Role of tropics in changing the response to Milankovich forcing some three million years ago

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[1] Throughout the Cenozoic the Earth experienced global cooling that led to the appearance of continental glaciers in high northern latitudes around 3 Ma ago. At approximately the same time, cold surface waters first appeared in regions that today have intense oceanic upwelling: the eastern equatorial Pacific and the coastal zones of southwestern Africa and California. There was furthermore a significant change in the Earth's response to Milankovich forcing: obliquity signals became large, but those associated with precession and eccentricity remained the same. The latter change in the Earth's response can be explained by hypothesizing that the global cooling during the Cenozoic affected the thermal structure of the ocean; it caused a gradual shoaling of the thermocline. Around 3 Ma the thermocline was sufficiently shallow for the winds to bring cold water from below the thermocline to the surface in certain upwelling regions. This brought into play feedbacks involving ocean-atmosphere interactions of the type associated with El Niño and also mechanisms by which high-latitude surface conditions can influence the depth of the tropical thermocline. Those feedbacks and mechanisms can account for the amplification of the Earth's response to periodic variations in obliquity (at a period of 41K) without altering the response to Milankovich forcing at periods of 100,000 and 23,000 years. This hypothesis is testable. If correct, then in the tropics and subtropics the response to obliquity variations is in phase with, and corresponds to, El Niño conditions when tilt is large and La Niña conditions when tilt is small. *INDEX TERMS:* 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 4231 Oceanography: General: Equatorial oceanography; 4267 Oceanography: General: Paleoceanography; *KEYWORDS:* ocean-atmosphere interactions, El Niño, glacial cycles, Milankovich cycles

1. Introduction

[2] The Milankovich forcing amounts to periodic variations in the distribution and intensity of sunlight associated with periodic variations in parameters that include the tilt of its axis, at a period of 41,000 years, the precession of the axis, at a period of approximately 23,000 years, and the eccentricity of the orbit at a period of 100,000 years. Some three million years ago (3 Ma) the response of the Earth's climate to this forcing amplified sharply at a period of 41K, but not at the other periods, even though there was no associated change in the forcing. This is evident in records of global ice volume [*Raymo*, 1998; *Clark et al.*, 1999], and the intensity of the west African and Indian monsoons [*deMenocal*, 1995]. Apparently a climate feedback came into play at that time and amplified the response to the variations in obliquity, but not precession or tilt. What feedback favored the 41K cycle while discriminating against the 23K and 100K cycles? To answer this question it is helpful to take a long-term perspective, to view the changes around 3 Ma in the context of developments over the course of the Cenozoic.

[3] Some sixty million years ago (60 Ma) the Earth was so warm that the poles were free of ice. Subsequently our planet experienced gradual global cooling as is evident in of Figure 1 (bottom). The climate changes that accompanied this cooling (see *Zachos et al.* [2001] for a review) were in association with the perpetual drifting of the continents. Superimposed on these climate changes driven by sources of energy internal to the Earth, are the periodic fluctuations associated with external energy sources, the Milankovich cycles mentioned above. Figure 1 (top) shows how markedly the response to the Milankovich forcing has changed over the past five million years. The obliquity cycle became the most prominent one around 3 Ma. That continued to be the case until about one million years ago at which stage the 100,000 year signal became dominant. This paper concerns only the changes around 3 Ma.

[4] The gradual global cooling that started 60 Ma was bound to lead to the appearance of glaciers once temperatures were sufficiently low. Apparently that happened first in Antarctica, then in Greenland, and later, around 3 Ma, on continents in high northern latitudes [*Jansen and Sjøholm*, 1991; *Raymo*, 1994]. This last event could have caused the response to Milankovich forcing to amplify because of an ice albedo feedback: a weakening of sunlight in summer in high latitudes can cause the failure of snow of the previous winter to melt in summer, resulting in an accumulation of
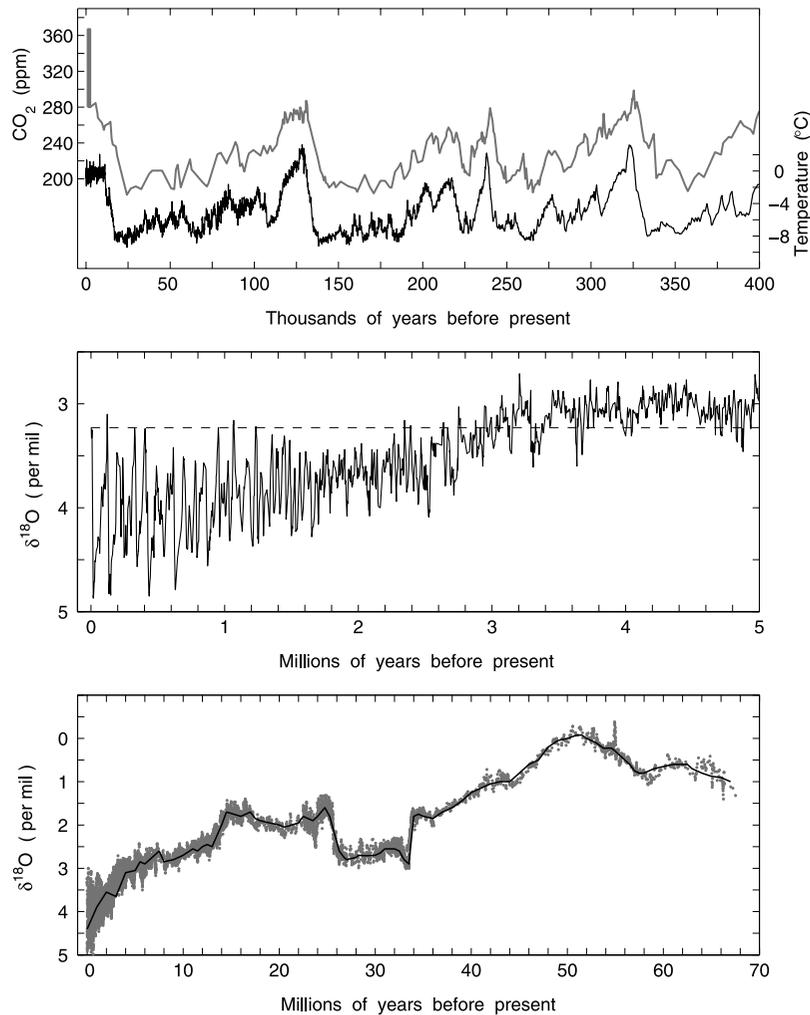
Figure 1. (top) The Vostok record showing temperature and CO_2 fluctuations over the past 400,000 years. The heavy vertical line along the left-hand axis shows the rise in CO_2 over the past century. After *Petit et al.* [1999]. (middle) Time series of $\delta^{18}O$ fluctuations, a measuer of ice volume, at site 659 in the eastern equatorial Atlantic. After *Tiedemann et al.* [1984]. (bottom) A composite of $\delta^{18}O$ record from Atlantic cores, as a proxy for deep ocean temperature for the past 60 million years. After *Lear et al.* [2000].

snow and an increase in the albedo of the Earth. This trend is reversed when the intensity of sunlight increases in summer, in high latitudes. Customarily, the intensity of sunlight at 65°N on 21 June, the date of the summer solstice in the northern hemisphere, is identified as the critical parameter that modulates the albedo feedback [*Crowley and North*, 1991]. One problem with this explanation is that sunlight in summer in polar regions is strongly affected by both the tilt of the Earth's axis and its precession. Why then does the response to variations in tilt, but not precession, amplify around 3 Ma? Apparently the climate changes of 3 Ma involve more than ice albedo feedbacks.

