere indicate that this happened during the early Pliocene and contributed significantly to the warm conditions at that time (24).

The appearance of cold surface waters, around 3 Ma, in regions remote from each other in fig.3, can be explained in terms of a global shoaling of the oceanic thermocline. (A contributing factor could have been the northward drift of Australia that restricted flow from the equatorial western Pacific into the Indian Ocean (25).) A shallower thermocline must involve changes in the oceanic heat budget shown in fig 3. In a state of equilibrium the loss of heat in high latitudes, mainly where cold continental air masses flow over the warm Gulf Stream and Kuroshio Current in winter, balances the gain, mostly in low latitude upwelling regions where cold water rises to the surface. In a state of equilibrium an increase in the loss of heat must be accompanied by an increase in the gain of heat, which requires a shoaling equatorial thermocline (26) .

The oceanic heat transport is effected by the meridional overturning of the oceanic circulation. A freshening of the surface waters in the extra-tropics, which increases the buoyancy of the upper ocean and inhibits overturning, can therefore reduce the heat transport. This is true for both the deep, slow thermohaline component of the circulation whose changes affect mainly the climate of the northern Atlantic (27), and also for the rapid, shallow wind-driven component whose changes affect mostly the tropics. Sufficiently large freshening in the extra-tropics can induce a perennial El Niño (28).

These arguments suggest that, during, and probably prior to the early Pliocene, when El Niño was perennial, oceanic heat gain at low latitudes and heat loss at high latitudes were minimal, and the thermocline was deep. The theory explains why the gradual global cooling during much of the Cenozoic (fig. 1) and the associated decrease in the temperature of the deep ocean (29), caused the thermocline to shoal (17). A threshold was reached around 3 Ma when the thermocline was so shallow that the winds could bring cold waters to the surface in the various upwelling zones (30).

The appearance of cold surface waters in the equatorial upwelling zones introduced feedbacks that affected the response of tropical sea surface temperatures to Milankovitch forcing. The winds influence sea surface temperatures and also depend on those temperatures because they blow from the cold towards the warm regions along the equator where the Coriolis force vanishes. This circular argument implies positive feedbacks between the ocean and atmosphere (31). In the case of El Niño the feedbacks depend on an adiabatic redistribution of warm surface waters as in fig. 4(a). On much longer time-scales they involve diabatic, vertical movements of the thermocline as in fig. 4(b) and influence the climatic response to obliquity variations (30). During times when obliquity and solar heating in high latitudes have maxima, reduced oceanic heat loss to the atmosphere can induce a deepening of the equatorial thermocline and a tendency towards El Niño conditions. This is possible only on time-scales sufficiently long for the tropical oceans to adjust to changes in higher latitudes. Obliquity causes annually averaged sunlight to vary for such prolonged periods (of thousands of years) that the tropical response is practically in phase with the extra-tropical forcing (16). The precession of the equinox, by contrast, has too short a time-scale for an adjustment in the oceanic heat budget to be possible. It induces seasonal changes in sunlight but has no effect on the annually averaged sunlight. These arguments explain why low latitude temperature records are dominated by obliquity variations and are nearly in phase with high latitude rather than local solar variations (30) and why they become amplified over the long term when the thermocline shoals (16) and air-sea feedbacks are strengthened.

Discussion

A major factor in the warmth of the early Pliocene was the persistence of El Niño in the Pacific; it contributed to global warming by causing the absence of stratus clouds from the eastern equatorial Pacific, thus lowering the planetary albedo, and by increasing the atmospheric concentration of water vapor, a powerful greenhouse gas. Today the atmospheric concentration of another greenhouse gas, carbon dioxide, is comparable to what it was in the early Pliocene, but the climate of the planet is not yet in equilibrium with those high values. It is possible that a persistence of high carbon dioxide concentrations could result in a return to a globally warm world if it were to melt glaciers and increase temperatures in high latitudes, and as a consequence cause the tropical thermocline to deepen by a modest amount, a few tens of meters. (Near the date line at the equator the thermocline is already so deep that its vertical excursions leave surface temperatures unaffected.)

