# Ocean Dynamics, Thermocline Adjustment, and Regulation of Tropical SST

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ABSTRACT

The role of tropical Pacific ocean dynamics in regulating the ocean response to thermodynamic forcing is investigated using an ocean general circulation model (GCM) coupled to a model of the atmospheric mixed layer. It is found that the basin mean sea surface temperature (SST) change is less in the presence of varying ocean heat transport than would be the case if the forcing was everywhere balanced by an equivalent change in the surface heat flux. This occurs because the thermal forcing in the eastern equatorial Pacific is partially compensated by an increase in heat flux divergence associated with the equatorial upwelling. This constitutes a validation of a previously identified ''ocean dynamical thermostat.''

A simple two-box model of subtropical–equatorial interaction shows that the SST regulation mechanism crucially depends on spatial variation in the sensitivity of the surface fluxes to SST perturbations. In the GCM, this sensitivity increases with latitude, largely a result of the wind speed dependence of the latent heat flux, so that a uniform forcing can be balanced by a smaller SST change in the subtropics than in equatorial latitudes. The tropical ocean circulation moves heat to where the ocean more readily loses it to the atmosphere. Water that subducts in subtropical latitudes and returns to the equatorial thermocline therefore has a smaller temperature perturbation than the surface equatorial waters. The thermocline temperature adjusts on timescales of decades to the imposed forcing, but the adjustment is insufficient to cancel the thermostat mechanism.

The results imply that an increase in the downward heat flux at the ocean surface, as happens with increasing concentrations of greenhouse gases, should be accompanied by a stronger equatorial SST gradient. This contradicts the results of coupled atmosphere–ocean GCMs. Various explanations are offered. None are conclusive, but the possibility that the discrepancy lies in the low resolution of the ocean GCMs typically used in the study of climate change is discussed.

## 1. Introduction

The tropical sea surface temperature (SST) pattern is strongly influenced by ocean dynamics. In some areas of the ocean, the net surface heat flux, and hence the ocean heat transport, is small and the SST is determined by a one-dimensional balance at the surface. This is the case for the west Pacific and west Atlantic warm pools. In contrast, upwelling on the equator and at the eastern boundaries of the Atlantic and Pacific creates a dynamical heat flux divergence that partially balances net surface heating. Oceanic heat flux divergence in the Tropics acts to lower tropical mean SST as is evident in the higher SSTs that result in numerical experiments in which the ocean heat transport is set to zero (e.g., Meehl and Washington 1986; Seager et al. 1988).

The spatial pattern of upwelling and SST is determined through coupled air–sea interaction. In the presence of eastern and western boundaries, mean easterlies cause equatorial upwelling, a thermocline that shoals toward the east and an equatorial cold tongue with a maximum intensity in the east. Warm SSTs induce convective heating in the atmosphere in the west, which forces surface easterlies over the basin. The wind response therefore feeds back positively, and coupled interactions establish the tropical climatology (Bjerknes 1969; Dijkstra and Neelin 1995).

The importance of coupled interactions in establishing the tropical climate suggests they will also play a role in climate change and variability. This is obvious for the case of interannual variability (Zebiak and Cane 1987) but should also be true for longer period variability, including the response to external forcings (e.g., increased greenhouse gases or Milankovitch cycles). However, in the literature on climate change, outside of areas of deep water formation, the ocean is commonly reduced to a passive, dynamically uncoupled, diffusive

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medium that merely acts to delay climate changes forced by radiative imbalances (e.g., Hansen et al. 1981; Lindzen 1993).

The notion of a dynamically inert ocean has been challenged by Clement et al. (1996; CSCZ hereafter) and by Sun and Liu (1996). Using a simple coupled model of the tropical Pacific, CSCZ demonstrate how equatorial ocean dynamics and associated ocean heat transport can act to oppose an imposed forcing and lead to changes in SST that are less than would be the case for local surface heat flux equilibration. Further, the coupled interaction alters the tropical climatology and the interannual variability. The dynamical model of CSCZ, and the box model of Sun and Liu (1996), are open to criticism because the temperature of the thermocline water that is upwelled does not respond to the variations in SST. Will thermocline adjustment eventually wipe out these responses?

Here we perform some experiments with a forced ocean general circulation model (GCM) that computes its own thermocline temperature. The experiment is again trivially simple. A uniform forcing is imposed, the model run to equilibrium, and the results examined to see whether the changes in the tropical SST identified by CSCZ can survive as the thermocline adjusts. Since our ocean GCM is uncoupled, and because we expect the coupled dynamics to enhance the ocean response, this would be a persuasive confirmation of the mechanism identified by CSCZ.

The purpose of this paper is to defend the contention that tropical ocean dynamics can play an active role in long-term climate variability and climate change. Their presence alters the pattern and magnitude of tropical SST change. Coupled GCMs that include ocean GCMs with low horizontal resolution, as is typically the case (e.g., Knutson and Manabe 1995; Tett 1995), do a poor job of resolving the equatorial upwelling and the tropical thermocline. They therefore risk misrepresenting these processes although, as it will turn out, it appears it is in their response to wind variations, not heating or cooling, that they slip up.

First, we provide a brief review of the experiments of CSCZ. Next, we introduce a simple two-box model of the interaction of the subtropical and tropical oceans that shows the conditions under which an ocean dynamical thermostat is theoretically possible. The results of the box model are used to explain the GCM results, which are then described.

## 2. Review of the ocean dynamical thermostat mechanism

Consider the response of the tropical Pacific SST to a spatially uniform heat flux forcing, which we will take as a heating. In the west Pacific, the ocean heat transport is essentially zero so the SST must warm by as much as is necessary for the the upward surface heat flux to balance the imposed downward flux. This is what we refer to as local surface heat flux equilibration. However, in the central and eastern equatorial Pacific, heat is diverged away from the equator by the mean upwelling. Here a portion of the imposed heating is balanced by cooling due to upwelling. The heat is moved poleward, some is lost to the atmosphere, and the rest is downwelled in the off-equatorial latitudes. Consequently, the equatorial SST changes by less in the east than it does in the west (Seager et al. 1988; CSCZ).

In the model of CSCZ (Zebiak and Cane 1987), the atmosphere responds to the strengthened equatorial SST gradient with enhanced convection in the west Pacific. This forces stronger surface easterlies to the east, which increase the upwelling and also cause the thermocline to shoal in the east. Both dynamical responses of the ocean cool the SST in the east Pacific and further strengthen the equatorial SST gradient. After this positive feedback has led to a new equilibrium, the SST anomaly in the eastern equatorial Pacific is actually negative. Elswehere, the SST has warmed but the basin mean change is less than would have occured through local surface flux equilibration. This justified use of the term ‘‘ocean dynamical thermostat.’’ CSCZ also found that the effectiveness of the thermostat varied seasonally. For the case of an imposed heating this resulted in a stronger seasonal cycle of SST in the east Pacific. Further, the interannual variability of the SST changes as the SST changes.

The main criticism of these results is that the heat that is diverged away from the equator and is downwelled in the subtropics is not allowed to alter the thermocline temperature. The GCM experiments described later are designed to see whether this adjustment erases the oceanic responses seen by CSCZ or just acts to damp them over the timescale of thermocline adjustment.

## 3. The two-box model of equatorial–subtropical interaction

The perturbed tropical ocean is represented by two boxes. The first box represents the near equatorial water lying above the thermocline and has perturbed temperature $T_1$. The second box represents the water in the subtropics and the thermocline water, all of which has the perturbed temperature $T_2$. The surface areas of the two boxes are given by $a_1$ and $a_2$ with $a_1 + a_2 = 1$. There is a perturbation dynamical heat flux between the boxes that is denoted by $Q_{12}$ and is positive if it cools box 1. Both boxes are subjected to a perturbation forcing $Q^*$ taken to be a heating and positive downward. The ocean will respond by increasing the upward surface heat flux. This is represented by the terms $\alpha_1 T_1$ and $\alpha_2 T_2$. Here, $\alpha_1$ and $\alpha_2$ represent the sensitivity of the surface heat flux to changes in SST. They can be derived by linearizing the bulk formulas for the surface heat flux and then differentiating with respect to the SST, or by numerical experiment (Seager et al. 1995b). Generally, $\alpha$ is positive and is dominated by the increase of latent heat flux with SST. However, this is significantly offset by a reduction of the longwave cooling of the surface as the SST rises (e.g., Seager et al. 1995b).