[5] Today, low sea surface temperatures are prominent aspects of intense oceanic upwelling in certain tropical and subtropical coastal and equatorial zones. That has not always been the case. *Kennet et al.* [1985] find that there was no surface temperature gradient along the equator in the early Miocene. Figure 2 shows that the surface temperatures were high until about 3 Ma, and then started to fall, in the

eastern equatorial Pacific [*Cannariato and Ravelo*, 1997; *Ravelo et al.*, 2001], and also off southwestern Africa [*Marlow et al.*, 2000]. Conditions off California [*Ravelo et al.*, 2001] and in the equatorial Atlantic (P. B. deMenocal, private communication, 2002) started changing in a similar manner at about the same time. Exceptionally warm surface water in the eastern equatorial Pacific is the signature of El Niño today. If we adopt that as the definition of El Niño then the Pacific experienced perennial El Niño conditions up to approximately 3 million years ago. (*Molnar and Cane* [2002] summarize much of the available evidence for the permanent El Niño conditions [see also *Chaisson and Ravelo*, 2000].) Is it possible that the processes that brought the perennial El Niño to an end around 3 Ma also contributed to the change in the response to Milankovich forcing, the increase in the amplitude of the 41K cycle?

[6] Today, a warming of the eastern equatorial Pacific, our definition of El Niño, is associated with an adiabatic horizontal redistribution of warm surface waters in the
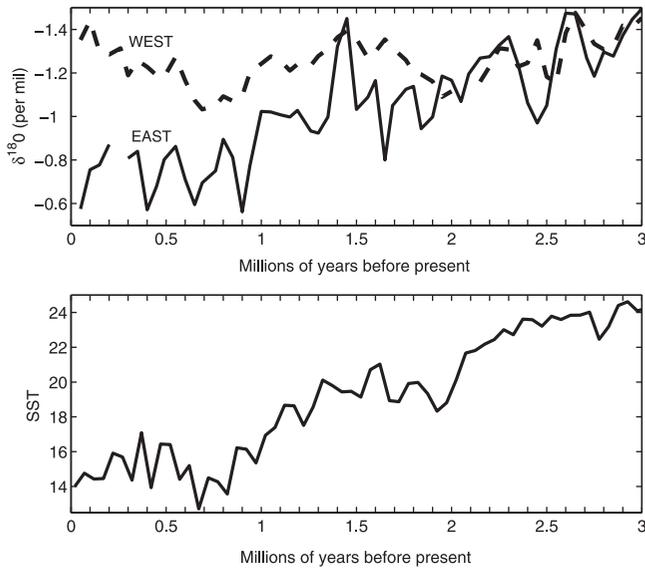
Figure 2. (top) Records from a site in the eastern equatorial Pacific (solid line) and from a site in the western equatorial Pacific (dashed line). The difference between the two signals indicates that the east-west gradient of sea surface temperature along the equator was absent earlier than about 3 million years ago. After *Ravelo et al.* [2001]. (bottom) Sea surface temperature record (in degrees centigrade) from a site off southwestern Africa. After *Marlow et al.* [2000]. The data have been smoothed with a Gaussian filter (100 kyr wide).

tropical Pacific. A very different, diabatic mechanism for warming the eastern equatorial Pacific is a net increase in the volume of warm surface waters and hence an increase in the spatially averaged depth of the thermocline. In this paper we argue that the global cooling during the Cenozoic gradually changed the thermal structure of the ocean. Specifically, we propose that a cooling of the deep ocean, documented by *Lear et al.* [2000] was accompanied by gradual shoaling of the thermocline. In the early Cenozoic, when temperatures in the deep ocean were in the neighborhood of 12°C, the thermocline was so deep that the winds were unable to bring deep, cold water to the surface in the upwelling zones of low latitudes. El Niño conditions were permanent. That continued to be the case, even as the thermocline shoaled gradually, until about 3 Ma. At that time, the thermocline was sufficiently shallow for further shoaling to reduce surface temperatures in upwelling zones. To substantiate these statements we discuss, in section 2, the factors that determine the oceanic stratification, including the depth of the thermocline. One of the implications of the discussion is that upwelling off Peru and off northwestern Africa must also have started around 3 Ma. Observations to test this hypothesis are as yet unavailable.