A deepening of the tropical thermocline requires a reduction in the oceanic heat loss in the extra-tropics (26). However, in certain atmospheric models, warm conditions in high latitudes depend on the atmosphere gaining heat from the oceans (32). This is also the case in the coupled ocean-atmosphere climate model that recently was used to simulate the early Pliocene (33). In that model, the oceanic heat loss in the extra-tropics is balanced by the gain of heat in the eastern equatorial Pacific. This gain is possible in spite of higher sea surface temperatures in low latitudes because temperature gradients along the equator, and presumably the depth of the equatorial thermocline, do not change significantly. This means that, in the model, maximum sea surface temperatures in the western tropical Pacific rise significantly above 30^0C. This is inconsistent with observations which indicate that, at no time in the past, were sea surface temperatures much higher than 30^0C. Are the models at fault, or is there a problem with the observations?

More data from the western tropical Pacific (and also from currently warm regions to the west of upwelling zones) are needed to determine the maximum temperatures over the last millions of years, and to determine whether observations of perennial El Niño are robust. If the information available at present should prove accurate, then temperatures in excess of 30^0C in some models, and problems in their ability to simulate a perennial El Niño, could be indicative of flaws in the models, in the parameterization of clouds for example. The models are designed to reproduce the world of today, but it is unclear how much confidence we should have in the simulations of very different climates.

Efforts to predict future global warming could benefit enormously from a better understanding of past climates, especially of the Milankovitch cycles for which the forcing functions are known precisely. The response to that forcing involves not only glaciers that wax and wane, but also equatorial sea surface temperatures that rise and fall. These phenomena could, to some degree be separate, each involving its own physical processes and feedbacks, in the same way that the response to seasonal variations in sunlight involves several different phenomena such as the Indian monsoons, coastal upwelling off southwest Africa, and severe winter storms in central Canada. An explanation for the monsoons does not explain the other features of the seasonal cycle. It is similarly unlikely that processes involving strictly ice-sheets will explain all aspects of the response to Milankovitch forcing. The important processes are likely to include, in addition to those discussed above, biogeochemical cycles such as the carbon cycle.

The obliquity cycles are of special interest because they started increasing in amplitude around 3 Ma, and then changed character again around 1 Ma. Initially it was thought that the latter change involved a shift of the dominant period from 40K to 100K (7). However, recent analyses indicate that the shift could have been from 40K to multiples of 40K, specifically 80K and 120K (34). Apparently obliquity continued to play a dominant role up to the present, in pacing glacial terminations, and also in equatorial Pacific variables (15, 35). To determine the relative importance of feedbacks involving ice-albedo, tropical ocean-atmosphere interactions, and biogeochemical cycles, it will be valuable to have a detailed description of the obliquity signal over the past several million years, inferred from a synthesis of the diverse measurements (productivity, sea surface temperature, thermocline depth, ice volume, atmospheric gas concentration etc.). Explaining the remarkable changes in this response to forcing that itself did not change significantly, is a major challenge for geoscientists.

References

1. Estimates of the concentration of carbon dioxide in the atmosphere during the early Pliocene range from 340-380 ppm (5,36) to 280-300ppm (37). Today the concentration has been in excess of 330ppm only a few decades, and have been elevated above 280ppm for only ten thousand years.

2. H. J. Dowsett, J. Barron, R. Poore, Mar. Micropaleontol. 27, 13 (1996).

3. H. J. Dowsett, M. A. Chandler, T. M. Cronin, G. S. Dwyer, *Paleoceanography* **20**, doi:10.1029/2005PA001133 (2005).

4. T. Crowley, *Mar. Micropaleontol*. **27**, 3 (1996).

5. M. E. Raymo, B. Grant, M. Horowitz, G. H. Rau, *Mar. Micropaleontol.* **27**, 313 (1996).

6. Geochemical and faunal-based estimates of sea surface temperature in the Early Pliocene, in the equatorial Pacific, are in basic agreement with each other as is evident in fig. 3 (17,19,20,21), and with the $\delta^{18}O$ record from planktonic foraminifers (22,30). The recent contradictory finding that the eastern tropical Pacific was relatively cool in the early Pliocene (38) is based on two, questionable data points, one of which is shown in fig.3 (the large blue circle at ~4.1Ma) and the other one lying outside the time range of this figure.