The heat balance for the boxes can be written

$$a_1 Q^* = a_1 \alpha_1 T_1 + Q_{12}, \tag{1}$$

$$a_2 Q^* = a_2 \alpha_2 T_2 - Q_{12}. \tag{2}$$

If the perturbation dynamical heat flux is zero, then the solution is

$$T_1 = Q^*/\alpha_1, \tag{3}$$

$$T_2 = Q^*/\alpha_2, \tag{4}$$

$$\tilde{T} = Q^*\left(\frac{a_1}{\alpha_1} + \frac{a_2}{\alpha_2}\right), \tag{5}$$

where $\tilde{T}$ is the basin mean SST. If there is a perturbation dynamical heat flux then the solution is

$$T_1 = \frac{Q^* - Q_{12}/a_1}{\alpha_1}, \tag{6}$$

$$T_2 = \frac{Q^* + Q_{12}/a_2}{\alpha_2}, \tag{7}$$

$$\overline{T} = Q^*\left(\frac{a_1}{\alpha_1} + \frac{a_2}{\alpha_2}\right) + Q_{12}\left(\frac{1}{\alpha_2} - \frac{1}{\alpha_1}\right). \tag{8}$$

For ocean heat transport to lower the basin mean SST, the second term in this last equation must be negative. This can occur for

$$Q_{12} > 0, \qquad \alpha_2 > \alpha_1, \tag{9}$$

$$Q_{12} < 0, \qquad \alpha_2 < \alpha_1. \tag{10}$$

This quite simply requires that the ocean transport heat from the part of the ocean where the surface heat flux is less sensitive to changes in SST to an area where the sensitivity is greater. That way the imposed forcing can be balanced by a smaller change in SST than would be the case if the forcing had to be balanced in each box individually. This result also shows that for $\alpha_2 = \alpha_1$ no amount of ocean heat transport can reduce the basin mean SST below the value derived from surface flux equilibration alone. A thermostat relies critically on variable $\alpha$. CSCZ used a fixed $\alpha$ in their experiments. Had the thermocline temperature been able to respond to the SST changes they would not have found any reduction of SST by the dynamics, only a delay while the thermocline adjusted.

We can extend this simple box model by noting that the heat transport is given by

$$Q_{12} = \nu(T_1 - T_2). \tag{11}$$

Here, $\nu = \rho c_p V$ where $\rho$ is the density, $c_p$ the specific heat of sea water, and $V$ the velocity between the two boxes accounting for meridional and gyre exchanges. Here, $\nu$ has units of W m$^{-2}$ K$^{-1}$. Substituting this expression for $Q_{12}$ we derive the solution

$$T_1 = \epsilon^{-1} Q^*(\alpha_2 + \beta), \tag{12}$$

$$T_2 = \epsilon^{-1} Q^*(\alpha_1 + \beta), \tag{13}$$

$$\overline{T} = \epsilon^{-1} Q^*(a_1\alpha_2 + a_2\alpha_1 + \beta), \tag{14}$$

$$\epsilon = \alpha_1\alpha_2 + \beta(a_1\alpha_1 + a_2\alpha_2), \tag{15}$$

$$\beta = \frac{\nu}{a_1 a_2}. \tag{16}$$

Since $a_1 + a_2 = 1$, for $\alpha_1 = \alpha_2 = \alpha$, we again see that $\overline{T} = Q^*/\alpha$ and also that $T_1 = T_2$. Hence, if $\alpha$ is fixed, not only can dynamics not affect the basin mean SST, they also cannot affect the SST of the individual boxes. This is explained quite easily. Imagine $T_1$ is greater than $T_2$, then $\nu(T_1 - T_2)$ is positive and the ocean removes heat from box 1 and puts it in box 2. That means that the forcing $Q^*$ in box 1 is partly compensated for by dynamical heat flux, which requires that $\alpha_1 T_1 < Q^*$. But box 2 is receiving heat from box 1, which requires that $\alpha_2 T_2 > Q^*$. If $\alpha_1 = \alpha_2$, this requires $T_2 > T_1$, which is inconsistent with our original assumption. The only solution for fixed $\alpha$ is $T_1 = T_2$.

For a thermostat to work we require that the basin mean SST ($\overline{T}$) be less than that derived through local heat flux equilibration [$\tilde{T}$, Eq. (5)]; namely,

$$\frac{\overline{T}}{\tilde{T}} = \frac{a_1\alpha_2 + a_2\alpha_1 + \beta}{(a_1\alpha_2 + a_2\alpha_1)\left[1 + \beta\left(\dfrac{a_1}{\alpha_2} + \dfrac{a_2}{\alpha_1}\right)\right]} < 1. \tag{17}$$

Some algebra demonstrates that this quantity is always less than 1 for $\alpha_1 \neq \alpha_2$ and for $\nu \neq 0$. Figure 1 shows plots of this quantity as a function of the model parameters. In Fig. 1a we show $\overline{T}/\tilde{T}$ as a function of $\nu$ and $P = \alpha_2/\alpha_1$ for $a_1 = a_2 = .5$. First we see that, beyond a certain value, increases in $\nu$ are unable to affect $\overline{T}$. This limit is reached when the heat transport is strong enough that $T_1 = T_2$. On the other hand, for any nonzero heat transport, increases in $\alpha_2$ will always further depress $\overline{T}$. Figure 1b shows $\overline{T}/\tilde{T}$ as a function of $\alpha_1$ and $P$ for $\nu = 8$ W m$^{-2}$ K$^{-1}$. As $a_1$ goes to either 0 or 1, the SST approaches the local surface flux equilibrium value because the system approaches a one-box limit. The thermostat is most efficient for equal size boxes.

For the physical range of interest, $\alpha_2$ is about 2–3 times larger than $\alpha_1$ (Seager et al. 1995b), $a_1 \sim a_2$ and $\nu \sim 12$, suggesting that $\overline{T}/\tilde{T}$ is about 0.8. While this may seem quite small, for a forcing that induces a 2.5 K SST response this amounts to a 0.5 K change in the meridional SST gradient, which is not insignificant. Remember, in a dynamically coupled system any such change would be amplified, as in the experiments of CSCZ.