[7] Up to 3 Ma approximately the thermocline was so deep that sea surface temperatures were determined mainly by the local flux of heat across the ocean surface. Surface temperatures therefore tended to be uniformly high throughout the tropics and their patterns were similar to those in Figure 3a. Once the winds could bring cold water to the surface, temperatures patterns resembled those in Figure 3b. In that figure temperatures increase from east to west along the equator even though the intensity of sunlight is independent of longitude; the temperatures are at a maximum north of the equator even though sunlight is most intense at the equator. These curious asymmetries, which reflect the dynamical response of the ocean to the winds, in turn influence the winds, a circular argument which implies that interactions between the ocean and atmosphere are unstable. The introduction of those interactions, around 3 Ma, significantly affected the Earth's climate because they involve the two parameters that strongly influence climate: the atmospheric concentration of greenhouse gases, especially water vapor; and the albedo of the Earth, especially the areal extent of the highly reflective stratus clouds over cold tropical waters. Estimates of these effects for certain recent El Niño episodes are modest [*Soden*, 2000] because those episodes were modest in amplitude and duration. Even the intense El Niño of 1997 was very brief. To return to the warm world of 3 Ma and earlier we need to consider the indefinite persistence of the conditions that at the peak of El Niño of 1997 lasted merely for a few months. Furthermore, earlier than 3 Ma, sea surface temperatures were high, not only in the equatorial Pacific, but in upwelling zones throughout the tropics and subtropics. Conditions at that time corresponded to more than a perennial El Niño.
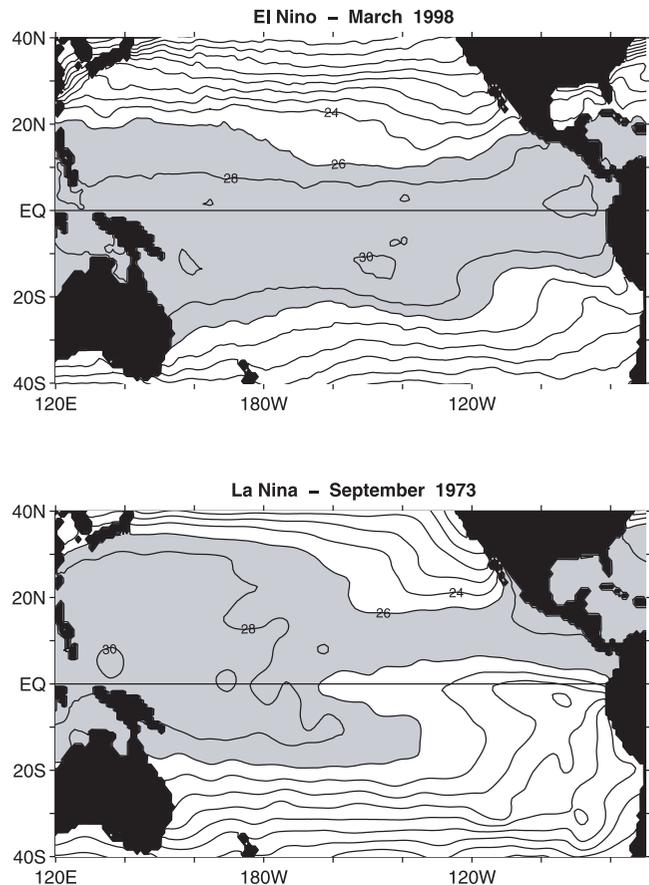
Figure 3. Sea surface temperature patterns (degrees centigrade) for (top) El Niño and (bottom) La Niña.
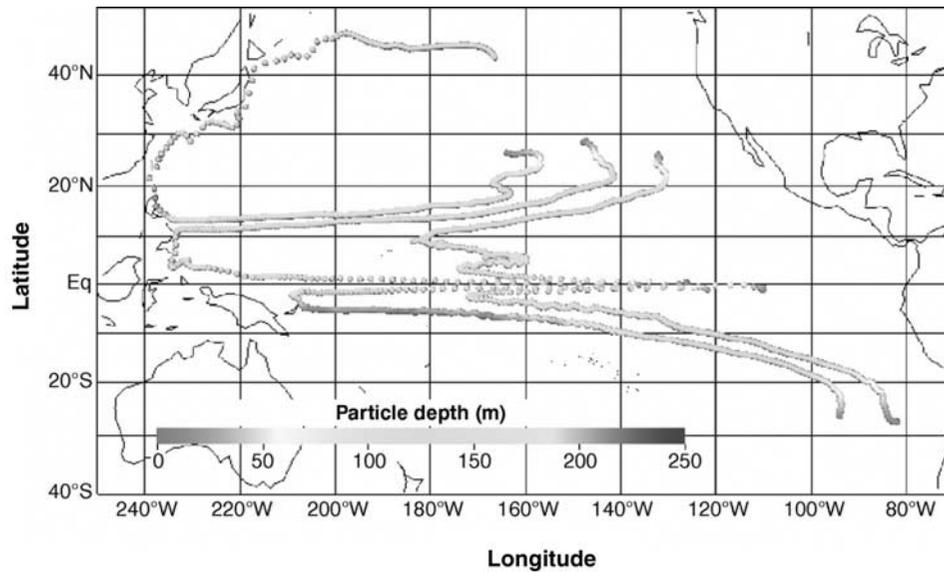
Figure 4. The paths of water parcels over a period of 16 years after subduction off the coasts of California and Peru as simulated by means of a realistic ocean model [*Harper*, 2000]. The colors indicate the depths of parcels; the distance between points is a measure of the speed of currents. Note the swift Kuroshio Current off Japan and the Equatorial Undercurrent. See color version of this figure at back of this issue.

[8] After discussing, in section 2, the factors that determine the oceanic stratification we turn, in sections 3 and 4, to the factors that influence the intensity of ocean-atmosphere interactions. These results are invoked to explain, in section 5, some of the phenomena that occurred around 3 million years ago. Section 6 explores why, around 3 Ma, the response to Milankovich forcing at a period of 41K but not 23K became important. Section 7 is a summary and discussion of the main results.

2. Thermal Structure of the Ocean

[9] Today, the thermocline is so shallow that a mere change in its tilt (downward to the west along the equator in the Atlantic and Pacific) strongly affects sea surface temperatures. Why is the thermocline so shallow? What processes maintain the thermocline and determine its spatially averaged depth? (The following discussion addresses these questions from a strictly oceanographic perspective so that surface conditions, the winds for example, are regarded as specified.) Oceanographers realized in the 18th century already that the cold waters that fill the deep ocean have their origin at the surface in the high latitudes. At first it was thought that this water gradually rises back into the upper ocean essentially everywhere, thus balancing the downward diffusion of heat and maintaining the thermocline. However, such a state of affairs implies values for oceanic diffusion that are very high, an order of magnitude larger than those actually measured in the thermocline [*Ledwell et al.*, 1993]. This result motivated studies of an ocean with very low diffusivities, except in a very shallow surface Ekman layer [*Luyten et al.*, 1983]. The wind-driven circulation in the upper layers of such an ocean is then the one shown in Figure 4. Water subducts in the subtropics off the western

coasts of continents, penetrates to a few hundred meters at most, and flows along isopycnal surfaces either to low latitudes, or participates in a subtropical gyre. The water that wells up at the equator returns to the subduction zone in poleward Ekman drift in the surface layer. Hence the vertical temperature profile of this ventilated thermocline is a reflection of the latitudinal temperature profile at the surface. The ventilated thermocline has been studied extensively [*Pedlosky*, 1996], but the available theories are incomplete. The theories require that the thermal structure of the ocean in the absence of any winds be given, and then describe how the winds alter that structure, deepening the thermocline in some regions, shoaling it elsewhere. In the case of a simple two-layer ocean this indeterminacy of the theory is explicit and appears as two parameters that need to be specified:

$$H^2 = H_e^2 + 2f^2(t/f)_y(x - x_e)/g'\beta \qquad (1)$$

This expression describes how an imposed wind stress τ causes the depth of the thermocline H to vary as a function of longitude x and latitude y. Here f is the Coriolis force and β is its latitudinal (y) gradient. The two parameters that need to be specified are H_e the depth of the thermocline along the eastern coast $x = x_e$, and the reduced gravity g' which is directly proportional to ΔT, the temperature difference across the thermocline. Because the temperature of the deep ocean is determined by the thermohaline circulation, that deep circulation is implicit in the theories for the wind-driven circulation of the ventilated thermocline. (In equation (1), the constant g', because it is proportional to ΔT, parameterizes the thermohaline circulation.) The parameter H_e remains an undetermined para-

meter. Its value is simply specified (as the depth of the thermocline at the eastern boundary) in most studies of the ventilated thermocline.

