

## Climate and the Tropical Oceans\*

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### ABSTRACT

An attempt is made to determine the role of the ocean in establishing the mean tropical climate and its sensitivity to radiative perturbations. A simple two-box energy balance model is developed that includes ocean heat transports as an interactive component of the tropical climate system. It is found that changes in the zonal mean ocean heat transport can have a considerable effect on the mean tropical sea surface temperature (SST) through their effect on the properties of subtropical marine stratus clouds or on the water vapor greenhouse effect of the tropical atmosphere. The way that the tropical climate adjusts to changes in the ocean heat transport is primarily through the atmospheric heat transport, without changing the net top of the atmosphere radiative balance. Thus, the total amount of low-latitude poleward heat transport is invariant with respect to changes in ocean circulation in this model. These results are compared with analogous experiments with general circulation models.

Doubled CO<sub>2</sub> experiments are performed with different values of ocean heat transport. It is found that the sensitivity of the mean tropical SST to doubled CO<sub>2</sub> depends on the strength of the ocean heat transport due to feedbacks between the ocean and subtropical marine stratus clouds and the water vapor greenhouse effect. In this model, the results are the same whether the ocean heat transports are determined interactively or are fixed.

Some recent studies have suggested that an increased meridional overturning in the ocean due to changes in the zonally asymmetric circulation can reduce the sensitivity of the tropical climate to increased CO<sub>2</sub>. It is found that, in equilibrium, this is not that case, but rather an increase in ocean heat transport, which involves increased equatorial upwelling, actually warms the tropical climate.

### 1. Introduction

The regulation of tropical sea surface temperatures (SSTs) has recently been the subject of considerable debate. The processes that have been proposed to play a dominant role in determining the temperature of the warmest SSTs include a feedback between SST and evaporation (Newell 1979; Hartmann and Michelsen 1993), high cloud cover (Ramanathan and Collins 1991), the atmospheric circulation (Wallace 1992; Fu et al. 1992; Pierrehumbert 1995), the ocean circulation (Sun and Liu 1996; Clement et al. 1996; Seager and Murtugudde 1997), and stratus cloud cover (Miller 1997). Together these works show that tropical SSTs are determined by a nonlocal balance involving numerous feedbacks in the atmosphere and ocean, and that attempts to understand the factors that determine the tropical SSTs using local balance arguments are futile. We take the approach that SSTs are greatly influenced by

the general circulation of the atmosphere and ocean at low latitudes, and, in this context, we address the question of the role of the ocean.

The mean tropical SST varies on interannual timescales in association with the El Niño–Southern Oscillation (ENSO). The changes in SST are primarily due to changes in the ocean circulation and fluctuations in the equatorial thermocline depth (Philander 1990). A warmer tropical climate occurs when there is less equatorial upwelling and a weakened meridional overturning ocean circulation; a cold tropical climate occurs when equatorial upwelling and meridional ocean overturning are enhanced. The atmosphere adjusts to the change in SST through a variety of mechanisms (Sun and Trenberth 1999) that impact the subtropical source regions of the equatorial thermocline water (Liu et al. 1994; Lu and McCreary 1995). However, these interannual swings in the tropical SST happen on a shorter timescale than it takes for the ocean to adjust to the off-equatorial forcing. How would the tropical climate adjust to a persistent change in the meridional ocean overturning? Can the ENSO analogy for tropical climate change be extended to longer timescales?

Cane et al. (1997) have suggested that a La Niña-like state can persist on century timescales. They argue that coupled interactions in the tropical Pacific produce

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an enhanced Walker circulation and an increased zonal mean meridional overturning in the ocean in response to increasing CO<sub>2</sub> concentrations. This change in ocean circulation cools the tropical climate and constitutes a negative feedback to greenhouse warming at least on the timescale over which the thermocline temperature adjusts. However, the modeling experiments used by Cane et al. (1997) to make their case do not take into account the adjustment of the entire troposphere over the extent of the Hadley cell to the SST changes. The zonally asymmetric interactions involving the Walker circulation that produce the La Niña-like state can presumably be maintained, since they arise as a result of a positive feedback between the ocean and atmosphere (Bjerknes 1969). However, can this state persist when the adjustment over the Hadley cell is taken in account?

This issue of the role of the meridional ocean overturning in the tropical climate has not been fully explored using general circulation models. Often in climate change studies with atmospheric general circulation models (AGCMs), the ocean is assumed to be either a passive mixed layer (e.g., Mitchell et al. 1995) or the amount of heat that it moves is fixed (e.g., Hansen et al. 1984). This practice, while avoiding the need to run an ocean GCM, has the shortcoming that the ocean is not in dynamical balance with the atmosphere. If, for example, the Hadley circulation speeds up under climate change, it should drive a more vigorous ocean circulation that may move heat poleward. This would reduce the need for the atmosphere to transport heat poleward and thus act as a negative feedback to the more vigorous atmospheric circulation. In models that do not require dynamical consistency, this negative feedback will not operate. What are the consequences of this omission?

To address these questions within the context of a single model, we develop a two-box model of the zonal mean tropical climate over the extent of the Hadley circulation. The Walker circulation is not modeled explicitly but is taken into account as an effective zonal asymmetry of the ocean that is specified. The one-dimensional model of Betts and Ridgway (1988, 1989) is extended to include the effects of horizontal heat transports in both the ocean and atmosphere and include dynamic coupling between the atmosphere and ocean. In section 2, the model is described. In section 3, we test the role of the ocean in establishing the mean climate by varying the efficiency with which the ocean transports heat. Section 4 compares the results of the box model to similar experiments with atmospheric GCMs. In section 5 we examine the potential role of the ocean in climate change using the box model and compare the conclusions with those of previous studies (Cane et al. 1997).

## 2. Model description

To examine the effect of ocean heat transports on the tropical climate, we formulate a two-box version of the

single cell model of Betts and Ridgway (1989, hereafter BR89). The BR89 model was designed to represent a tropical mean. Here, we divide the Tropics into a warm pool and a subtropical cold pool, and allow energy to be transported between the two boxes in both the atmosphere and ocean. The model continues in the spirit of Pierrehumbert (1995), Miller (1997), and Larson and Hartmann (1999, hereafter LH99). The warm pool is assumed to be the ascending region of the Hadley cell, where convection occurs in isolated towers but descent occurs over most of the area, and the cold pool is considered to be the descending region of the Hadley cell. In this model, as in previous box models of its type, the areas of the ascending and descending regions are specified. Miller (1997) has calculated that for the current climate the warm pool occupies about 40% of the area of the Tropics. The issue of what controls the relative areas and how this may change is an important issue for tropical climate change, yet this is not the focus of the current work. We assume that they are of equal area. The energy budget over this circulation is closed after accounting for a transport to midlatitudes in the ocean and atmosphere that are specified. The model solves for subcloud and cloud layers, temperature and humidity above the convective boundary layer (CBL), radiative cooling of the troposphere, and subsidence into the CBL. The notation used here follows that of BR89.