7. Zachos J., M. Pagani, L. Sloan, E. Thomas, K. Billups, *Science* **292**, 686 (2001).

8. G. H. Haug, D. M. Sigman, R Tiedemann, T. F. Pedersens, M Sarnthein, *Nature* **401**, 779 (1999).

9. L. C. Sloan, T. J. Crowley, D. Pollard, *Mar. Micropaleontol.* **27**, 51 (1996).

10. G. H. Haug, R Tiedemann, *Nature* **393**, 673 (1998).

11. A. Klocker, M. Prange, M. Schulz, *Geophys. Res. Let.* **32**, doi:10,1029/ 2004GL021564 (2005).

12. N.W. Driscoll, G.H. Haug, *Science* **282,** 436 (1998).

13. G.H. Haug, R. Tiedemann, R. Zahn, A.C. Ravelo, *Geology* **29,** 207 (2001).

14. M. E. Raymo, K. Nisancioglu, *Paleoceanography* **18**, doi:10.1029/2002PA000791 (2003).

15. Z. Liu, T. Herbert, *Nature* **427**, 720 (2004).

16. K. Lawrence , Z. Liu, T. Herbert, (in review).

17. M. Wara, A. C. Ravelo, M. L. Delaney, *Science* **309**, 758 (2005).

18. J. R. Marlow, C. B. Lange, G. Wefer, A. Rosell-Mele, *Science* **290**, 2288 (2000).

19. A. M. Haywood, P. Dekens, A. C. Ravelo, M. Williams, *Geochemistry, Geophysics, Geosystems* **6**, doi:10.1029/ 2004GC000799 (2005).

20. P. S. Dekens, A.C. Ravelo, M. McCarthy, (in review).

21. A. C. Ravelo, D. H. Andreason, M. Lyle, A. O. Lyle, M. Wara, *Nature* **429**, 263 (2004).

22. P. Molnar, M. Cane, *Paleoceanography* **17**, doi:10.1029/2001PA000663 (2002).

23. P. B. deMenocal, *Science* **270**, 53 (1995).

24. M. Barreiro, S. G. Philander, R. C. Pacanowski, A. V. Fedorov, Clim. Dyn. published online, doi:10.1007/s00382-005-0086-4 (2005).
25. M. Cane, P. Molnar, *Nature* **411**, 157 (2001).
26. G. Boccaletti, R. Pacanowski, S. G. Philander, A. V. Fedorov, *J. Phys. Oceanogr.* **34**, 888 (2004).
27. S. Manabe, R. J. Stouffer, *Nature* **378**, 165 (1995).
28. A. V. Fedorov, R. C. Pacanowski, S. G. Philander, G. Boccaletti, *J. Phys. Oceanogr.* **34**, 1949 (2004).
29. C. H. Lear, H. Elderfield, P. A. Wilson, *Science* **287**, 269 (2000).
30. S. G. Philander, A. V. Fedorov, *Paleoceanography*, **18**, 1045 (2003).
31. H. A. Dijkstra, J. D. Neelin, *J. Climate* **12**, 1630 (1999).
32. M. Winton, *J. Climate* **16**, 2875 (2003).
33. A. M. Haywood, P. J. Valdes, *Earth Planet. Sci. Lett.* **218**, 363 (2004).
34. P. Huybers, C. Wunsch, *Nature* **434**, 491 (2005)
35. L. Beaufort, T. de Garidel-Thoron, A. C. Mix, N. G. Pisias, *Science* **293**, 2440 (2001)
36. J. van der Burg, H. Visscher, D. L. Dilcher, V. M. Krshner, *Science* **260**, 1788 (1993).
37. M. Pagani, K. H. Freeman, M. A. Arthur, *Science* **285**, 876 (1999).
38. R. E. M. Rickaby, P. Halloran, *Science* **307**, 1948 (2005).
39. The location and depth of the ODP sites used for fig.3:
ODP site 806 (0°N, 159°E, 2520 m); ODP site 847 (0°N, 95° W, 3373m); ODP site 1014 (33˚N, 120˚W, 1165m) ; ODP site 1237 (16˚S, 76˚W, 3212m) ; ODP site 1084 (26˚S, 13˚E).
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Legends for figures