[10] What constraint determines the value for H_e, or equivalently, the volume of warm water in the upper ocean? *Bocaletti et al.* [2003] propose a balanced heat budget for the ocean as the relevant constraint. This is a constraint on the areal integral of the heat flux across the ocean surface and can be satisfied in an infinite number of ways. One possibility is for that flux to be zero locally, everywhere, at each point in the ocean. That is not the state of affairs today but is more likely to be the case when upwelling zones disappear, for example, in the warm world earlier than 3 Ma. In such a world the first term on the right-hand side of equation (1) has the same parameter dependence as the second term [*Bocaletti et al.*, 2003] so that the depth of the thermocline is proportional to $(\tau/\Delta T)^{1/2}$. An increase in ΔT, or a cooling of the deep ocean, therefore translates into a shoaling of the thermocline. (If the oceanic stratification is weak, if the value of g' tends to zero, then the effect of the wind penetrates to considerable depths and the thermocline is very deep.) These results imply that, during the Cenozoic, as the deep ocean cooled and the density difference across the thermocline increased, the thermocline became shallower and shallower. (This assumes that the winds remain unchanged, a matter to which we return shortly.)

[11] The appearance of upwelling zones in low latitudes around 3 Ma meant that the ocean had a net gain of heat in the tropics. In a state of equilibrium, the gain must be balanced by a loss elsewhere so that the net flux of heat across the ocean surface can be zero. Today the regions of heat gain and loss are those shown in Figure 5. The oceanic circulation effects a poleward transport of heat, from the upwelling zones in low latitudes where the ocean gains heat, to the high latitudes where the ocean loses heat. The processes that control the gain and loss of heat are practically independent of each other. The loss, in high latitudes, depends mainly on atmospheric conditions, for example on the temperature of cold air masses moving off continents over the warm Gulf Stream and Kuroshio currents. The gain of heat in the upwelling zones of low latitudes depends on oceanic processes such as the depth of the thermocline. The heat gain decreases when the thermocline deepens (during El Niño for example) because sea surface temperatures and hence evaporation increases. When the thermocline shoals, the heat gain increases. To maintain a balanced heat budget, an increase in the oceanic heat loss in high latitudes requires an increase in the heat gain in low latitudes. The ocean accomplishes this by shoaling the tropical thermocline, and by increasing the heat transported poleward by the wind-driven circulation of the ventilated thermocline. Conversely, reduced heat loss in high latitudes results in deeper tropical thermocline and a smaller poleward transport of heat.

[12] These results have been confirmed by means of calculation using the model similar to the one that produces the results in Figure 4. In such models the thermohaline circulation is parameterized. The circulation of the ventilated thermocline shown in Figure 4 is obtained by forcing a realistic, high-resolution general circulation model of the ocean with observed winds. The model starts from a
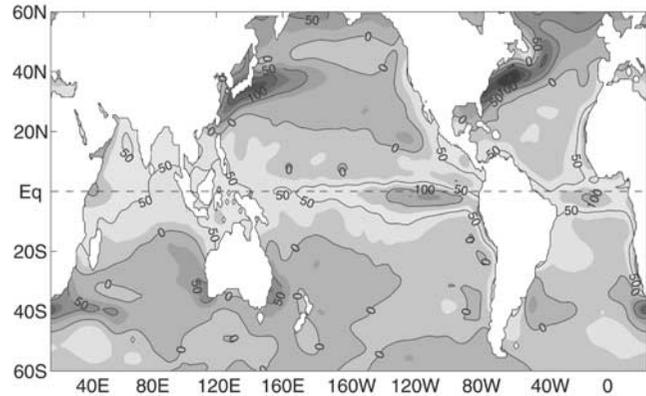
Figure 5. Annual mean net heat flux across the ocean surface in watts/m^2 [after *Da Silva et al.*, 1994]. The heat gain in the upwelling zones of low latitudes equals the heat loss in high latitudes. See color version of this figure at back of this issue.

specified, realistic thermal structure for the ocean. Calculations continue for a few decades to permit the upper ocean to adjust to the imposed (realistic) winds and surface heat flux conditions. Densities in the deep ocean remain essentially the same as specified initially and represent the implicit thermohaline circulation of this model. This approach exploits the difference in the timescales of the ventilated thermocline (a few decades) and the thermohaline circulation (a few centuries). Only the wind-driven circulation of the upper ocean is allowed to come into equilibrium with the winds. To explore how the upper ocean is affected by changes in the thermohaline circulation, it is necessary to alter the temperature of the deep ocean, and to repeat the calculations for a few decades so that the upper ocean can adjust. Such calculations confirm the inferences made from equation (1): a cooling of the deep ocean can cause a shoaling of the thermocline. Additional calculations indicate that the heat budget at the surface does indeed have a strong influence on thermocline depth on timescales of decades. (A more detailed description of these calculations is given elsewhere [*Bocaletti et al.*, 2003].) Recent analyses of paleoclimate data from the equatorial Pacific and Indian Oceans are consistent with these theoretical results. *Beaufort et al.* [2001] find that, over the past 200,000 years, the thermocline in the equatorial Pacific was shallower during glacial periods, deeper during interglacial periods.

[13] In summary, the absence of tropical and subtropical oceanic upwelling zones during most of the Cenozoic suggests that the heat budget was in approximate balance locally at each point in the ocean so that horizontal heat transports in the ocean were of secondary importance. During this period the thermocline gradually shoaled, without significantly affecting sea surface temperatures. This was the state of affairs until about 3 Ma at which stage the thermocline was sufficiently shallow for the wind to bring cold water from below the thermocline to the surface in upwelling zones. The heat budget at the surface then moved further and further from a local balance, toward one in which the poleward transport of heat became

important. This introduced new factors that influenced the thermocline depth, and increased the sensitivity of the Earth's climate to Milankovich forcing. Before discussing that issue, we turn our attention to the factors that control the intensity of the ocean-atmosphere interactions that came into play around 3 Ma.

3. Ocean-Atmosphere Interactions in the Tropics

[14] Interactions between the ocean and atmosphere require a thermocline sufficiently shallow for the winds to bring cold water to the surface. How shallow does the thermocline have to be? An answer is available from measurements in the Pacific of today. Along the equator in that ocean, the thermocline slopes downward to the west. It is so deep in the neighborhood of the dateline, on the order of 150 meters, that its vertical movements there have essentially no influence on local surface temperatures. In the eastern equatorial Pacific, where the thermocline is less than 50 meters deep, its vertical movements are highly correlated with surface temperature variations. (This is the reason for the success of shallow water models of the ocean in simulating El Niño.) Along the coast of Peru it has similarly been observed that, even when there is little change in the local winds, the temperature of the water that wells up is high when the thermocline is deep, during El Niño for example. The main conclusion to be drawn from these observations is that a transition to persistent El Niño conditions requires a relatively modest change, in the thermal structure of the tropical oceans. A deepening of the thermocline by merely tens of meters could prove adequate.