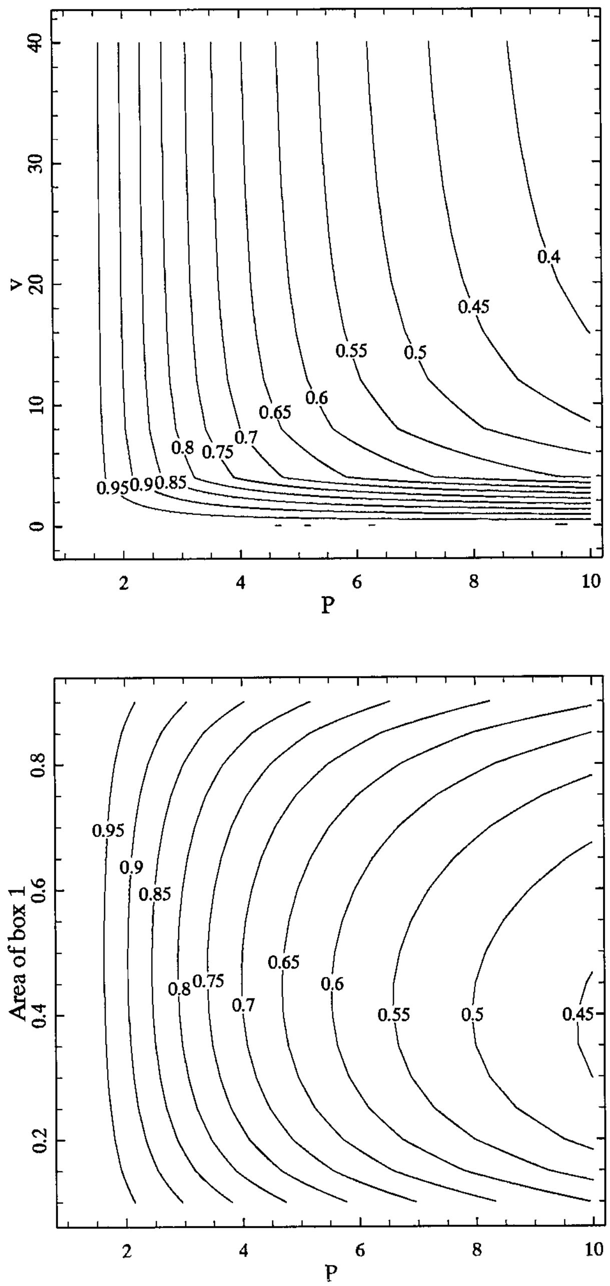


FIG. 1. Solutions of the two-box model. Ratio of the basin mean SST with interactive heat transport to the basin mean SST with fixed heat transport for various model parameters. (a) The ratio as a function of the mass exchange between the boxes ($v$, in units of W m$^{-2}$ K$^{-1}$ and of the ratio of the sensitivity of the surface heat flux to SST variations ($P = \alpha_2/\alpha_1$) for $a_1 = a_2$, and (b) the ratio as a function of the area of box 1 ($a_1$) and of $P = \alpha_2/\alpha_1$ for $v = 8$ W m$^{-2}$ K$^{-1}$.

## 4. Ocean GCM experiments

### *a. Model description*

The ocean GCM used is that described by Murtugudde et al. (1996). Its design rests on that of Gent and Cane (1989). The model domain covers the tropical and subtropical Pacific Ocean extending to 40°N and 40°S. The model has 11 layers, using a sigma coordinate, plus an ocean mixed layer. The mixed-layer depth and temperature are computed using the hybrid vertical mixing scheme of Chen et al. (1994). The model resolution is stretched from 0.3° lat at the equator to 1° poleward of 20°. The longitudinal grid is 0.3° at the boundaries and 1° in the interior. Diffusion is provided by application of a fourth-order Shapiro filter applied every 8 h. The model is forced with stresses derived from the Florida State University analysis (Goldenberg and O'Brien 1981). The solar radiation is the all-sky satellite product from the Earth Radiation Budget Experiment (Li and Leighton 1993). Other components of the surface heat flux are computed by the model. The cloud cover needed in the longwave formula is taken from the International Satellite Cloud Climatology Product (Rossow and Schiffer 1991). Poleward of 30°N and 30°S, the model SST is relaxed to observed values (Levitus 1982).

The GCM is coupled to the advective atmospheric mixed layer (AML) model of Seager et al. (1995a). The AML model solves for the air temperature and humidity by balancing advection, radiation, the surface fluxes, and entrainment at the top of the AML. Fluxes are computed from the SST and the modeled AML temperature and humidity. Only the winds, solar radiation, and cloud cover need to be externally specified. The AML model was designed to enable the computation of surface heat fluxes in the absence of specification of quantities (such as air humidity) over which the ocean has direct control. Simulations of tropical SST with the coupled GCM–AML model (but a more limited meridional domain) have been described by Murtugudde et al. (1996). They demonstrate that the model is quite successful at simulating the SST in all three tropical oceans.

### *b. Model experiments*

First, we integrate a control run for 35 yr subject to climatological, seasonally varying forcing, by which time the model is close to equilibrium. We then add a uniform 10 W m$^{-2}$ surface flux forcing and continue the integration for 70 more years, by which time the model has equilibrated to the new forcing field. All other forcing fields remained the same as in the control experiment. (To check that the control run was in equilibrium, we extended its integration for a total of 90 yr. At depth, the temperature changes between year 35 and year 90 were less than 0.2 K, indicating not only that the model was in equlibrium after 35 yr, but also that this weakly diffusive model is able to maintain a thermocline. The SST drift was less than 0.06 K.) In Fig. 2 we show the difference between the model SST 70 yr after the perturbation heat flux was added and the SST from the end (year 35) of the control run. The SST has increased everywhere but not in a uniform way. The maximum change in SST is in the equatorial west Pacific. The SST has changed by less in the eastern equa-

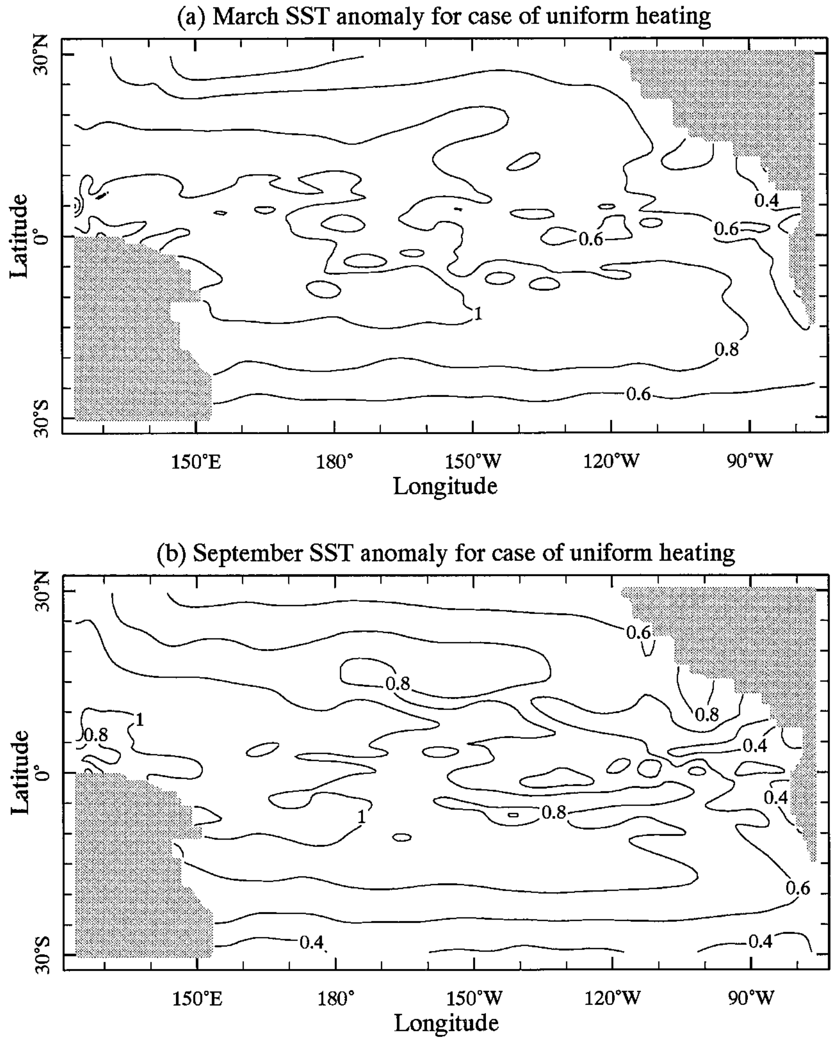


FIG. 2. Difference in SST between year 70 of the case with a uniform additional 10 W m$^{-2}$ imposed heating and the control run climatology for (a) March and (b) September.

torial Pacific and in the off equatorial regions. Poleward of 30°N and 30°S, this is explained by the relaxation, but equatorward this is not the case. Figure 3 shows a Hovmoller diagram of the SST anomaly along the equator. The east–west gradient has strengthened by 0.6°C. The maximum increase in the gradient occurs in September and October, which roughly agrees with the seasonal dependence found by CSCZ and constitutes a strengthening of the seasonal cycle.