### a. Bulk CBL equations

Figure 1 shows a schematic representation of the box model. Employing the mixing line assumption, sensible heat budgets can be written for the CBL that include transport terms between the boxes for the warm pool and cold pools:

$$\begin{aligned} \omega_{T_w}(\theta_{T_w} - \theta_{MO_w}) + \frac{g}{c_p} F_{\theta_w} \\ = \omega_{T_c}(\theta_{MO_w} - \theta_{MO_c}) + \left( \frac{g\bar{\theta}}{c_p T} \right) \Delta N_{T_w} \end{aligned} \quad (1)$$

$$\begin{aligned} \omega_{T_c}(\theta_{T_c} - \theta_{MO_c}) + \frac{g}{c_p} F_{\theta_c} \\ = \left( \frac{g\bar{\theta}}{c_p T} \right) \Delta N_{T_c} + \frac{g}{c_p} F_{\theta_{exp}}. \end{aligned} \quad (2)$$

Moisture budgets can be written as

$$\omega_{T_w}(q_{T_w} - q_{MO_w}) + \frac{g}{L} F_{q_w} = \omega_{T_c}(q_{MO_w} - q_{MO_c}) \quad (3)$$

$$\omega_{T_c}(q_{T_c} - q_{MO_c}) + \frac{g}{L} F_{q_c} = \frac{g}{L} F_{q_{exp}}. \quad (4)$$

The subscripts *w* and *c* indicate values for the warm and cold pools, respectively. The subscript *T* indicates values at the inversion. Here  $\theta_0$ ,  $q_0$  are the ocean surface

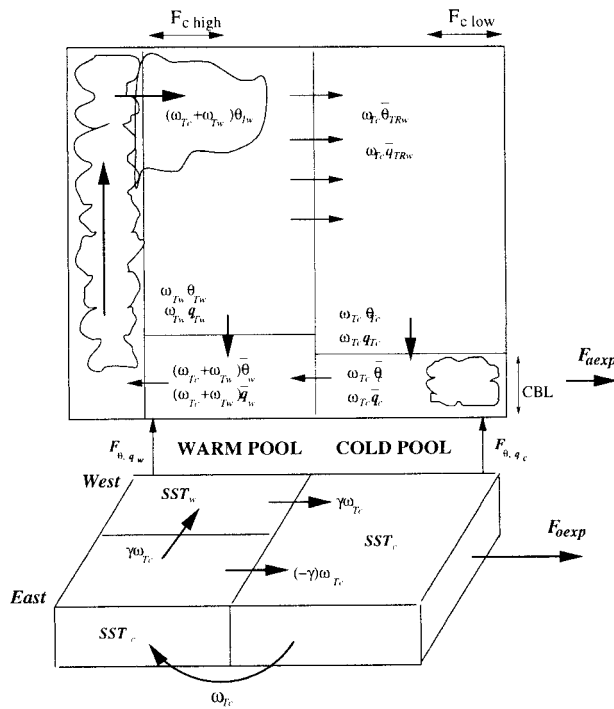


FIG. 1. Schematic representation of the box model. In the atmosphere, air ascends with the moist static energy of the subcloud layer in the warm pool,  $c_p \theta_{MOw} + Lq_{MOw}$ , and arrives at the tropopause with all of the moisture condensed out and a potential temperature of  $\theta_{lw}$ , where  $c_p \theta_{lw} = c_p \theta_{MOw} + Lq_{MOw}$ . Some air descends locally in the warm pool through the warm pool inversion,  $\omega_{r'w}$ . Some is advected into the cold pool, where it descends into the CBL and is advected back into the warm pool in the subcloud layer. The amount of air descending in the cold pool,  $\omega_{r'c}$ , is considered to be the total overturning mass flux associated with the Hadley cell. In the ocean, water is advected out of the warm pool into the subtropics. There, it is subducted and upwells onto the equator. The total overturning mass flux in the ocean is set equal to that in the atmosphere,  $\omega_{r'c}$ . A fraction  $(1 - \gamma)$  of the water is advected out of the equatorial cold pool directly into the subtropics and does not affect the heat budget of the warm pool.

temperature and saturation specific humidity;  $\theta_{MO}$ ,  $q_{MO}$  are the potential temperature and specific humidity of the surface air;  $F_\theta$ ,  $F_q$  are the sensible and latent surface heat fluxes;  $\Delta N_T$  is the radiative flux divergence across the CBL; and  $F_{\theta exp}$ ,  $F_{q exp}$  are the latent and sensible heat transports to midlatitudes that are assumed to occur only in the cold pool CBL as in Miller (1997). Values for these quantities are taken to be  $15 \text{ W m}^{-2}$  for the latent and sensible heat fluxes (after Peixoto and Oort 1993). The values of  $\omega_r$  are the modified subsidence of BR89,  $\omega'_r = \omega_r(1 - \alpha)$ , where  $\alpha = 0.25$  as in BR89, and we have dropped the prime. The subsidence referred to in the text will be this modified subsidence.

We use the bulk aerodynamic formulas to model the surface heat fluxes:

$$F_\theta = \omega_0(\theta_0 - \theta_{MO}), \quad F_q = \omega_0(q_0 - q_{MO}), \quad (5)$$

where  $\omega_0$  is the surface wind speed multiplied by  $\rho g C_D$ ,

where  $C_D$  is an exchange coefficient and equals  $0.1 \text{ Pa s}^{-1}$  unless otherwise noted.

*b. Thermodynamic structure of the CBL*

Following BR89, we assume that the thermodynamic properties of the CBL are given by a mixing line between the saturation points of the subcloud and above inversion air (Betts 1982). The subcloud layer is assumed to be well mixed below clouds and very nearly well mixed in the clear-sky portion. In the cloud layer, humidity decreases, and potential temperature increases quite steeply in the clear portion. In the cloud portion, mixing of environmental air is allowed such that the cloud liquid water content is 0.4 of its adiabatic value up to an inversion base that is specified to be 50 mb below cloud top. In the inversion, the liquid water content decreases linearly from a factor of 0.4 to 0 of its adiabatic value. This is supposed to account for the entrainment of dry air from above that reduces the liquid water content in the top of the cloud.