[15] Stability analyses of ocean-atmosphere interactions confirm the inferences made in the previous paragraph. These analyses fall into two categories: (i) those concerned with interannual, oscillatory modes involving a rapid, horizontal redistribution of warm surface waters across the equatorial Pacific; and (ii) those concerned with long-term conditions, the time-averaged intensity of the trade winds for example. In the case of (i) we are interested in how the properties (period, amplitude, etc.) of the oscillations depend on background conditions described by parameters such as the spatially averaged depth H of the thermocline, the temperature difference ΔT across the thermocline, and the intensity τ of the trade winds at the equator. For which values of these parameters are there no oscillatory modes? Stability analyses indicate that, today, the observed Southern Oscillation, between complementary La Niña and El Niño conditions, is marginally unstable [*Neelin et al.*, 1998; *Fedorov and Philander*, 2001]. A modest increase in the values of H or ΔT, or a decrease in the value of the time-averaged zonal winds can result in the disappearance of the oscillations because the ocean-atmosphere interactions on which they depend become too weak.

[16] When our interest shifts to the dependence of the background conditions themselves on ocean-atmosphere interactions then we focus on a variable such as the time-averaged sea surface temperature in the eastern equatorial Pacific and ask how sensitive that temperature is to modest changes in the values of H, τ and ΔT. Figure 6 provides some answers. (To obtain these results we used methods
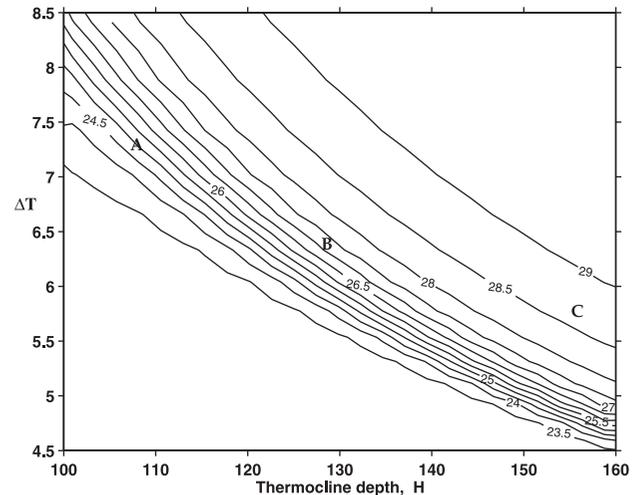
Figure 6. The effect, on sea surface temperatures in the eastern equatorial Pacific, of changes in the spatially averaged depth of the thermocline H, and the temperature difference across the thermocline ΔT, two parameters that affect the intensity of ocean-atmosphere interactions (after A. Fedorov and S. G. Philander, How climatological conditions over tropical oceans depend on ocean-atmosphere interactions, submitted to *Journal of Climate*, 2003, hereinafter referred to as Fedorov and Philander, submitted manuscript, 2003).

similar to those of *Dijkstra and Neelin* [1995]. Conditions today, which correspond to the neighborhood of point A in Figure 6, are seen to be such that a small change in H or ΔT results in dramatic changes in surface conditions and hence in climatic conditions. If the thermocline were to deepen slightly then the winds would bring warmer water to the surface, and would thus decrease the zonal temperature gradient. This would cause the winds to weaken, causing even warmer surface waters and so on. For large values of H, variations in the depth of the thermocline have little effect on surface conditions because the winds are unable to bring cold water below the thermocline to the surface. For a shallow thermocline (values of H much smaller than 100 meters if we are at point A in Figure 6) the results are again insensitive to changes in H. This happens because the water below the thermocline has a uniform temperature (in this model) so that more intense entrainment of cold water does not result in lower temperatures.

[17] We postulate that conditions during the Miocene, when the thermocline was deeper than today and when ΔT had a smaller value, corresponded to the neighborhood of point C in Figure 6. At that time surface conditions were insensitive to modest changes in the values of H or ΔT so that ocean-atmosphere interactions were inhibited. Subsequently the global cooling resulted in a move toward B which was reached around 3 Ma. Today we are near point A.

[18] Ocean-atmosphere interactions can have a surprising effect on the response of the eastern equatorial Pacific to an increase in solar radiation. In the absence of those interactions, more intense sunlight causes a rise in surface temperatures. In the presence of those interactions surface
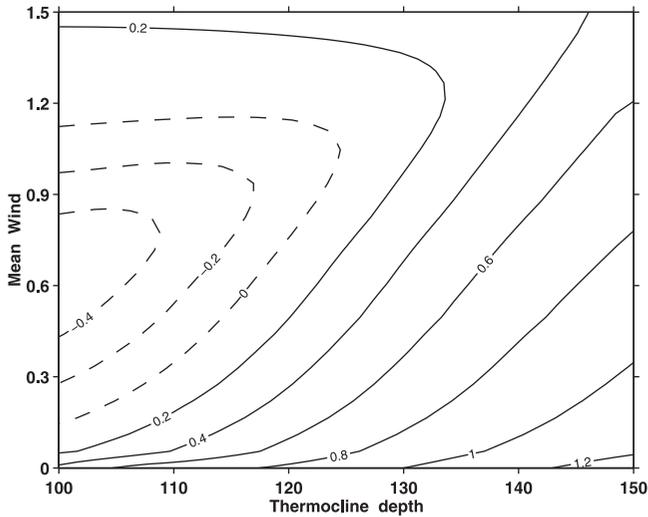
Figure 7. The response of sea surface temperatures in the eastern equatorial Pacific to a higher value of solar radiation, for different values of the thermocline depth H, and the intensity of the zonal wind stress τ (Fedorov and Philander, submitted manuscript, 2003).

temperatures depend mainly on the local flux of heat in the western equatorial Pacific but matters are different in the east. There, wind-induced upwelling strongly influences surface temperatures. It is possible for more intense sunlight to have a negligible effect in the east. If temperatures increase in the west, but remain almost unchanged in the east, then the zonal temperature gradient increases. That gradient intensifies the surface winds, which cause more upwelling. Hence the net effect of more intense sunlight can be lower surface temperatures in the east. The result clearly depends on the degree to which ocean-atmosphere interactions come into play and hence depends on factors such as the thermocline depth. Figure 7 shows how the surface temperature in the eastern equatorial Pacific is affected, for different values of the winds and thermocline depth, when the increase in solar radiation is 5 watts/m^2. (The model is essentially the same as the one that produces the results in Figure 6.) For a deep thermocline, and weak winds τ, ocean-atmosphere interactions are weak and the response is as expected: the surface temperatures increase. As the thermocline shoals, the rise in temperature becomes smaller until it is zero along the dashed line. Inside the dashed line ocean-atmosphere interactions cause the temperatures in the eastern Pacific to decrease in response to more intense sunlight.