Figure 1.

Top: Variations in $\delta^{18}O$ over the past 60 million years. Note that the time-scale changes at 3 million years. The Milankovitch cycles are modest in amplitude up to ~3 million years but then start amplifying (7).
Bottom: Fluctuations in temperature and in the atmospheric concentration of carbon dioxide over the past 400,000 years as inferred from Antarctic ice-core records. The vertical red bar is the increase in atmospheric carbon dioxide levels over the past two centuries and before the year 2006.

Figure 2.

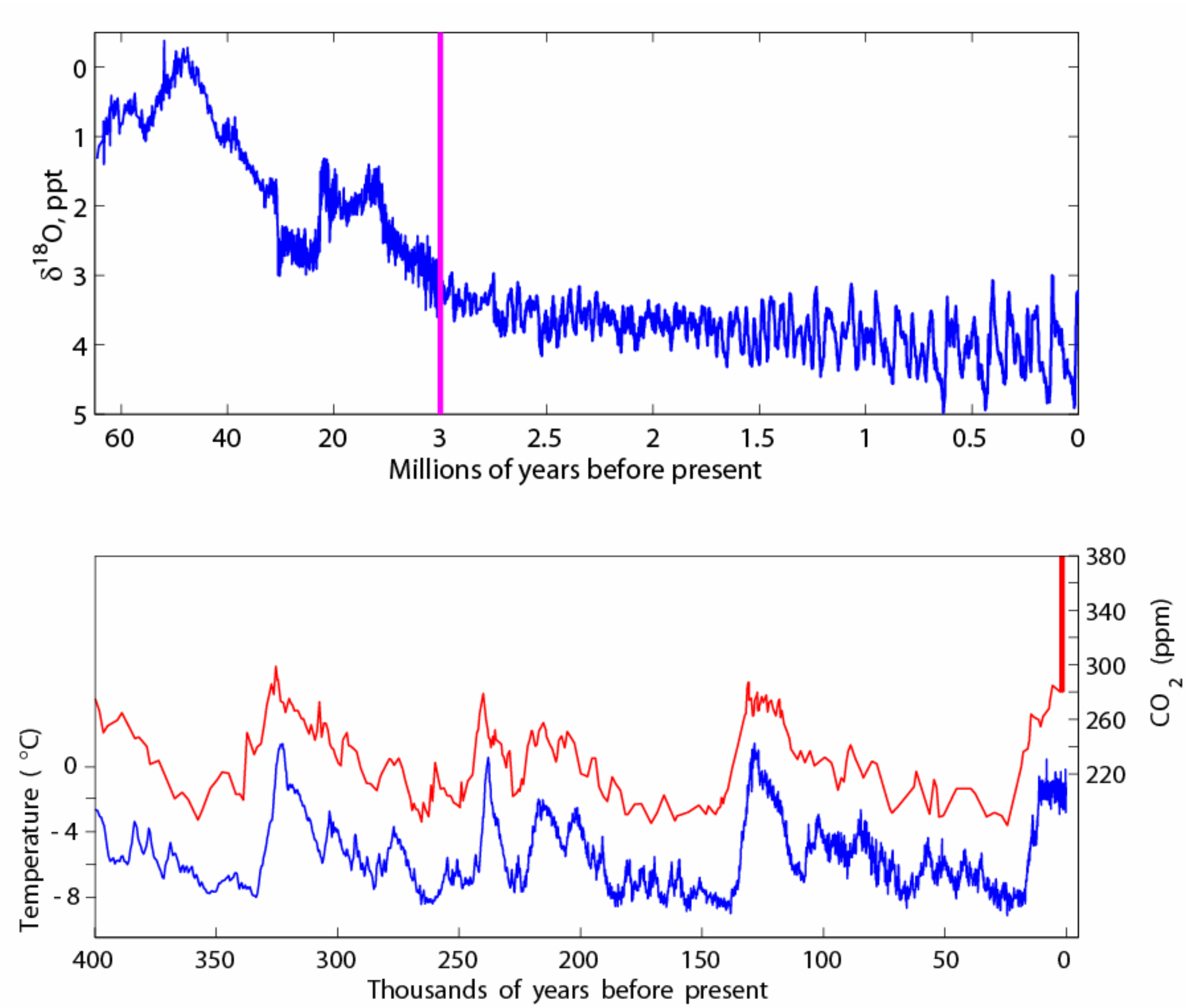
(a) Annual mean sea surface temperature pattern in ^{0}C. The dots show approximate locations of the deep cores used to calculate the time-series of SST displayed in fig. 3.
(b) Annual mean rainfall pattern in mm/day.
(c) The annual mean heat flux into the ocean in low latitudes, and out of the ocean in higher latitudes (in W/m^2), after daSilva *et al* (1994). Note a slightly different latitudinal extent of (c) as compared to (a) and (b)