To illustrate how the new climatology is established, we show a sequence of plots of the temperature along the equator as a function of depth (Fig. 4). Two years after the addition of the uniform heating, the SST has warmed by 0.6°C in the west and by 0.4°–0.6°C in the east. The temperature perturbation penetrates to greater depth in the west than the east. In the west the net surface heat flux is close to zero and the SST warms so that its heat loss to the atmosphere increases by as much as the imposed forcing. In the east, however, the ocean is cooled by upwelling of cold water from below. As the SST warms in response to the imposed forcing, the vertical temperature gradient increases so that the upwelling cooling increases and partially compensates for the imposed heating. Hence the SST needs to change by less here than in the west. In other words, in the east some of the added heat is diverged away from the equator and deposited at higher latitudes.

Some of the heat diverged into the subtropics will be lost to the atmosphere by surface fluxes but some will be subducted. Lu and McCreary (1995) and Liu et al. (1994) have provided cogent demonstrations of the pathways by which water moves from the subtropics into the equatorial thermocline. While water subducted in the western part of the subtropical gyre remains within the gyre, water subducted in the eastern third of the basin moves equatorward and westward at depth, equatorward in the western boundary current and then east in the Equatorial Undercurrent (EUC). Approximately

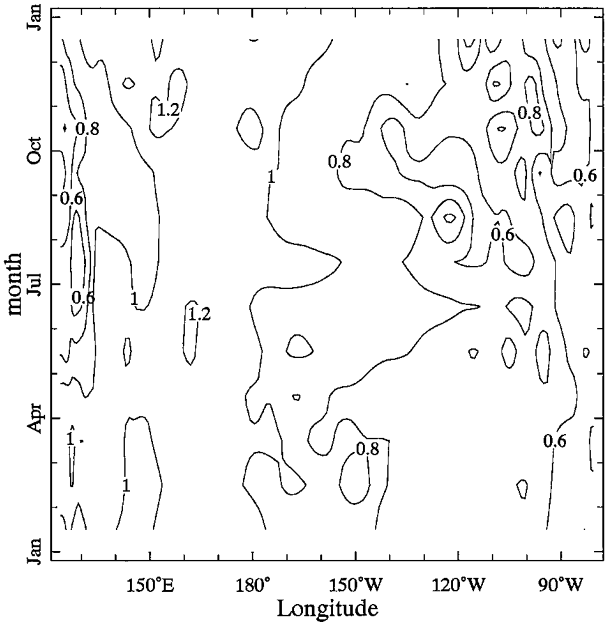


FIG. 3. The seasonal cycle of the difference in equatorial SST between year 70 of the case with a uniform additional 10 W m$^{-2}$ imposed heating and the control run climatology.

two-thirds of the EUC transport is contributed in this way with the remainder arising from recirculation in the ocean interior and from a shallow equatorial cell (Liu et al. 1994).

These results imply that some of the heat subducted in the subtropics will move back at depth to the equator and modify the thermocline temperature. This is not allowed for in the model of CSCZ. It is possible that the thermocline temperature will warm such that the vertical temperature gradient returns to its value before the perturbation was added. If that happened, then the SST in the equatorial east Pacific would eventually warm by as much it does in the west.

The sequence of longitude-depth sections of the temperature perturbation in Fig. 4 shows this is not the case. Instead, even after 70 yr, the vertical temperature gradient in the east is stronger than in the control run. Seager et al. (1995b) show that, in the Tropics, the sensitivity of the surface flux to SST perturbations increases with latitude. This derives, in part, from the increase of the latent heat flux, which, in turn, follows from the latitudinal increase in the wind speed. In addition, the flux sensitivity over the east Pacific cold tongue is low because of high relative humidity and low SSTs even though winds here are quite strong. It follows that a given heat flux forcing can be balanced by a smaller SST change in the subtropics than in the near equatorial region. As illustrated in the simple two-box model, the enhanced meridional heat flux is acting to move heat to areas where the ocean can more easily lose heat to the atmosphere. The water that is subducted in the subtropics therefore has a smaller temperature perturbation than that on the equator. If it maintained this temperature as it made its way into the EUC it could only reduce, not remove, the cooling by upwelling that offsets the surface heating in the east. On its way it mixes with surrounding water, which reduces the efficiency of the subtropical–equatorial thermocline exchange and delays the thermocline adjustment. However, as in the box model, it is the variable sensitivity of the surface fluxes to SST that is essential for the dynamical mechanism of SST regulation to work.

Do the subtropical–equatorial pathways in our model conform to those outlined by Liu et al. (1994) and Lu and McCreary (1995)? In Fig. 5a we show a longitude and depth plot of the zonal and vertical velocities on the equator for July of the final year of the simulation. Although the flow varies seasonally, for brevity we show just one month. The water upwelled on the equator is derived from the EUC and ultimately from near the western boundary. In Fig. 5b we show a map of the zonal and meridional flow at 100-m depth, which is close to the core of the model EUC in the west. This shows how weak westward flow in the subtropics feeds into strong western boundary currents that supply the EUC. In addition, there is a convergence of water from south of the equator into the EUC in the ocean interior, which has no counterpart in the north. All these features agree with the results of Liu et al. and Lu and McCreary. The latter authors ascribe the absence of an interior route in the north to the creation of a potential vorticity barrier by the intertropical convergence zone. In Fig 5c, we show a map of the vertical velocity at 50-m depth. There is downward motion in the subtropics (and with maximum values well equatorward of the relaxation zones that extend poleward of 30°). Although it is hard to tell how the destination of downwelling water varies with longitude, these results are consistent with the argument that the source of water upwelled on the equator is subtropical water that subducts, moves west to the boundary, equatorward in the boundary current, and then into the EUC.

In equilibrium, the surface layer of the ocean model must be losing an amount of energy equivalent to the applied forcing. We computed the anomalous heat flux across 30°N and 30°S, the anomalous surface heat flux, and the anomalous heat flux into the lowest, dynamically inactive model layer. The 10 W m$^{-2}$ forcing is taken up by an increase in the upward surface flux by 7.8 W m$^{-2}$ and a 2 W m$^{-2}$ flux into the lowest, dynamically inactive model layer. The latter occurs by vertical diffusion. The change in the meridional heat flux into higher latitudes is very small. The SST change, averaged over 30°N to 30°S, is about 0.8 K.

The results are not linear relative to the sign of the forcing. Figure 6 shows a vertical section of the equatorial temperature change for the case of a uniform 10 W m$^{-2}$ cooling. The change in SST gradient is less than