4. Ocean-Atmosphere-Land Interactions Along Coasts

[19] Sea surface temperature variations in coastal zones such as the one off California are determined mainly by two factors that are often in opposition: in summer, a rise in temperatures because of more intense sunlight is countered by more intense upwelling because of stronger alongshore winds. The winds are associated with an offshore zone of high pressure. Why is that pressure maximum during summer when sunlight is most intense? (Usually heating results in lower surface pressures.) The answer to this puzzle involves ocean-atmosphere-land interactions that are, as yet, practically unexplored. Such interactions probably played a crucial role in the climate changes of 3 Ma.

[20] The air that rises in the convective zones of low latitudes, in regions of low surface pressure, descends in the subtropics where the surface pressure is high. These features would be essentially independent of longitude on a water-covered globe. The presence of continents introduces zonal variations; in summer the subtropical high-pressure zone is prominent over the oceans, while pressure is lower over the continents because they attain higher temperatures. If the sea surface temperatures depend strictly on the flux of heat across the ocean surface, which is the case when the thermocline is sufficiently deep, then convection in low latitudes is shared by ocean and land, provided the land is saturated with moisture. If certain land processes cause moisture levels on land to be relatively low, then the convection is mainly over the warmest part of the oceans, in the lowest latitudes [*Chou et al.*, 2000]. In the case of a shallow thermocline that allows winds to bring cold water to the surface, feedbacks between the ocean, atmosphere and land come into play [*Hoskins*, 1996; *Rodwell and Hoskins*, 2001]. Now the winds cause a lowering of sea surface temperatures so that convection occurs mainly over land. The air that rises there subsides westward and poleward, over the adjacent ocean, suppressing convection there and intensifying the offshore high-pressure zone. The associated winds induce coastal upwelling and an offshore Ekman drift of cold water that further intensifies the high-pressure zone and the winds. Highly reflective stratus clouds tend to form over these cold waters. Hence the difference between worlds with and without coastal upwelling could amount to a significant change in the water vapor content of the atmosphere, and in the albedo of the Earth.

5. Transition From Perennial to Intermittent El Niño

[21] Any attempt to identify climate changes associated with a gradual elevation of the thermocline must keep a number of complicating factors in mind. At any time, the winds cause the depth of the thermocline to have considerable spatial variations. For example, the thermocline is shallow near the equator and deepens as we start moving poleward, in accord with theories for the ventilated thermocline (see equation (1)). The thermocline also deepens from east to west because of the prevailing trade winds in the tropics. Because of these spatial variations, a gradual shoaling of the thermocline will have surface manifestations at different times in different regions. This happened during the transition from El Niño to La Niña in 1998. An isolated patch of cold surface waters first appeared along the equator to the west of the Galapagos Islands in mid-1998. At that time the surrounding waters were still very warm, and the region of equatorial upwelling was separated from the regions of coastal upwelling along the western coasts of the Americas. Conditions were essentially symmetrical about the equator. Subsequently the upwelling zones off

Peru and along the equator became linked, and warm surface waters were confined to a narrow band to the north of the equator. There are indications that the transition from a warm to a cold world during the Pliocene followed a similar sequence, over a much, much longer period. Hence the various climate changes in different parts of the globe around 3 Ma are unlikely to have been simultaneous, but probably occurred in a definite order. *Chaisson and Ravelo* [2000] and *Ravelo et al.* [2001] propose that cold surface waters in a low latitude upwelling zone first appeared in the eastern equatorial Pacific around 2 Ma. It is possible that, at that time, conditions were still symmetrical about the equator so that atmospheric convection was possible over the warm waters on both sides of the equator. Under those conditions, tropical convection was not necessarily confined to a narrow Intertropical Convergence Zone (ITCZ). (For example, today, over the western tropical Pacific, the convection covers broad areas and is not confined to a narrow zone.) (*Hovan* [1995] and *Billups et al.* [1999] report an equatorward shift of the ITCZ around 4 Ma but it is not clear that a narrow ITCZ existed at that time.)

[22] As the thermocline continued to shoal, cold surface waters started appearing in the coastal upwelling zones. At some stage the upwelling off Peru became linked to that along the equator, asymmetries relative to the equator appeared, and warm surface water was confined to a narrow band north of the equator. Now atmospheric convection in the eastern tropical Pacific was limited to a narrow ITCZ. This final transition from perennial to intermittent El Niño and La Niña, presumably around 3 Ma, would have contributed to drier, cooler conditions in the eastern tropical Pacific, and also in eastern equatorial Africa. (The change in the neighborhood of the Galapagos Islands is consistent with the DNA analysis of Darwin's finches on those islands [see *Grant*, 1999].)

6. Response to Milankovich Forcing

[23] Before 3 Ma, in the absence of low surface temperatures in tropical and subtropical upwelling zones where the ocean today gains most of its heat, the oceanic heat budget was balanced mostly locally. Sea surface temperature variations were then determined mostly by processes that are captured by a one-dimensional mixed layer model that neglects horizontal currents. In such a warm world, the Milankovich forcing induced, in low latitudes, mainly variations in the contrast between land and sea and hence affected mainly the monsoons. This response changed around 3 Ma because the appearance of cold surface waters in the upwelling zones of low latitudes shifted the oceanic heat budget from a local to a nonlocal balance. Since 3 Ma the ocean has been gaining heat in the upwelling zones, and its circulation has been transporting that heat to higher latitudes to be lost there. This means that, since 3 Ma, conditions in high latitudes have been influencing the depth of the thermocline in low latitudes as explained in section 2. For example, since 3 Ma increased oceanic heat loss in high latitudes has translated into a shallower tropical thermocline and hence increased heat gain. The balance between loss and gain, and hence the connection between oceanic con-

ditions in high and low latitudes, is possible only on the timescale of the ventilated thermocline circulation which transports the heat poleward, a few decades at least. Hence a balanced heat budget is not a constraint on a seasonal timescale which is much too short. Seasonal fluctuations in atmospheric conditions in high latitudes can affect the local, seasonal flux of heat across the ocean surface, but can not influence oceanic conditions in the tropics. A minimum timescale for a connection between low and high latitudes in the ocean is one reason why, around 3 Ma, Milankovich forcing started inducing different responses at periods of 23,000 and 41,000 years.