Figure 3.

Top: Sea surface temperature (^{0}C) records in the western equatorial Pacific (red line, ODP site 806) and in the eastern equatorial Pacific (blue line, site 847) both based on Mg/Ca and adapted from (17), and that for the eastern Pacific based on alkenones (green dots) and adapted from (20). Larger circles are for the data based on Mg/Ca but from ref. (38): for ODP sites 806 (red) and 847 (blue). Pink shading denotes the early Pliocene. For the discussion see (6).
Bottom: $U^{k'}_{37}$ (alkenone-based) sea surface temperature records for the California margin (black, ODP site 1014), the Peru margin (blue, site 1237), and the West African margin (green, site 1084). From ref. 21, 19 and 18, respectively. The locations of the ODP sites are shown in fig.2; for the exact geographical locations see (39).

Figure 4.

A sketch that shows the difference between (a) an adiabatic, horizontal redistribution of heat that characterizes the interannual oscillations between El Niño and La Niña, and (b) a slow diabatic increase in the oceanic heat storage associated with an increase in obliquity for example. Thermocline movements in the equatorial plane are shown schematically. Perennial El Niño conditions would correspond to an extreme of (b). Ocean-atmosphere interactions cause changes in the thermocline on longer timescales to be a combination of (a) and (b).



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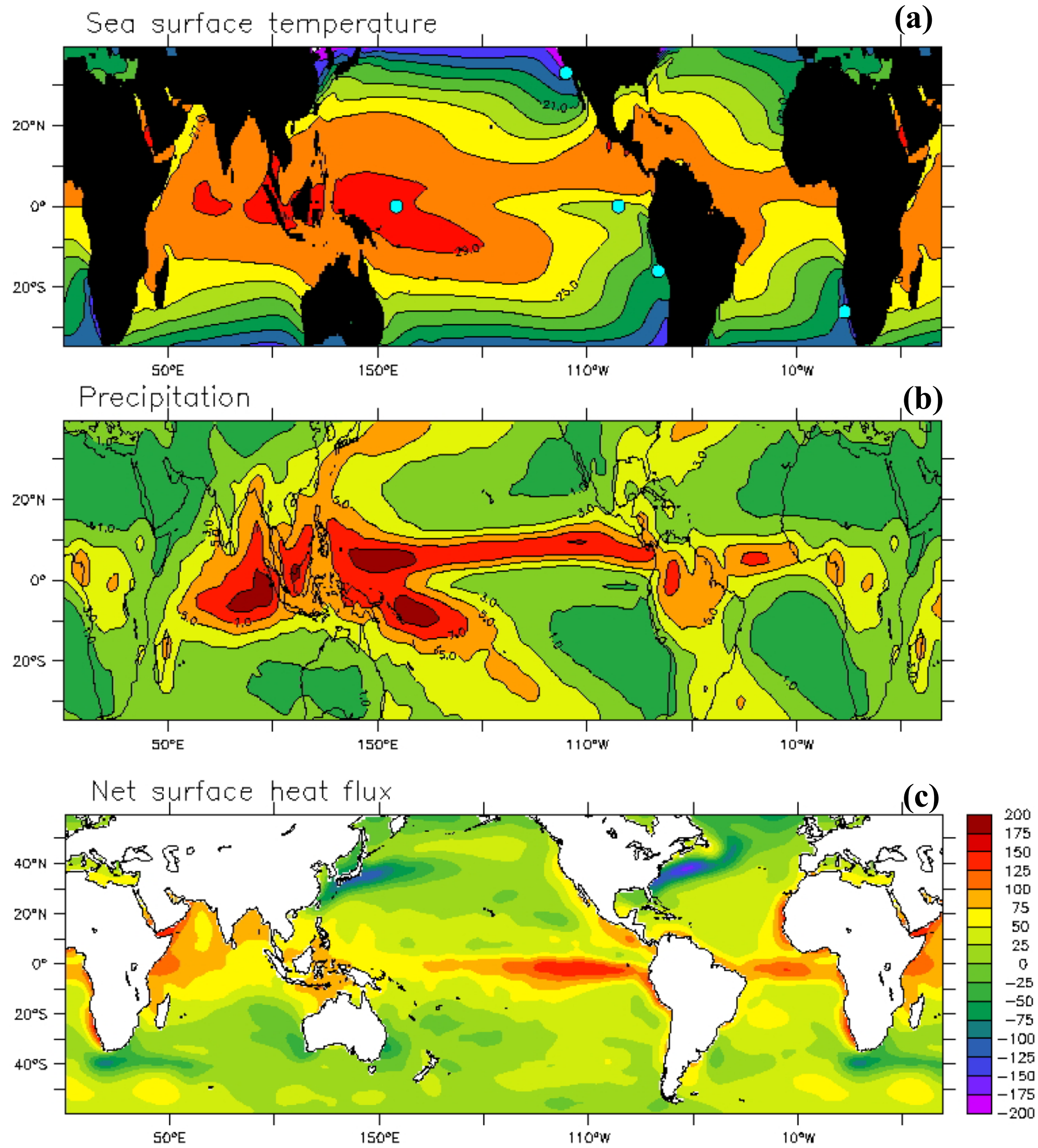