The cloud-top pressure is defined as the pressure at which a saturated subcloud parcel reaches a level of neutral buoyancy and is found as in BR89.

*c. Free tropospheric thermodynamic structure*

We assume that, in the warm pool, the tropospheric structure above the CBL follows a moist adiabat with the saturation equivalent potential temperature,  $\theta_{ES}$ , equal to the subcloud CBL equivalent potential temperature  $\theta_E$  ( $\theta_{MOw}$ ,  $q_{MOw}$ ). Dynamics requires that temperature gradients in the tropical free troposphere be small (Schneider 1977; Held and Hou 1980; Pierrehumbert 1995). To ensure this we require that the cold pool free troposphere has the same temperature as that of the warm pool, as was done in previous two-box models (Pierrehumbert 1995; Miller 1997; LH99).

The specific humidity immediately above the CBL,  $q_T$ , is taken to be equal to the saturation humidity on the moist adiabat at a given temperature,  $T_D$ . BR89 use  $T_D = -7^\circ\text{C}$ , but we found that in the warm pool, a value of  $T_D = 0^\circ\text{C}$  gave more reasonable values for  $q_T$  ( $6\text{--}7 \text{ g kg}^{-1}$ ). In the cold pool,  $T_D = -7^\circ\text{C}$  gives values of  $q_T$  of approximately  $4\text{--}5 \text{ g kg}^{-1}$ , which are reasonable for the subtropics (Albrecht et al. 1995). Air at the tropopause is assumed to be 30 mb below its saturation level (BR89). The specific humidity between the CBL top and the troposphere is computed assuming a linear change in the subsaturation pressure. The result is a CBL with high relative humidities (around 80%) and a rather dry free troposphere in both the warm pool and cold pool. In rough agreement, a recent compilation of VIZ sonde humidity data from the tropical west Pacific shows that between 700 and 200 mb, the peak in frequency of sampled relative humidities occurs at about 20% with a wet CBL below with relative humidities of about 80% (Spencer and Braswell 1997). In contrast,

Pierrehumbert (1995) and LH99 assumed that the entire warm pool is moistened by convection and has a constant relative humidity profile with rather large values (above 50%). Our model appears to be more in agreement with reality.

The tropopause is assumed to be at the pressure where the moist adiabat reaches 195 K. Above the tropopause the temperature is fixed at this 195 K.

#### d. Free tropospheric energy budget

To determine the subsidence into the CBL, we use the free tropospheric energy budget. We assume a linear decrease in subsidence in the warm pool free troposphere from a value of  $\omega_{T_c} + \omega_{T_w}$  at the tropopause to  $\omega_{T_w}$  at the inversion and a linear increase in subsidence in the cold pool from 0 at the tropopause to  $\omega_{T_c}$  at the inversion. This gives a constant divergence in the warm pool and a constant convergence in the cold pool with a magnitude of  $\omega_{T_c}$  integrated over the free troposphere. This gives the free tropospheric energy budget for the warm and cold pools:

$$\begin{aligned} (\omega_{T_w} + \omega_{T_c})\theta_{1w} - \omega_{T_w}\left(\theta_{T_w} + \frac{L}{c_p}q_{T_w}\right) - \omega_{T_c}\left(\bar{\theta}_{\text{TRw}} + \frac{L}{c_p}\bar{q}_{\text{TRw}}\right) \\ = \frac{g}{c_p}\Delta N_{\text{TRw}} - \left(\frac{g\bar{\theta}}{c_p T}\right)\Delta N_{T_w} \end{aligned} \quad (6)$$

$$\begin{aligned} \omega_{T_c}\left(\bar{\theta}_{\text{TRw}} + \frac{L}{c_p}\bar{q}_{\text{TRw}} - \theta_{T_c} - \frac{L}{c_p}q_{T_c}\right) \\ = \frac{g}{c_p}\Delta N_{\text{TRc}} - \frac{g}{c_p}\Delta N_{T_c}, \end{aligned} \quad (7)$$

where  $c_p\theta_{1w}$  is the moist static energy of air at the tropopause in the warm pool, and  $\Delta N_{\text{TR}}$  is the radiative flux divergence of the entire troposphere. Here  $\bar{\theta}_{\text{TRw}}$ ,  $\bar{q}_{\text{TRw}}$  are the average potential temperature and humidity over the warm pool free troposphere (from the top of the CBL,  $p_T$ , to the tropopause,  $p_1$ ):

$$\bar{\theta}_{\text{TRw}} = \frac{\int_{p_{T_w}}^{p_{1w}} \theta_w dp}{\int_{p_{T_w}}^{p_{1w}} dp} \quad (8)$$

$$\bar{q}_{\text{TRw}} = \frac{\int_{p_{T_w}}^{p_{1w}} q_w dp}{\int_{p_{T_w}}^{p_{1w}} dp}. \quad (9)$$

The sum of the terms  $(L/c_p)\bar{q}_{\text{TRw}} - (L/c_p)q_{T_c}$  in Eq. (7) is the cooling due to reevaporation of hydrometeors in the cold pool. That is, the humidity at the inversion is the result of both the process of evaporation of hydrometeors and advection of water vapor directly from the warm pool.

From the assumptions of the free tropospheric thermodynamic structure, we can define the quantity  $\theta_D$ , which is the potential temperature of air on the moist adiabat that has temperature  $T_D$  and saturation humidity  $q_T$ . Thus, we have the relation  $c_p\theta_D + Lq_T = h_D^* = c_p\theta_{1w}$  if we assume that air at the tropopause is dry. Using this relation, and combining Eqs. (6) and (7) with the CBL sensible heat budget [Eqs. (1) and (2)], we derive equations for the subsidence in each box:

$$\omega_{T_w} = \frac{g\Delta N_{\text{TRw}} - c_p\omega_0(\theta_{0w} - \theta_{\text{MOw}}) + c_p\omega_{T_c}\left(\theta_{\text{MOw}} - \theta_{\text{MOc}} - \frac{1}{c_p}\Delta h\right)}{c_p(\theta_{Dw} - \theta_{\text{MOw}})} \quad (10)$$

$$\omega_{T_c} = \frac{g\Delta N_{\text{TRc}} - c_p\omega_0(\theta_{0c} - \theta_{\text{MOc}}) + gF_{\theta\text{exp}}}{c_p\left(\theta_{Dc} - \theta_{\text{MOc}} - \frac{1}{c_p}\Delta h\right)}, \quad (11)$$

where

$$\Delta h = c_p\theta_{1w} - (c_p\bar{\theta}_{\text{TRw}} + L\bar{q}_{\text{TRw}}). \quad (12)$$

In Eq. (12),  $\Delta h$  is the difference in moist static energy between air coming into the warm pool at the tropopause (out of the top of the convective towers) and that leaving the warm pool troposphere,  $(c_p\bar{\theta}_{\text{TRw}} + L\bar{q}_{\text{TRw}})$ , which

has been lowered through evaporative and radiative cooling.