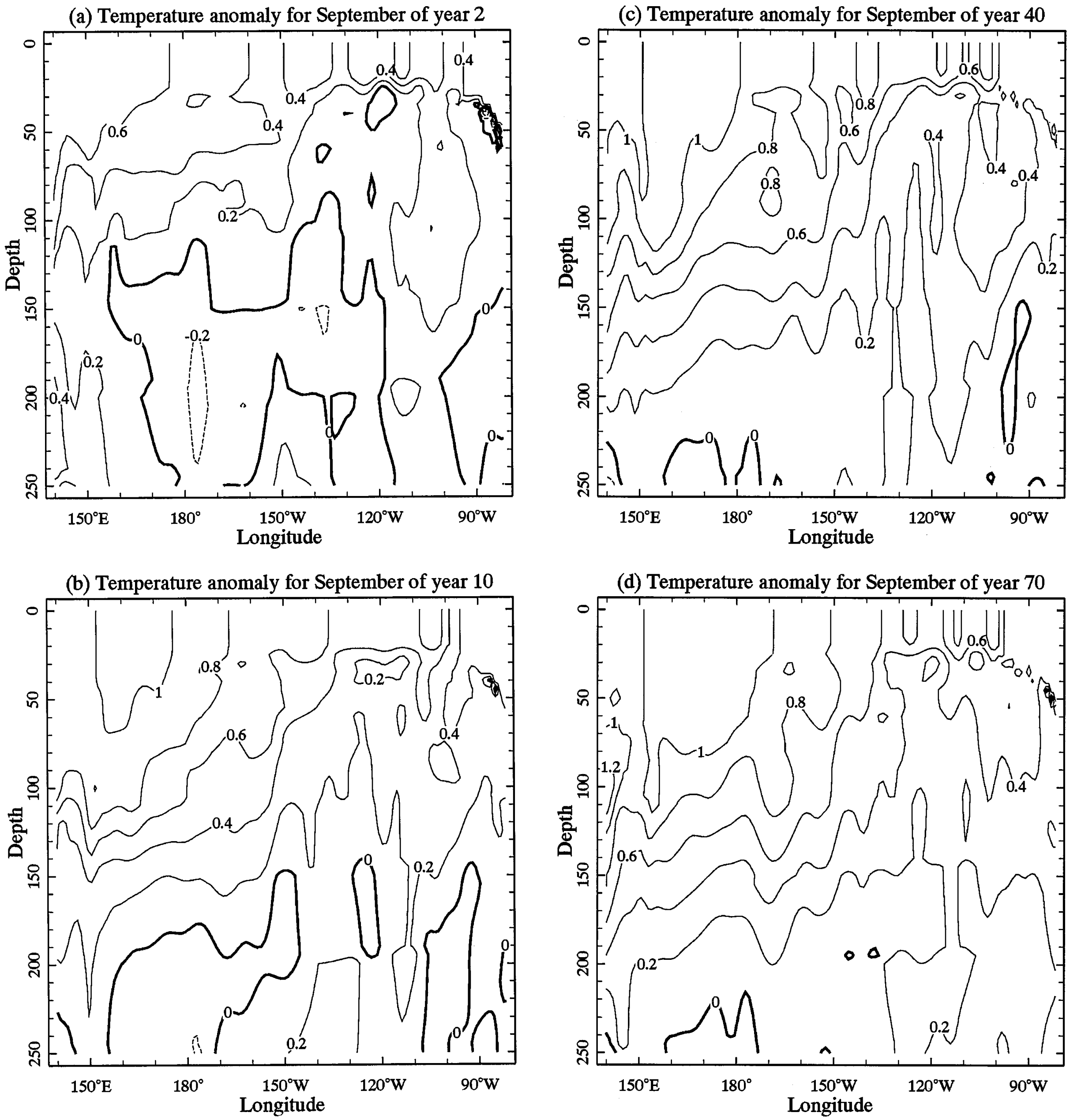


FIG. 4. Difference in equatorial temperature as a function of depth between the case with a uniform additional 10 W m$^{-2}$ imposed heating and the control run climatology for September of (a) year 2, (b) year 10, (c) year 40, and (d) year 70.

for the heating case because of a larger change in the east Pacific SST in the cooling experiment. Comparison of Fig. 6 with Fig. 4d shows that the maximum temperature change penetrates to greater depth in the west Pacific in the cooling case. Some of this water moves east near the top of the thermocline and upwells, which causes a greater change in the temperature of the upwelled water in this experiment than in the warming case. In the cooling case, the vertical stability is reduced, whereas in the warming case it is increased. For cooling, the SST changes are more readily communicated to depth, the thermocline adjustment is greater, and the thermostat mechanism is slightly weaker. The basin mean SST change was $-0.9$ K. In the remainder of the paper we restrict ourselves to consideration of the case with an imposed heating.

## 5. The SST response for fixed ocean heat transport

Currently, the complete ocean heat transport is not stored as part of the model's output. Hence it cannot be

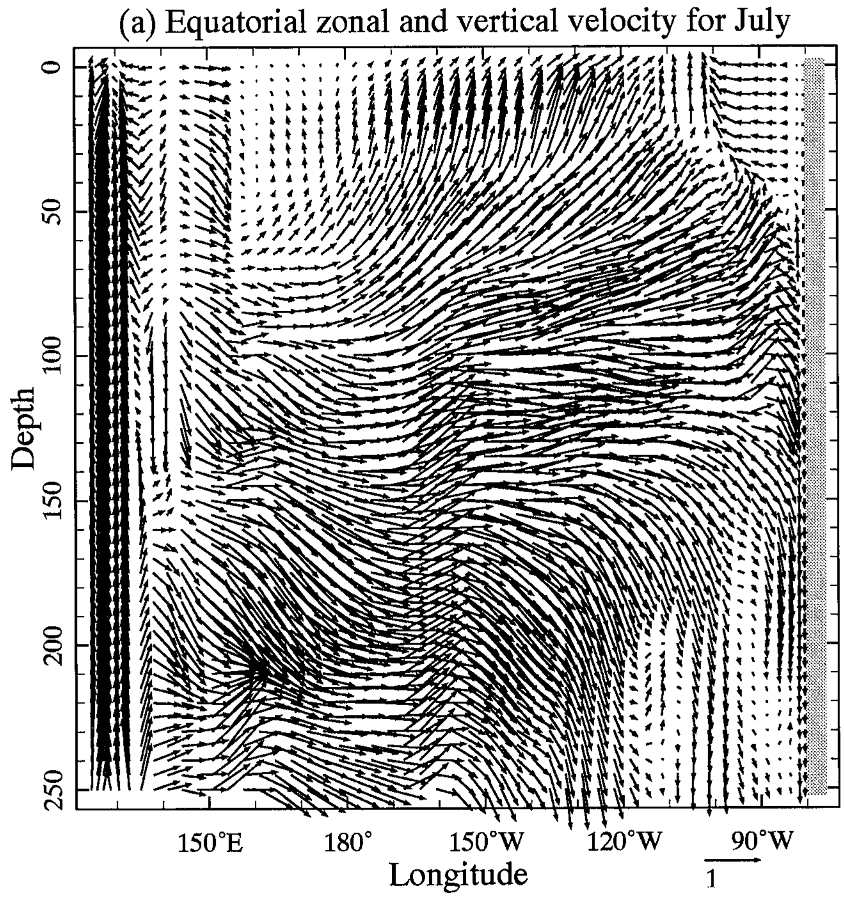


FIG. 5. (a) Zonal and vertical velocities on the equator, (b) zonal and meridional velocities at 100-m depth, and (c) the vertical velocity at 50-m depth for July of year 70. In (a) the vertical velocities have been scaled by $5 \times 10^4$ relative to the zonal velocities. In (a) and (b) the arrow at the lower right represents 1 m s$^{-1}$, and in (c) the vertical velocity in m s$^{-1}$ has been multiplied by $10^6$.

reinserted in the model to assess how the SST would have changed in the presence of the forcing if these transports had remained fixed. Instead we address this problem using the AML model coupled to a 50-m fixed depth ocean mixed layer. The SST equation for the GCM control run can be written as

$$\overline{T}_t + \overline{D} = \overline{Q}(\overline{T}), \tag{18}$$

where $D$ represents the combined effect of dynamical heat transports and diffusion, and $\overline{Q}(\overline{T})$ is the surface heat flux. We want to know the SST given by the equation

$$T_t + \overline{D} = Q(T) + Q^*, \tag{19}$$

where $Q^*$ is the forcing. Subtracting, we derive an equation for the SST perturbation $T' = T - \overline{T}$ relative to the control run

$$T_t' = Q(T) - \overline{Q}(\overline{T}) + Q^*. \tag{20}$$

The coupled AML–ocean mixed layer model was initialized with the GCM control run SST ($\overline{T}$) and integrated forward for 4 years. Here, $Q(T)$ is the surface heat flux computed by the AML model forced with the total SST, $T = T' + \overline{T}$. The forcing $Q^*$ was 7.8 W m$^{-2}$, which corresponds to the forcing imposed on the GCM less the diffusive heat flux into the lowest model layer.