[24] Milankovich forcing amounts mainly to a modification of the seasonal forcing. Precessional variations redistribute sunlight incident on the Earth over the course of a year. The intensity of sunlight increases in summer and decreases in winter in one hemisphere, while doing the opposite in the other hemisphere. This is evident in the top panel of Figure 8. The net effect of the precessional change, integrated over a year, is zero. Such fluctuations, as explained in the previous paragraph, are too rapid to affect the oceanic heat budget which is balanced on timescales of decades and more. Fluctuations in the distribution of sunlight associated with changes in tilt are very different because they have a very long timescale.

[25] Figure 8 (bottom) shows how an increase in tilt by 2 degrees, without any changes due to precession, affects the distribution of sunlight. The net amount of sunlight incident on the Earth is unchanged but there is a latitudinal redistribution: the summer hemisphere receives more sunlight, especially near its pole, while the winter hemisphere receives less. (In the bottom panel of Figure 8 the larger amplitude in polar regions is offset by the smaller area of those regions.) At a fixed latitude the fluctuations in sunlight, averaged over a year, are not identically zero, as in the case of precessional changes, but are nonzero and vary with latitude as shown in the right-hand panel. When tilt is large, the intensity of sunlight in high latitudes increases. Such conditions persist for half the obliquity cycle, for some 20,000 years. Because of the constraint of a balanced heat budget, the reduced oceanic heat loss in high latitudes is associated with a deepening of the tropical thermocline, with persistent El Niño conditions. When tilt is small, conditions correspond to La Niña. These arguments imply that the changes in atmospheric and oceanic conditions from small to large tilt should include: a deepening of the thermocline throughout the tropics and subtropics; weakened coastal upwelling off the western coasts of the Americas and Africa in both hemisphere; increased sea surface temperatures in the eastern equatorial Pacific; a weakened slope of the equatorial thermocline; equatorward movement of the ITCZ; increased rainfall over the eastern equatorial Pacific and eastern equatorial Africa. In addition, there would have been a fluctuation (at a period of 41K) in the properties of the interannual Southern Oscillation between El Niño and La Niña, changes in its period, etc.

[26] These results find support in analyses of a seafloor core off northwestern Africa which indicate that, since 3 Ma, the obliquity fluctuations have been in phase with fluctuations, at the same period, in local thermocline depth
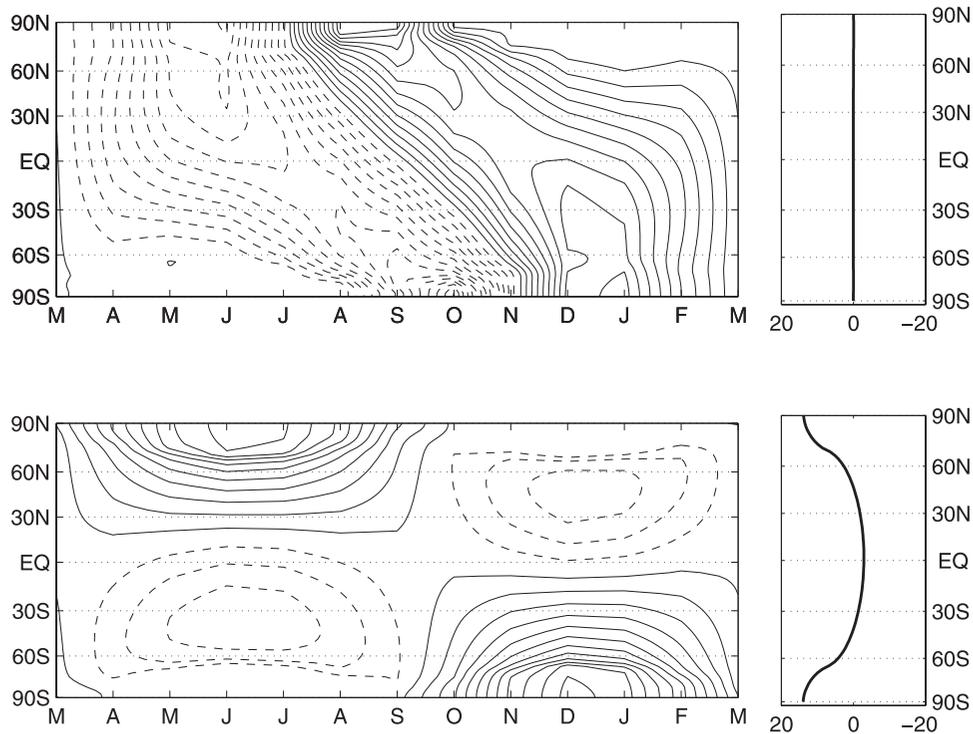
Figure 8. Changes in the incident sunlight, as a function of month and latitude (top) when perihelion changes from its current date (in January) to six months later without any change in obliquity and (bottom) when obliquity changes from 22 to 24 degrees without any change in precessional effects. The time integral over 12 months of the changes in sunlight as a function of latitude, in the right-hand panels, is identically zero for changes in precession. Solid lines indicate an increase in sunlight; dashed lines a decrease. The contour interval is 1.4 watts/m^2.

(P. B. deMenocal, private communication, 2001). A similar result is available for data that describe the stratification of the upper ocean in the northern Pacific (G. H. Haug, private communication, 2001). This agreement between observations and theory is very tentative because the method used to date the cores from the ocean floor introduces considerable uncertainty in the determination of the phase.

[27] Variations in obliquity can induce variations in thermocline depth and hence in the intensity of ocean-atmosphere interactions. This will influence a variety of regional phenomena, and also the global climate (by affecting the atmospheric concentration of water vapor, and the areal extent of stratus clouds over regions of upwelling.) To what degree did the impact of the tropics on the global climate affect the waxing and waning of glaciers? P. B. deMenocal (private communication, 2001) finds that, off northwestern Africa, at a period of 41K, fluctuations in thermocline depth led those in ice volume by some 5000 years. Given that glaciers grow very slowly, a lag should be expected even if sunlight at 60 N is their controlling factor. Hence the lag described by deMenocal does not preclude the possibility that fluctuations in ice volume depend mainly on ice albedo feedbacks, while conditions in the tropics are contributing factors.

[28] Lea [2001] recently summarized evidence that, during the last glacial maximum, the melting of continental ice sheets lagged some thousands of years behind increases in sea surface temperatures in certain regions. For example, temperatures off California started rising during the coldest part of the ice age. Herbert et al. [2001] propose that the changes off California are linked to the presence of the massive ice sheet over northern America at the time, but it is entirely possible that the changes are related to variations in the depth of the thermocline that modulate ocean-atmosphere interactions in low latitudes. What then matters is not the intensity of sunlight at 60 N in June, but the tilt of the Earth's axis for example.