#### e. Subcloud layer energy budget

As in BR89, we use a subcloud layer energy balance to close the system. In common with previous models, we assume that the entrainment flux of potential temperature at the cloud base equals a fixed proportion,  $k$ , of the sensible heat flux (BR89; Bretherton 1993; Nicholls and LeMone 1980). Including the sensible heat transport, the sensible heat budget for the subcloud layer of each box is

$$\omega_0(1+k)(\theta_{0w} - \theta_{MOw}) = \omega_{Tc}(\theta_{MOw} - \theta_{MOc}) + \frac{g}{c_p} \Delta N_B \quad (13)$$

$$\omega_0(1+k)(\theta_{0c} - \theta_{MOc}) = \frac{g}{c_p} F_{\theta_{exp}} + \frac{g}{c_p} \Delta N_B. \quad (14)$$

Here,  $\Delta N_B$  is the radiative flux divergence across the subcloud layer. The value of  $k$  was chosen to be 0.25 as in BR89.

#### f. Ocean surface energy budget

The energy budget of the ocean surface can be written as

$$Q_s - Q_{lw} - F_\theta - F_q = F_o + F_{oexp}, \quad (15)$$

where  $Q_s$  is the incident solar radiation,  $Q_{lw}$  is the net longwave cooling,  $F_\theta$  the sensible heat flux, and  $F_q$  the evaporative heat flux. The expression  $F_{oexp}$  is the horizontal transport of heat in the ocean to the midlatitudes which is zero for the warm pool and  $30 \text{ W m}^{-2}$  for the cold pool (Peixoto and Oort 1993).

The term  $F_o$  is the transport associated with the meridional overturning circulation in and above the thermocline between the subtropics and Tropics. It is assumed that water with the temperature of the cold pool SST is subducted in the subtropics and upwells on the equator and, likewise, water with the SST of the warm pool is advected out of the warm pool and into the cold pool. The heat transport is then determined by the product of the overturning mass flux and the SST difference. The surface stress exerted on the ocean by the atmosphere must be equal and opposite to that exerted on the atmosphere by the ocean,  $\tau_{xo} = -\tau_{xa}$ . If we assume Ekman balance and average over the atmospheric CBL and ocean mixed layer, assuming the stress goes to zero at the top and bottom, respectively, then

$$fV_a = -fV_o, \quad (16)$$

where  $V_a$  and  $V_o$  are the total mass transports of the atmosphere and ocean frictional layers, which are equal and opposite. Our definition of the overturning mass transport in the atmosphere is the subsidence rate in the cold pool, and the heat transport between the warm and cold pools in the ocean can be written as the product of  $\omega_{Tc}$  and the SST difference between the boxes. Values for  $\omega_{Tc}$  are about  $0.05 \text{ Pa s}^{-1}$ , which amounts to an overturning mass transport for both the ocean and atmosphere of about  $2.5 \times 10^{11} \text{ kg s}^{-1}$  over the area of the Hadley circulation ( $20^\circ$  lat), which is somewhat large compared with observations (Peixoto and Oort 1993) and coupled model results (Manabe et al. 1990). The radiative fluxes and static stability that the model computes and that go into determining the cold pool subsidence compare favorably with observations. One possibility for the excess strength of the model Hadley cell is that evaporative cooling is being overestimated. Per-

haps the delivery of water vapor to the cold pool free troposphere is accomplished entirely by advection from the warm pool without local evaporation (e.g., Salathé and Hartmann 1997).

In the real world, the ocean has meridional boundaries that set up a zonal pressure gradient at low latitudes. This results in a zonal tilt of the thermocline such that it is only in the eastern part of the basin that water from the subtropics upwells to the surface on the equator. In the western part of the basin, the upwelled water comes from above the thermocline and has little effect on the surface energy budget. In this configuration, subtropical water reaches the warm pool through zonal advection. However, because of the poleward Ekman drift at the equator, some of the water that is advected westward along the equator does not reach the warm pool before it is advected into the subtropics and thus does not affect the surface heat budget of the warm pool (Liu and Huang 1997). This is illustrated in Fig. 1. A fraction of the water upwelled into the cold pool on the equator,  $(1 - \gamma)$ , is advected directly to the subtropics. Thus, the heat transport between the warm and cold pools can be written as

$$F_{ow} = \gamma \omega_{Tc} \frac{c_{po}}{g} (\text{SST}_w - \text{SST}_c) = -F_{oc}. \quad (17)$$

Writing  $\gamma = 1$  would describe zonal symmetry, and  $\gamma = 0$  means that there is no ocean heat transport. In the real world, the value would be somewhere in between. Liu and Huang (1997) have argued that most of the poleward Ekman flow should occur in the western part of the basin due to westward wind drift and that  $\gamma > 0.5$ . However, all of the water that upwells onto the equator does not follow the circuit described by Fig. 1. Much of the upwelled water will undergo this kind of overturning circuit *within* the warm pool, or it goes directly from the cold tongue to the cold pool without entering the warm pool, hence reducing the efficiency of the meridional ocean heat transport. We suggest that values for  $\gamma$  should be smaller, and a value of about 0.2 yields an ocean heat export from the warm pool of about  $20 \text{ W m}^{-2}$ , with an SST difference between the warm and cold pools of about 5 K.

This configuration allows us to examine how different treatments of the ocean might affect the tropical mean climate in GCMs. The case with  $\gamma = 0$  represents a mixed layer ocean with no heat transport. The  $q$ -flux models will have a fixed ocean heat transport. Coupled GCMs will have an effective value of  $\gamma$ , and the value will depend on the longitudinal asymmetry of the model.