Figure 7 shows the SST perturbation computed in this way. It has a maximum on the equator and decreases with latitude. Near the far northern and southern boundaries the decrease occurs because the forcing was not applied there, but within the deep Tropics and subtropics it occurs because the sensitivity of the surface heat flux to changes in SST increases with latitude as previously explained. In contrast to the GCM experiments, there is a broad maximum of SST change on the equator. Near the coasts, the SST change is small because the air temperatures and humidity over land remained fixed as the forcing was applied. This advective effect is limited to areas very close to the shore. Further offshore, the equatorial east Pacific SST change is limited by advection of drier and cooler air from the southeast.

The difference in the eastern equatorial Pacific be-

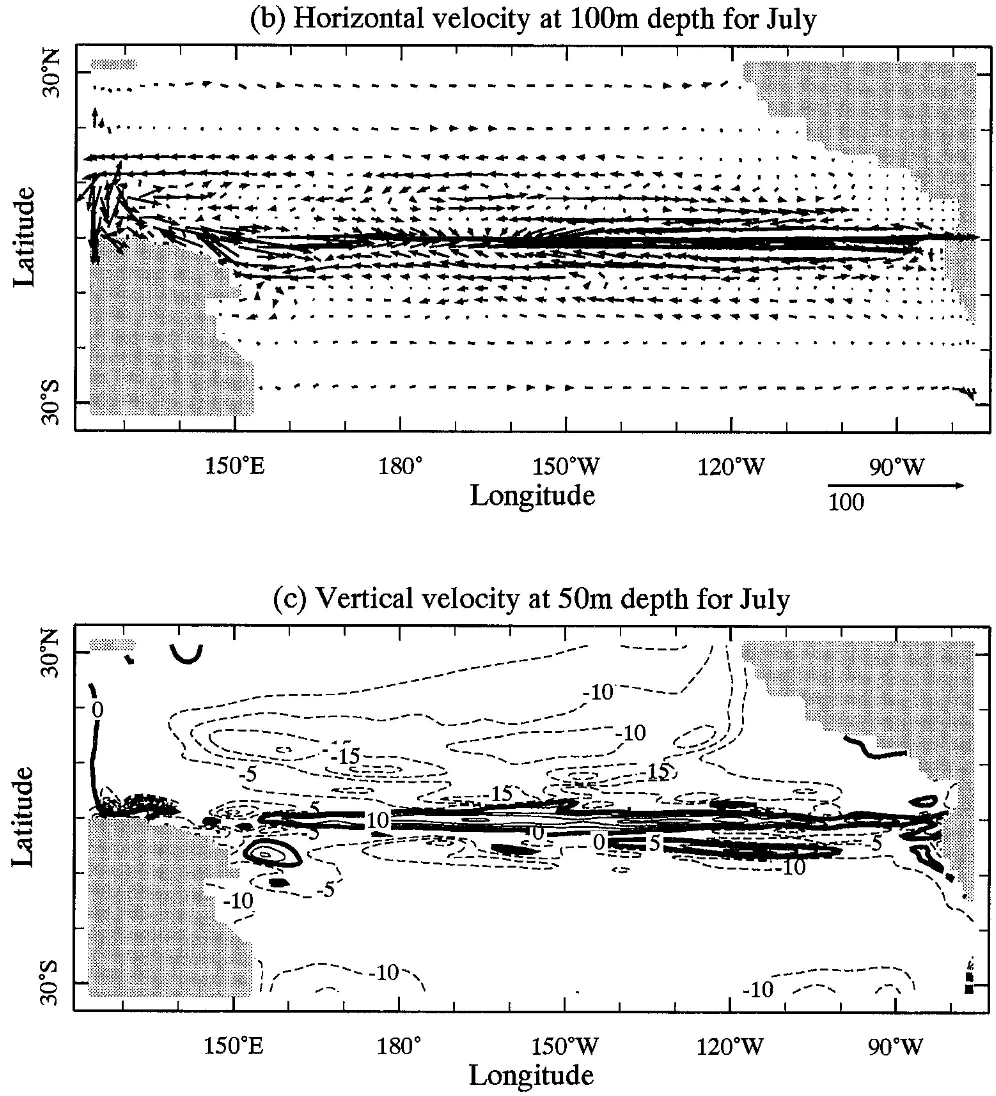


FIG. 5. *(Continued)*

tween this simulation and the case with the GCM is particularly apparent in September. During this time, cooling by equatorial upwelling is strong and causes a minimum SST change in the GCM. No minimum is seen in the AML–OML model, where the dynamical heat fluxes are held constant. The basin mean SST change in this experiment is about 1.2°C—one and a half times that in the GCM. As suggested by the two-box model, the ocean's dynamical response can act to lower the tropical SST change relative to the change induced by a purely thermodynamic response. The reduction is actually greater than the box model suggests. The box model assumes subtropical SSTs are translated to the equatorial thermocline without alteration. In reality, and in our GCM, the vertical heat transfer in the ocean depends on surface mixing, subduction, and horizontal mixing by transients and the subtropical signal becomes diluted. The damping of the SST regulation mechanism is less efficient in the GCM than in the box model.

## 6. Implications for studies of climate change and variability

Coupled GCM simulations of the response to increased $CO_2$ show a decrease in the strength of the east–west SST gradient in the tropical Pacific (e.g., Tett 1995; Knutson and Manabe 1995, KM hereafter). The increased $CO_2$ is felt at the surface in terms of a warmer atmosphere and reduced longwave cooling of the ocean surface, primarily associated with increased water vapor. Globally integrated, the ocean must increase its upward heat flux to balance.

Knutson and Manabe (1995) claim that adjustment to

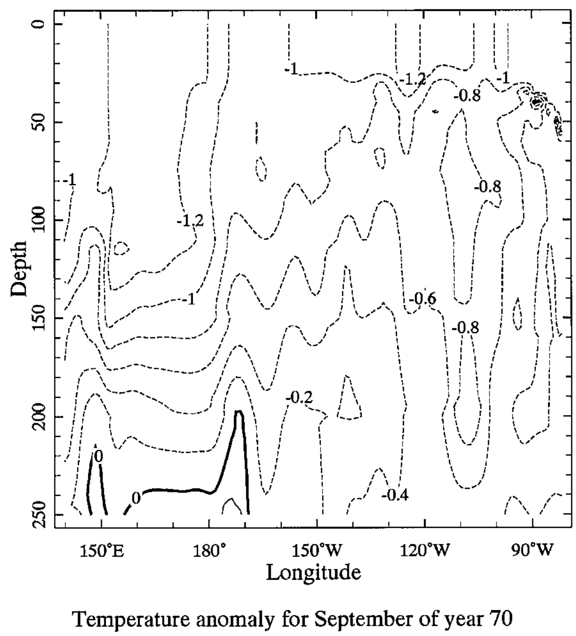


FIG. 6. Difference in equatorial temperature as a function of depth between September of year 70 of the case with a uniform 10 W m$^{-2}$ imposed cooling and the control run climatology.

the increased downward heat flux requires that the east–west SST gradient decrease. Because of the nonlinearity of the Clausius–Clapeyron equation, the latent heat flux rises less rapidly with SST in the colder east than in the warmer west. A uniform forcing requires the east to warm by more than the west to restore balance. This mechanism is included in the flux calculation of our AML model but is clearly overwhelmed by the tendency of dynamics to cool the east. It also is not evident in the coupled AML–OML integrations. The reduction in the SST gradient discussed by KM is weak, only 0.5°C on a warming of 4.5–5°C. It could easily be overwhelmed by other processes. For example, in our case it is partly offset by a tendency for the response to a heating to give enhanced downward longwave flux in the west relative to the east (Seager et al. 1995b).