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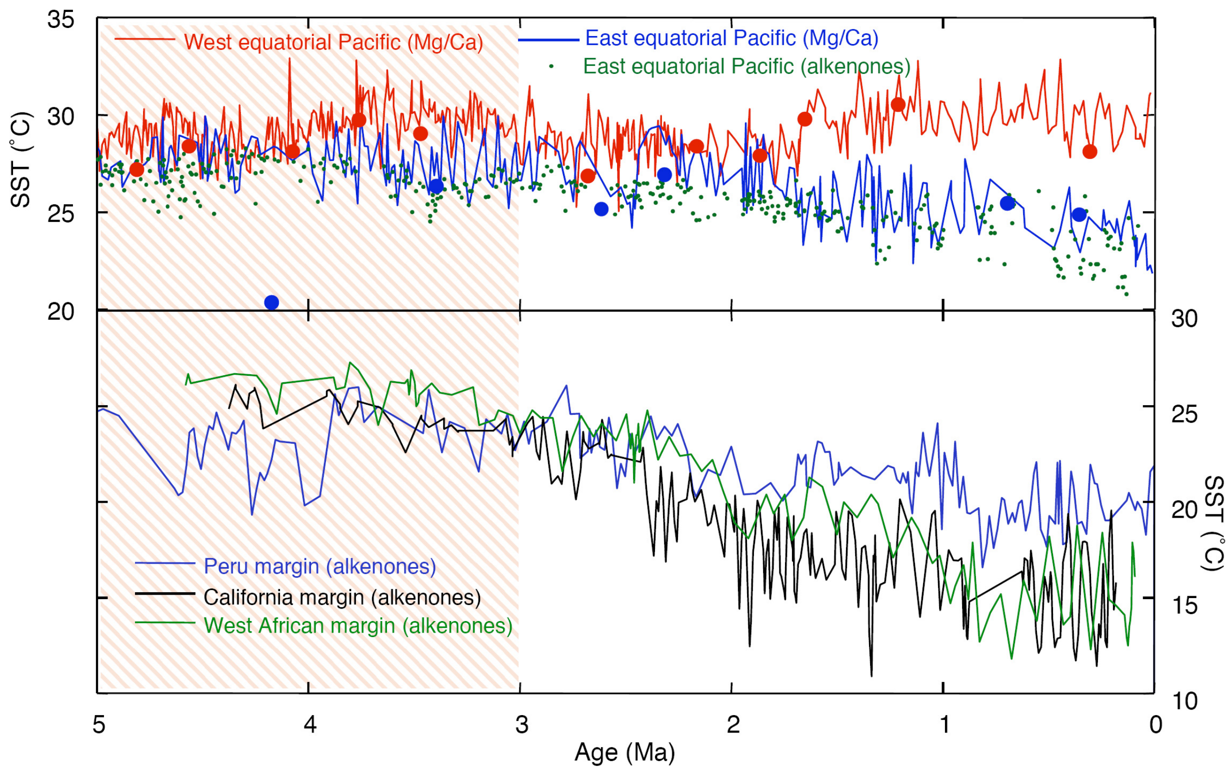

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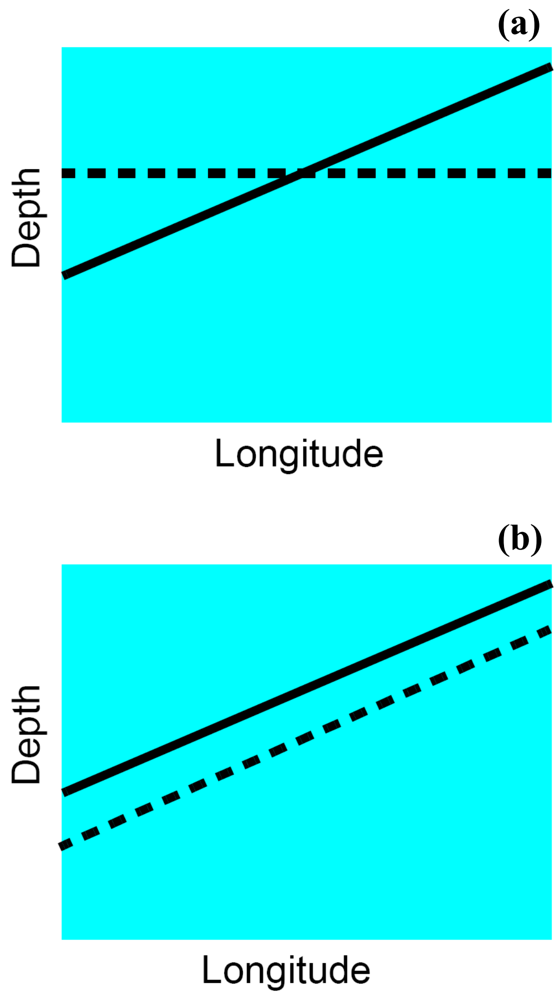

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