This aspect of the model is different from previous two-box tropical models of the Tropics where ocean heat transports were specified (Pierrehumbert 1995; Miller 1997; LH99). Here, we require that the ocean be dynamically consistent with the atmosphere, while crudely allowing for longitudinal asymmetries in the ocean.

### g. Tropical energy balance

If we add the CBL latent and sensible heat budgets [Eqs. (1) and (3)] to the free tropospheric energy budget [Eq. (6)] of the warm pool, we find that

$$\omega_{T_w}(c_p\theta_{1w} - h_{MOw}) + \omega_{T_c}(c_p\theta_{1w} - \bar{h}_{TRw} + h_{MOc} - h_{MOw}) = g\Delta N_{TRw} - gF_{\theta w} - gF_{q w}, \quad (18)$$

where  $h = c_p\theta + Lq$  is the moist static energy. We have made the assumption that air is delivered to the tropopause in the warm pool by convection with the moist static energy of the subcloud layer and that the humidity at the tropopause is zero. Thus  $c_p\theta_{1w} = h_{MOw}$ , and Eq. (18) reduces to

$$\omega_{T_c}(h_{MOc} - \bar{h}_{TRw}) = g\Delta N_{TRw} - gF_{\theta w} - gF_{q w}. \quad (19)$$

If we then add the surface energy budget, we find that energy balances for the warm pool, cold pool, and the entire Tropics are satisfied by

$$\begin{aligned} \frac{\omega_{T_c}}{g}[(\bar{h}_{TRw} - h_{MOc}) + \gamma c_{po}(SST_w - SST_c)] &= Q_{TOAw}, \\ \frac{\omega_{T_c}}{g}[(\bar{h}_{TRw} - h_{MOc}) + \gamma c_{po}(SST_w - SST_c)] &= -(Q_{TOAc} - F_{\theta exp} - F_{q exp} - F_{o exp}), \\ F_{\theta exp} + F_{q exp} + F_{o exp} &= Q_{TOAw} + Q_{TOAc}, \end{aligned} \quad (20)$$

where the terms on the right-hand side are the net top of the atmosphere (TOA) fluxes in the warm and cold pools and are defined as positive downward. That is, the net radiative gain of the warm pool is balanced by the heat export in the ocean and atmosphere, which is proportional to the SST difference between the warm pool and cold pool and the difference in moist static energy between the poleward and equatorward flows. The case with  $\gamma = 0$  is when the ocean transports no heat, and with  $\gamma = 1$ , the efficiency of ocean heat transports is a maximum.

### h. Radiation model

The longwave scheme is derived from that of Ramanathan and Downey (1986). It is a nonisothermal emissivity scheme that uses four broad bands and has been well calibrated against available line-by-line and narrowband models. Clouds are assumed to have an emissivity of 1. The solar transfer scheme is that of Lacis and Hansen (1974) but uses the more recent coefficients derived by Ramaswamy and Freidenreich (1992). Cloud reflection, absorption, and transmission are computed from the cloud liquid water and water vapor contents following Stephens (1978). The  $CO_2$  absorption scheme follows from Kiehl and Briegleb (1991).

An annual mean value for the incident solar radiation is used for each of the boxes ( $412 \text{ W m}^{-2}$  for the warm

pool, and  $387 \text{ W m}^{-2}$  for the cold pool). Annual mean values for the daily mean solar zenith angle [weighted by the incoming solar radiation as in Hartmann (1995)] of  $41.8^\circ$  and  $45.6^\circ$  were used for the warm and cold pools, respectively.

The clear- and cloudy-sky fluxes are derived separately and the mean values are obtained from weighted averages of the clear- and cloudy-sky values. In the warm pool, we divide the area into a fraction covered by clear skies,  $(1 - F_{\text{high}})$ , and a fraction covered by high clouds,  $F_{\text{high}}$ , with the fraction of high cloud specified. The high clouds are considered to be anvil clouds, which cover a much larger area than the clouds associated with the cumulus scale updrafts. We do not include the effects of warm pool low clouds on the radiative fluxes. While this is somewhat inconsistent with the assumption of a partly cloudy CBL, satellite data suggest that the effect of low clouds on the TOA balance is negligible in the warm pool (Hartmann et al. 1992). The temperature and humidity profiles for the free troposphere in the cloudy skies are the same as those for the clear skies, except that in the cloudy portion, we use the saturation humidity as the specific humidity. The liquid water content in the cloud is set to be 0.01 of the saturation humidity as in Slingo (1987). The cloud top is assumed to occur at 250 mb. We tune the cloud depth such that the shortwave and longwave cloud forcing cancel at the TOA as in satellite observations (Hartmann et al. 1992; Pierrehumbert 1995). We used a cloud base at 450 mb that gives a shortwave forcing of about  $-60 \text{ W m}^{-2}$  and about  $+60 \text{ W m}^{-2}$  for the longwave forcing for a warm pool high cloud fraction of 40%.

Low clouds are the only cloud type in the model cold pool. These have the most significant affect on the TOA balance in the subtropics (Hartmann et al. 1992; Klein and Hartmann 1993). In some experiments a cloud fraction of 30% is assumed, which is similar to observations (Hartmann et al. 1992). We will also use an empirical relation for cloud cover based on the low-level static stability from Klein and Hartmann (1993) to see how low clouds operate within the tropical climate.

## 3. The role of the ocean in establishing the mean tropical climate

### a. The case of fixed cold pool low cloud cover

The results in this section are all for the case of cold pool low cloud cover fixed at 30%. First, we ask whether heat transports in the ocean are necessary to maintain a tropical climate that is not in a state of a runaway greenhouse. In the current climate, the deep Tropics are the site of a local minimum in outgoing longwave radiation (OLR), an observation that led Pierrehumbert (1995) to postulate that atmospheric heat transports bring the Tropics into balance at realistic temperatures by advecting heat to the relatively dry cold pool “radiator fins.” However, there is also less evaporation, a flux of

energy from the ocean into the atmosphere, in the warm pool than in the cold pool. The minima in OLR and evaporation suggest that the local cooling mechanisms of *both* the atmosphere and the ocean are not particularly efficient in the warm pool, and that it is easier for the general circulation of the ocean and the atmosphere to bring the system into balance than it is to achieve a local balance. We can estimate how much the ocean cools the warm pool by averaging the net surface heat flux over the warm pool region (10°S–10°N and 140°E–180°). Values are typically about 20 W m<sup>-2</sup> (Oberhuber 1988). If the ocean did not export heat, how would the excess heat input to the warm pool be dealt with? Presumably, evaporation in the warm pool would have to increase, which could be accomplished by an increase in SST. Hallberg and Inamdar (1993), however, noted that as SST increases, longwave cooling of the surface *decreases*. Would this lead to a runaway greenhouse?