Why do the results of our GCM apparently contradict those of Knutson and Manabe (1995)? There are a few possibilities.

- *A uniform heating perturbation is irrelevent to study of the effects of increasing greenhouse gases.* We assessed this using the radiative transfer schemes for water vapor and $CO_2$ of Ramanathan and Downey (1986) and Kiehl and Briegleb (1991) and the European Centre for Medium-Range Weather Forecasts analysis to represent the unperturbed atmospheric state. We computed the changes in downward longwave flux and tropospheric flux convergence forced by a doubling of $CO_2$. The change in surface flux increases toward the poles because the 15-$\mu$m $CO_2$ band system overlaps the pure rotation line and continuum absorptions of water vapor, which are increasingly saturated at high water vapor amounts. The flux convergence, which would force a change in the atmospheric temperatures that would then impact the ocean, is dominated by the change in radiation to space, which is largest in the Tropics because of the temperature dependence of the 15-$\mu$m bandwidth. Within the Tropics, both flux changes are quite spatially uniform. We also computed the increase in downward longwave radiation forced by a uniform atmospheric warming and fixed relative humidity, similar to the response shown by KM. Again it is quite uniform in the Tropics. These experiments suggest an imposed uniform heating is relevant to both the initial $CO_2$ perturbation and the equilibrated state.
- *The ocean response to increased downward heat flux depends on the ocean model resolution.* The perturbation SST equation for the GCMs (ignoring nonlinear terms) is given by

$$T'_t + \bar{u}T'_x + u'\bar{T}_x + \bar{v}T'_y + v'\bar{T}_y + \bar{w}T'_z + w'\bar{T}_z = \bar{Q} + Q', \quad (21)$$

where barred quantities denote the mean and primed quantities denote the perturbations. Here, $Q'$ is the perturbation heat flux, which in our case includes the imposed forcing. In the eastern equatorial Pacific, the vertical upwelling terms dominate the horizontal advections. Further, in the case of our experiments, the winds retain the current annual cycle so $w' \sim 0$ (there are small contributions introduced by changes in the models' temperature). Therefore,

$$T'_t + \bar{w}T'_z \sim \bar{Q} + Q', \quad (22)$$

Knutson and Manabe (1995) use a much coarser resolution model than ours. Their resolution is 4° of latitude and 6 layers in the upper 1000 m, whereas ours is 1/3° on the equator and 12 layers. Consequently, our model is better able to model the equatorial upwelling.

We repeated our experiment using lower horizontal resolution but the same vertical resolution and then again using both lower horizontal and lower vertical resolution. In Fig. 8, we show the changes in SST along the equator for these experiments. There is no compelling evidence that the change in SST gradient is sensitive to the model resolution. In each case, the amount of water upwelled is about the same although it is spread over a larger range of latitude in the coarse resolution model. This could affect the upwelling advection because the subsurface temperature varies with latitude. The effect is weak. Note that in these experiments $T_z'$ is varying in response to the surface forcing and does not have any dynamically induced component.
- *Greater warming at mid and high latitudes, and its communication to deeper in the ocean by subduction and convection, will ultimately reduce the vertical sta-*

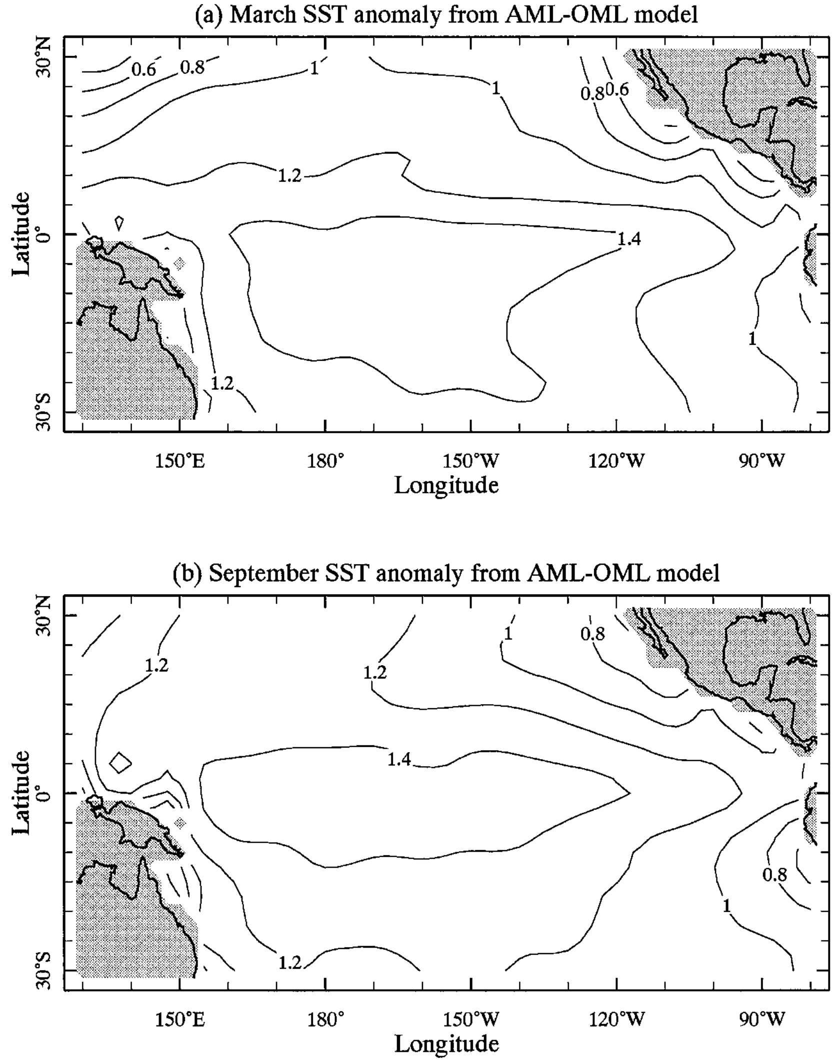


FIG. 7. Difference in SST between that computed with the coupled AML–OML model subject to a uniform 7.8 W m$^{-2}$ imposed heating perturbation and the GCM control run climatology for (a) March and (b) September.

*bility, increase the temperature of upwelled water, and reduce the equatorial SST gradient.* This mechanism is absent in our model and that of CSCZ and it is true that the reduction of the gradient in the KM model does not occur until 100 yr of the perturbed run has been completed. Furthermore, the subsurface temperatures in their model warmed by more than the surface temperatures (T. Knutson 1996, personal communication). However, in KM's model, the reduction in equatorial SST gradient occurs *before* the vertical stability reduces and is therefore not a direct result of the latter.

These considerations suggest that the initial response to increasing $CO_2$ in KM's model should be a stronger SST gradient. This will occur unless the modeled surface flux feedback they identify is stronger than ours or if the atmospheric response in their GCM (winds, clouds, etc.) acts to oppose relative warming in the west. However, we cannot abandon the issue of ocean model resolution. In high-resolution models the coupled dynamical feedback relies heavily on generation of Kelvin waves that cause shoaling of the thermocline in the east Pacific. These waves are poorly represented in coarse resolution models, which, as a result, have weak thermocline and SST variability in the eastern equatorial Pacific (Philander et al. 1992). This is consistent with the model ENSO behavior presented by Knutson et al. (1996). While this will affect the adjustment to a perturbation, it is unlikely to affect the climatological equilibrium for which a given equatorial wind stress must be closely balanced by the same pressure gradient (and therefore thermocline slope) irrespective of the model