7. Discussion and Summary

[29] The climate changes that the Earth experienced around 3 Ma were a long time in the making. Over millions of years, gradual global cooling, and associated changes in the oceanic thermal structure, ultimately led to the appearance of continental glaciers in the northern hemisphere, and independently to the emergence of low sea surface temperatures in equatorial and coastal upwelling zones. The latter phenomena led to changes in the patterns of atmospheric convection in low latitudes, amplified versions of those that accompany the transition from El Niño to La Niña today, thus contributing to significant regional and global climate changes. The introduction of feedbacks involving ocean-atmosphere interactions around 3 Ma altered the Earth's response to Milankovich forcing, mainly at a period of 41K,

because only at that period does the forcing amount to a change in the intensity of sunlight averaged over a year at a fixed latitude. The additional feedbacks that came into play some 1 Ma ago and caused further changes in the response to Milankovich forcing, an amplification of the 100K signal for example, are yet to be identified.

[30] The drifting of the continents over the past 60 million years continually changed the Earth's geography and, among others, opened and closed various critical seaways, thus influencing climate. For example, around 30 Ma, the opening of an uninterrupted seaway around Antarctica was accompanied by a fall in temperatures in high southern latitudes [*Zachos et al.*, 2001], presumably because of changes in the oceanic circulation [*Toggweiler and Samuels*, 1995]. The more recent closure of the Indonesian seaway [*Cane and Molnar*, 2001], and of the Panamanian seaway [*Stehli and Webb*, 1985; *Mikolajewicz and Crowley*, 1997] undoubtedly affected the oceanic circulation, especially the deep thermohaline circulation. Whether the associated changes in the depth of the thermocline influenced the global climate, is doubtful because at the time, some 4.5 Ma or earlier, the depth of the thermocline was such that its vertical movements had little surface manifestation. (See, e.g., Figure 2, which shows that low surface temperatures in upwelling zones occurred only around 3 Ma.) The paleo-records provide plentiful evidence of changes in the thermohaline circulation. To what extent those changes contributed to climate changes is unclear.

[31] Of crucial importance to the various climate changes around 3 Ma was the role played by the oceanic circulation which has two components: (i) the shallow wind-driven circulation of the ventilated thermocline which adjusts to changes in forcing on a timescale of a few decades; and (ii) the deep thermohaline circulation which has a timescale of millennia. Whereas (ii) transports heat northward mainly in the Atlantic, (i) contributes to that northward transport and, in addition, carries a large amount of heat poleward in the Pacific and southward in the Indian Ocean. Of direct importance to the atmosphere is not the oceanic heat transport but the sea surface temperature patterns. Most previous climate studies have focused on (ii) and find that changes in (ii) influence sea surface temperatures mainly in the far northern Atlantic with only a very modest effect in the tropics. (See, e.g., *Manabe and Stouffer* [1988, Figure 1] for the effect of a shut down of the thermohaline circulation.) This study focuses on (i) and argues that changes in the

circulation of the ventilated thermocline have a strong influence on sea surface temperatures in the upwelling zones of low latitudes. In reality both (i) and (ii) are important; that is, only the thermohaline circulation could have effected the cooling of the deep ocean over the past 60 million years, but the connections between (i) and (ii) are at present unclear. Both involve vertical conveyor belts, with surface water sinking in high latitudes in the case of (ii), in the subduction zones of the subtropics in the case of (i). Water returns to the surface in oceanic upwelling zones in the case of (i). Exactly where the water participating in the thermohaline circulation rises to the surface is at present a major puzzle. Another puzzle concerns the factors that determine the intensity of the thermohaline circulation. Mixing processes in the deep ocean are likely to be important to the answers of both puzzles. The distinction between (i) and (ii) blurs when mixing across the thermocline is strong. That may be the case in some models but is not the case in reality. Diffuse Kuroshio's and Gulf Streams, with little heat transport in the ventilated thermocline, because of coarse horizontal resolution in certain models, also blur the distinctions between (i) and (ii). Whether, at present, there are climate models that treat both (i) and (ii) realistically is unclear. On the more theoretical side it is also unclear which factors determine the partitioning of heat transport between (i) and (ii). Many issues remain to be explored.

[32] The climate changes that occurred some three million years ago are of considerable interest in connection with the rapid rise in the atmospheric concentration of carbon dioxide over the past century (which is represented by the vertical line in Figure 1 (top)). That figure shows that because the present happens to be a warm interglacial period and because the Earth's response to Milankovich forcing has grown significantly over the past million years, conditions today are close to those of three million years ago. Is it possible that the current rise in the concentration of greenhouse gases, an exponentially growing perturbation, can cause a return of permanent El Niño conditions?

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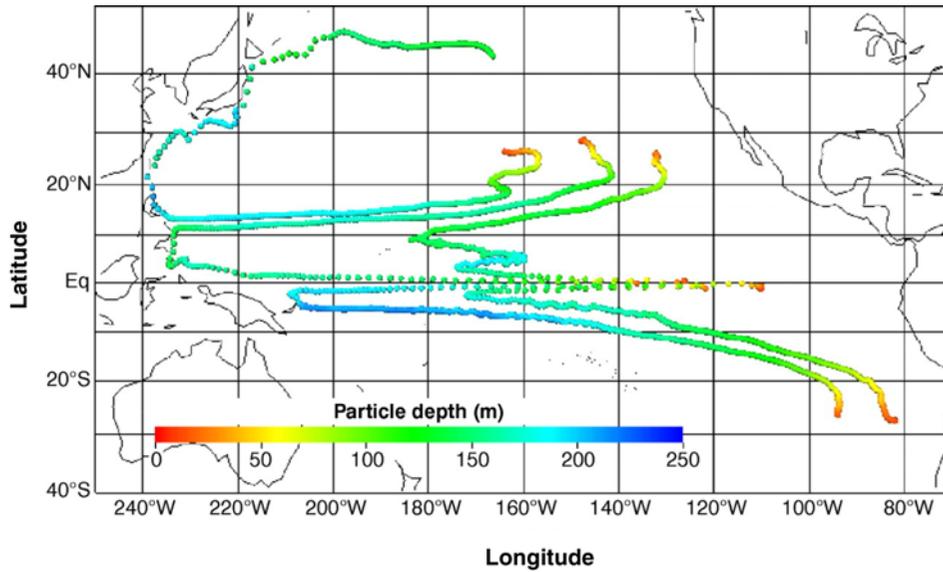
Figure 4. The paths of water parcels over a period of 16 years after subduction off the coasts of California and Peru as simulated by means of a realistic ocean model [*Harper*, 2000]. The colors indicate the depths of parcels; the distance between points is a measure of the speed of currents. Note the swift Kuroshio Current off Japan and the Equatorial Undercurrent.

Figure 5. Annual mean net heat flux across the ocean surface in watts/m^2 [after *Da Silva et al.*, 1994]. The heat gain in the upwelling zones of low latitudes equals the heat loss in high latitudes.