To address these questions, we perform experiments with the box model varying the efficiency of ocean heat transports ( $\gamma$ ) from 0 to 1. The model parameters are set to  $T_{Dw} = 0^\circ\text{C}$ ,  $T_{Dc} = -7^\circ\text{C}$ ,  $F_{\theta_{\text{exp}}} = 15 \text{ W m}^{-2}$ ,  $F_{q_{\text{exp}}} = 15 \text{ W m}^{-2}$ ,  $F_{o_{\text{exp}}} = 30 \text{ W m}^{-2}$ , and warm pool and cold pool wind speeds multiplied by  $\rho g C_D$  are set to 0.1 Pa s<sup>-1</sup>.

We expect that changes in the ocean heat transport will be mainly balanced by evaporation. In the absence of any feedbacks in the atmosphere, with increased ocean heat transport, the warm pool should cool and the cold pool should warm, and the temperature change should be smaller in the warm pool because of the larger change in saturation water vapor pressure at higher temperatures. Hartmann and Michelsen (1993) suggested that this large sensitivity of evaporative heat flux to SST at high temperatures would act to keep the warm pool SSTs quite stable. They make their argument based on the assumption of fixed relative humidity and an evaluation of how changes in the local SST will alter the net longwave cooling of the ocean (Hallberg and Inamdar 1993). In the model presented here, however, there are numerous feedbacks in the atmosphere operating both locally and nonlocally that affect the surface heat fluxes.

Figure 2a shows the change in the warm pool, cold pool, and mean SST. There is a small increase in cold pool SST (about 1 K) over the range of  $\gamma$ , while the warm pool SST decreases by about 3 K. The change in ocean heat transport over  $\gamma = 0$  to 1 is about 40 W m<sup>-2</sup>. This is mainly balanced by a decrease in evaporation in the warm pool and an increase in the cold pool. The evaporative heat flux is related to both the SST and a modified relative humidity (RH) and can be written as

$$F_q = \frac{L\omega_0}{g} q_s(\text{SST})(1 - \text{RH}), \quad (21)$$

where  $\text{RH} = r q_s(\theta_{\text{MO}})/q_s(\text{SST})$  and  $r$  is the actual relative humidity of the subcloud air. The sensitivity of

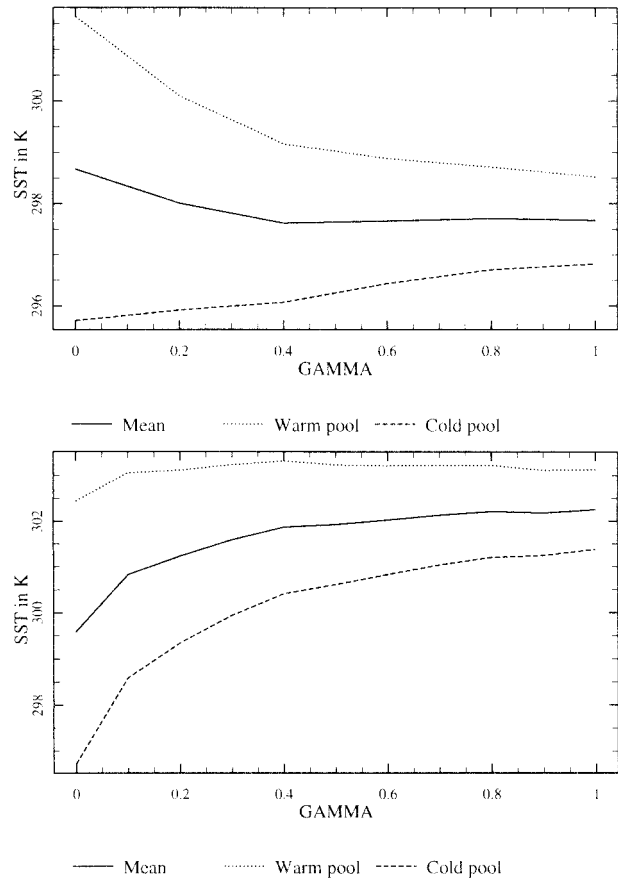


FIG. 2. Warm pool (dotted), cold pool (dashed), and mean (solid) SST as a function of the efficiency of ocean heat transports,  $\gamma$ , in the box model (in K) with fixed cold pool cloud cover and (a) varying low cloud reflectivity and (b) cloud shortwave radiative fluxes set at values for  $\gamma = 0$ .

evaporative heat flux to SST for a fixed surface relative humidity is typically about 7 W m<sup>-2</sup> K<sup>-1</sup> for SSTs around 300 K (Hartmann and Michelsen 1993) and is smaller for lower SSTs. The changes in SST in the box model are far too small to balance the 40 W m<sup>-2</sup> change in ocean heat flux subject to fixed relative humidity. Instead, the relative humidity in each box changes in such a way that the change in SST with respect to ocean heat transport is reduced.

An expression for the relative humidity can be derived from the CBL humidity budget, which can be written as an equation for the surface humidity of the warm and cold pools as

$$q_{\text{MOw}} = \frac{1}{(\omega_0 + \omega_{T_w})} (\omega_0 q_{sw} + \omega_{T_w} q_{T_w} - \omega_{T_c} (q_{\text{MOw}} - q_{\text{MOc}})), \quad (22)$$

$$q_{\text{MOc}} = \frac{1}{(\omega_0 + \omega_{T_c})} (\omega_0 q_{sc} + \omega_{T_c} q_{T_c}). \quad (23)$$