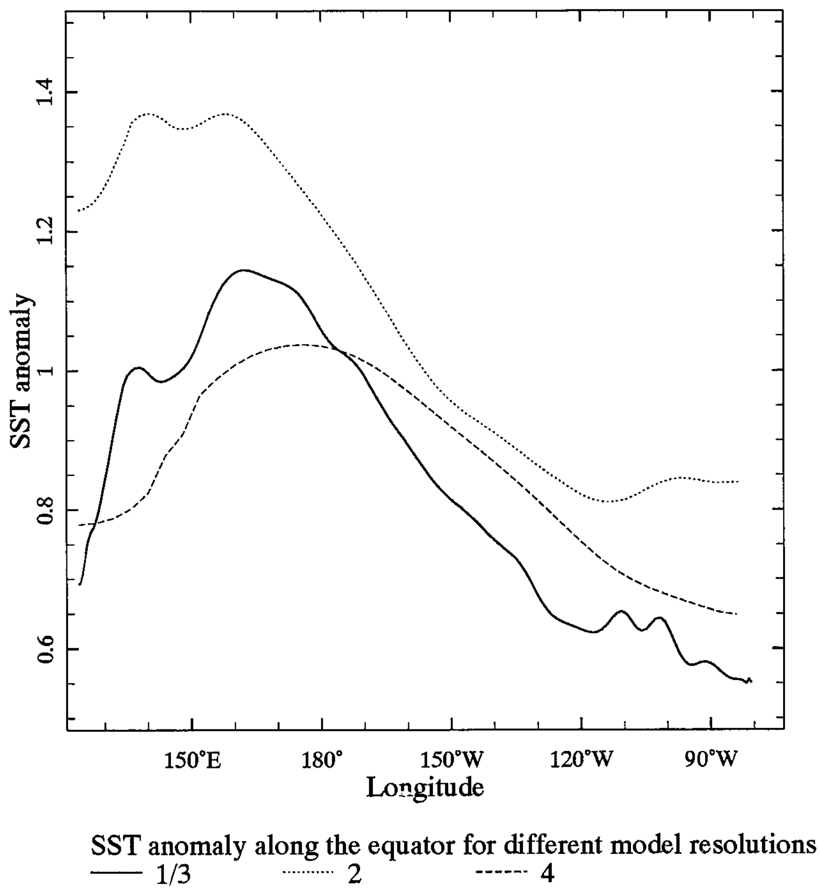


FIG. 8. The annual mean change in SST along the equator for a uniform 10 W m$^{-2}$ heating as a function of model resolution. The cases shown are for the standard model with 1/3° meridional resolution on the equator and 12 layers, a case with 2° resolution and 12 layers and one with 4° resolution and 6 layers. The results are for the 70th year after the perturbation heating was added.

resolution. Hence, if different coupled models respond to a forcing by increasing the zonal SST gradient and if the atmosphere responds by increasing the zonal wind stress, then the end result will be increased sea level and thermocline gradients despite differing ocean model resolution.

In the coupled model of CSCZ, this adjustment leads to a positive feedback because the shoaling thermocline in the east brings cooler water to the surface. If the model thermocline is too diffuse, as it is in GCMs with low vertical resolution (e.g., Lau et al. 1992), the effect of a shoaling thermocline on the temperature of upwelled water may be slight and effectively disable the positive feedback. This may allow negative feedbacks involving surface fluxes to win and lead to a weaker SST gradient. This possibility needs to be explored with fully coupled models in an attempt to explain the contradiction between our results and those of KM.

## 7. Conclusions

The basin mean change of SST due to an imposed uniform heat flux forcing is less in the presence of ocean dynamics than it is in the case of local equilibration by surface heat fluxes. This occurs because changes in vertical transfer of heat by the constant equatorial upwelling partially balance the imposed forcing. If the sensitivity of the surface heat flux to perturbations in SST is spatially uniform, then this effect will wear out as the temperature of upwelled water adjusts. This temperature is given in terms of the SST in the subtropical source regions for the equatorial thermocline water. If, however, the sensitivity of the surface fluxes to SST perturbations varies in space then it is possible that the dynamical SST regulation will persist.

We demonstrated using a simple two-box model that the necessary condition for dynamical regulation of tropical SST is that the ocean move heat from regions of low sensitivity of surface fluxes to SST perturbations to regions of higher sensitivity. In the high sensitivity regions a forcing flux can be balanced by a smaller SST change. The meridional gradient of SST then increases in response to a uniform forcing, which generates increased transport of heat from the equator to the subtropics. The subtropical SST adjusts by as much as is needed to balance the imposed forcing plus the increased dynamical heat flux convergence and is less than the SST adjustment on the equator. In time water subducted in the subtropics makes its way back into the equatorial thermocline and is upwelled but its temperature perturbation is less than that of the SST so the partial compensation of the forcing by a perturbed upwelling flux is not cancelled out.

We verified this mechanism using an ocean GCM thermodynamically coupled to a model of the atmospheric mixed layer. The sensitivity of the fluxes to changes in SST increases with latitude and is dominated by the change in the latent heat caused by increasing mean wind speed. The SST response to a uniform forcing is greatest on the equator but also has a minimum value in the eastern equatorial Pacific. Hence, the equatorial zonal SST gradient increases when a uniform heating is applied. After a few years only the temperature of the near-surface waters has changed, but after a decade the thermocline waters warm up as water subducted in the subtropics makes its way into the Equatorial Undercurrent. With time, the warming makes its way to lower depths. More than half a century is required to reach an equilibrium and, when it does, the altered SST gradient remains.

For comparison we performed an experiment with coupled atmosphere and ocean mixed layers and fixed ocean dynamical heat fluxes to assess how the SST would change by local heat flux equilibration alone. This case has no minimum SST change in the equatorial east Pacific and has a basin mean SST change 50% larger than the GCM.

Although these experiments use the simplest forcing possible, we believe they are relevant to studies of climatic change. They demonstrate the role that ocean dynamics can play in regulating both the pattern and magnitude of the tropical SST response to atmospheric forcing. In experiments with coupled GCMs subjected to increasing amounts of $CO_2$, the zonal SST gradient weakens (KM; Tett 1995), which is the opposite sign of what we would expect from our results. However, the coupled GCMs used have coarse resolution. We demonstrated that this will have little effect on the SST response to a thermodynamic forcing. However, dynamical coupling of the ocean and atmosphere will amplify the initial increase in the zonal SST gradient as in the experiments of Clement et al. (1996). The amplification depends on the sensitivity of eastern equatorial Pacific SST to changes in thermocline depth. The sensitivity is less in low vertical resolution models with diffuse thermoclines. The fact that the latent heat flux is less sensitive to SST changes in the eastern equatorial Pacific than in the west apparently wins out in the low resolution models and reduces the SST gradient.

In the literature on climate change, the ocean is considered to be of importance in two ways. First, the thermohaline circulation is considered to be potentially unstable and changes in the character of its operation are thought to have possible global climatic impacts. Second, the ocean is considered as a diffusive medium that through its large mass acts to delay the response of the climate system to imposed forcing (Hansen et al. 1981; Lindzen 1993). Here we suggest another mechanism, involving tropical ocean dynamics, whereby the ocean can influence climatic change. The mechanism operates on a timescale of decades and involves the thermal adjustment of the thermocline.

Coupled GCMs predict that, in response to increasing $CO_2$, temperatures will warm by more in mid and high latitudes than in the Tropics. This will ultimately reduce the strength of the thermocline and, through equatorial upwelling, act to reduce the equatorial SST gradient. In time, this effect could overwhelm the tropical SST regulation mechanism. Even if the SST regulation mechamism ultimately loses out to more powerful feedbacks, it is still relevant to understanding the recent history of the earth's climate and its evolution over the next few decades. If our contention is correct, then attempts to address these issues using models with fixed ocean heat transports (e.g., Mitchell et al. 1995) or low-resolution ocean models risk overlooking a decadal scale mechanism by which the tropical SST equilibrates to an external forcing. The ultimate arbiter of whether an ocean dynamic regulatory mechanism is at work must be a careful analysis of the observational record. This is under way and will be reported on shortly.

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