In the cold pool, as the SST difference decreases with

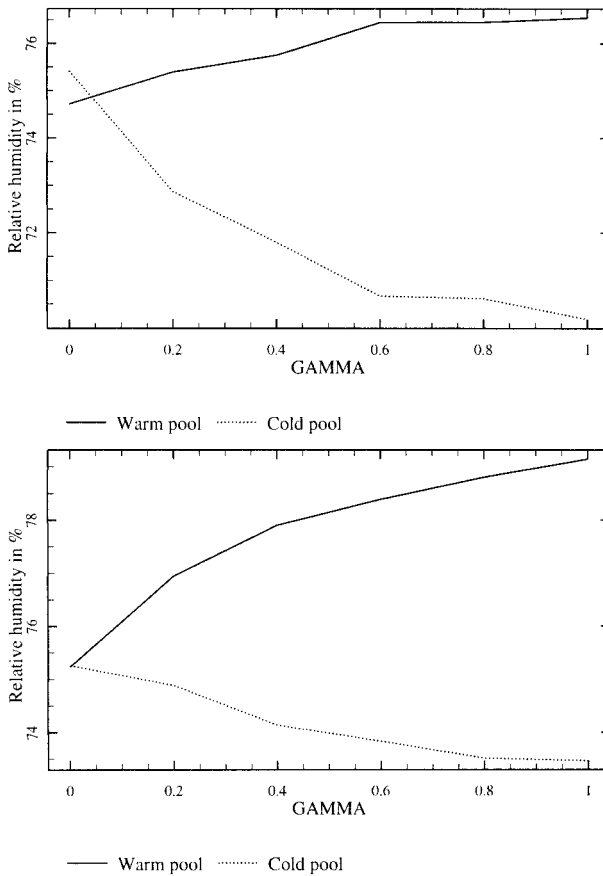


FIG. 3. Warm pool (solid) and cold pool (dotted) surface relative humidity (defined as  $q_s/q_s$ ) as a function of  $\gamma$  for fixed cloud cover and (a) varying cloud reflectivity and (b) cloud shortwave radiative fluxes set at values for  $\gamma = 0$ .

$\gamma$ , the cold pool static stability decreases since the temperature of the lower free troposphere is related to the equivalent potential temperature of the warm pool. A decrease in static stability increases the subsidence (BR89). The change in subsidence is large over  $\gamma$ , and since  $\omega_0 q_{sc} > \omega_{Tc} q_{Tc}$ , the relative humidity for the cold pool can be written approximately as

$$RH \approx \frac{\omega_0}{(\omega_0 + \omega_{Tc})}. \quad (24)$$

As the cold pool subsidence increases with  $\gamma$ , the surface relative humidity decreases (Fig. 3a), which allows the evaporative heat flux to increase with  $\gamma$  without much of a change in SST.

In the warm pool, on the other hand, the surface relative humidity increases, while the subsidence remains fairly constant over  $\gamma$  due to opposing local and nonlocal feedbacks. As the warm pool SST cools, the atmosphere moves to a drier and less steep moist adiabat. This local adjustment lowers the static stability and tends to increase the subsidence. However, a decreased horizontal SST gradient reduces the cooling of the warm pool CBL

by horizontal heat transport in the atmosphere, which tends to decrease the subsidence. The result is little change in warm pool subsidence over  $\gamma$ . Changes in relative humidity of the warm pool are, instead, controlled by the last term in Eq. (22). As  $\gamma$  increases, the advection of dry air into the warm pool decreases. Thus, the relative humidity increases and the evaporation decreases as the ocean cools the warm pool. The sensitivity of evaporative heat flux to SST in the warm pool is about  $12 \text{ W m}^{-2} \text{ K}^{-1}$ , considerably larger than typical values derived for fixed relative humidity.

The fact that the cold pool SST changes by less than the warm pool is counter to predictions based on the nonlinear increase of the saturation water vapor pressure with SST. At higher SST, as in the warm pool, less temperature change is needed to accomplish the same change in evaporation than at colder temperatures. Instead, it is processes in the atmosphere that determine the relative change in the warm pool and cold pool, and hence the mean tropical SST. As  $\gamma$  increases, the SST difference between the warm pool and cold pool decreases. This reduces the static stability of the cold pool lower troposphere, which means that the entrainment of warm air into the CBL must come from higher in the atmosphere and the CBL depth will increase (BR89). This has two effects on the radiative properties of the tropical atmosphere. On the one hand, a deeper CBL means an increase in the overall moisture in the cold pool atmosphere, which would make it more opaque to longwave radiation. On the other hand, a deeper CBL also increases the total liquid water content of the clouds and hence their reflectivity (Stephens 1978). The latter effect is stronger, and thus the increased heating of the cold pool with  $\gamma$  by the ocean heat transports is partly offset by a decrease in incident solar radiation at the surface, which restricts the warming of the cold pool.

While the feedback between cloud liquid water content and reflectivity is plausible, we cannot be sure about its magnitude. Thus, we perform the same experiments but hold the cloudy-sky shortwave radiative fluxes in the cold pool fixed at the values for  $\gamma = 0$ . Figure 2b shows that when this cloud reflection feedback is eliminated, the cold pool warms significantly over  $\gamma$  (more than 4 K), the warm pool SST warms slightly (about 1 K), and the mean temperature increases by about 2.5 K. The larger temperature change in the cold pool is now consistent with what is predicted by the nonlinearity of the Clausius–Clapeyron equation. However, the warming of the warm pool is not what one would expect from the relation between evaporation and SST. Evaporation decreases in the warm pool as the ocean transports more heat, while the SST increases. The decrease in evaporation with increasing ocean heat transport is brought about entirely by an increase in relative humidity as warmer and wetter air is advected into the warm pool from the cold pool (Fig. 3b). In this case, the deepening of the CBL with  $\gamma$  has a warming effect since it increases



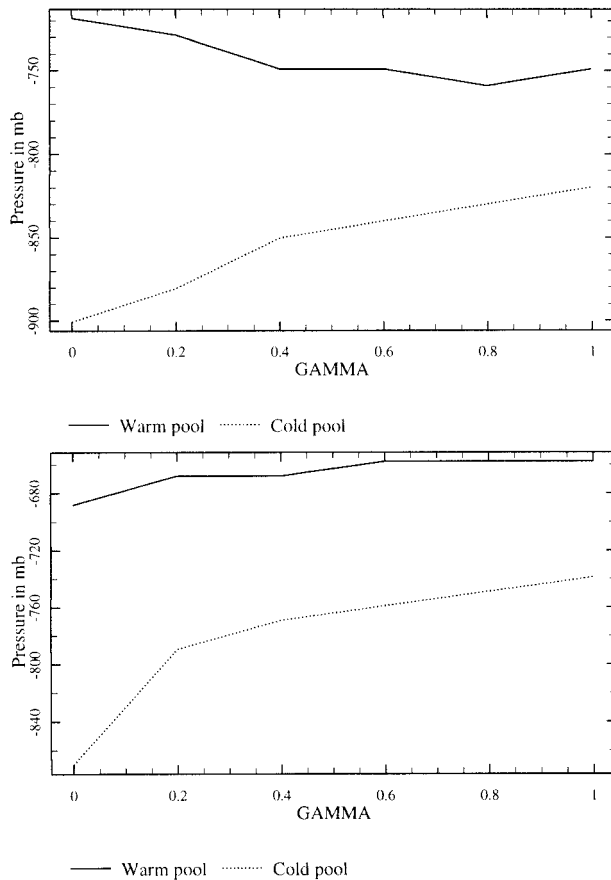


FIG. 4. Warm pool (solid) and cold pool (dotted) cloud-top pressure in mb as a function of  $\gamma$  for fixed cloud cover and (a) varying cloud reflectivity and (b) cloud shortwave radiative fluxes set at values for  $\gamma = 0$ .

the greenhouse capacity of the tropical atmosphere. As a result, the entire tropical climate warms and the CBL increases in both the warm and cold pools (Fig. 4b) as the ocean heat transport increases.

In these experiments, ocean heat transports are not essential in preventing a runaway greenhouse. Even in the case of no ocean heat transport ( $\gamma = 0$ ), the warm pool SST is still at reasonable values (Figs. 2a,b). As Pierrehumbert suggested, atmospheric heat transports alone can bring the system into balance. Figure 5 shows the partitioning of heat transports between the ocean and atmosphere as a function of  $\gamma$ . The total amount of poleward heat transport from the warm pool is independent of the efficiency of ocean heat transports: any change in ocean heat transports is compensated by an opposite change in the atmospheric heat transport. The explanation for this result lies in the physics of the greenhouse effect and the assumptions made about the humidity of the free troposphere. The total amount of poleward heat transport from the warm pool equals the net radiative gain of the warm pool [Eq. (20)]. Because the net cloud forcing of the warm pool is required to

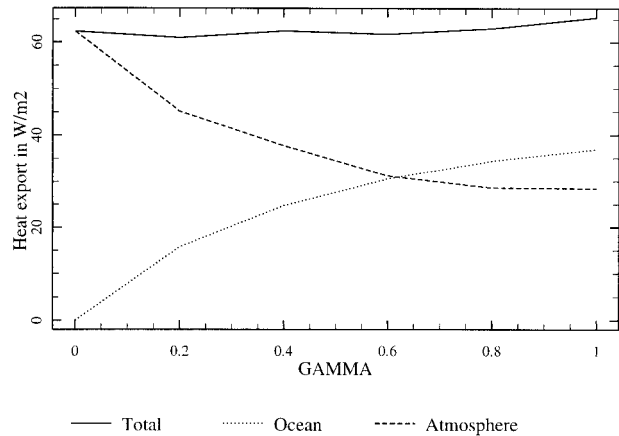


FIG. 5. Heat export from the warm pool in the ocean (dotted), atmosphere (dashed), and ocean + atmosphere (solid) as a function of the efficiency of ocean heat transports,  $\gamma$ . The total heat transport remains invariant over  $\gamma$  because of the insensitivity of warm pool radiative gain. Changes in the ocean heat transport are compensated by an opposite change in the atmospheric heat transport.

be close to zero, and because there are no low clouds in the warm pool, it is variations in the clear-sky OLR that most impact the radiation balance in the warm pool. The clear-sky OLR is controlled by the temperature and humidity of the atmosphere, which tend to have competing effects. As the warm pool temperature increases, for example, the free troposphere moves onto a warmer moist adiabat, which will tend to increase the OLR. However, as the temperature increases, the saturation specific humidity also increases. Given the model assumptions, the humidity of the entire free troposphere must also increase with the free tropospheric relative humidity remaining almost the same whatever the value of  $\gamma$ . In addition, a warmer warm pool will drive the CBL deeper and increase the humidity of the lower troposphere. Both these effects will tend to decrease the OLR. Shine and Sinha (1991) have also pointed out that changes in the humidity in this part of the atmosphere have a significant impact on the radiation budget of the planet.

The competing effects of temperature and humidity mean that the warm pool TOA balance is relatively insensitive to changes in the ocean heat transport. The sensitivity of OLR to SST in the model is about  $1.5 \text{ W m}^{-2} \text{ K}^{-1}$ , which is in close agreement with the observed spatial correlation between SST and OLR. On the other hand, the atmospheric heat transports vary approximately as  $\omega_{T_c}(c_p/g)(\text{SST}_\omega - \text{SST}_c)$ . Values for  $\omega_{T_c}$  are on the order of  $0.05 \text{ Pa s}^{-1}$ , which gives a sensitivity of atmospheric heat transports to warm pool SST, assuming a constant cold pool SST, of about  $5 \text{ W m}^{-2} \text{ K}^{-1}$ . Therefore, the warm pool more readily adjusts to changes in the ocean heat transport via changes in the horizontal transport of heat in the atmosphere than it can by altering the TOA radiative balance.

The reduction in tropical mean SST with increased SST gradients (Fig. 2b) is consistent with what was found in Hartmann and Michelsen (1993), but for a different reason. In their model, an increased SST gradient drives a stronger mass overturning in the atmosphere, which enhances evaporation and cools the Tropics. In the current model, the ultimate constraint is the TOA energy balance. Since the total radiative gain does not change, as SST gradients increase the mass overturning must decrease so that the total heat transport remains the same. The reduction in tropical mean SST with increased SST gradients comes about through feedbacks involving the effect of the CBL on the greenhouse capacity of the tropical atmosphere. Thus, the mean tropical SST is not simply the result of the coupling between the atmospheric circulation and the surface fluxes as Hartmann and Michelsen (1993) suggest, but rather is controlled by the adjustment of the atmospheric thermal and humidity structure to changes in the strength of the ocean heat transport.

#### b. The case of interactive cold pool low cloud cover

We now consider low-level marine stratus clouds in the cold pool as an interactive part of the tropical climate system. We will allow the cloud fraction to vary with the lower tropospheric dry static stability. Klein and Hartmann (1993) reported a linear relationship between lower tropospheric dry static stability and cloud cover as

$$f_c = 5.70(\theta_{700} - \text{SST}) - 55.73. \quad (25)$$

We find that this relation gives values of cloud cover that are too large in this model, and we instead use an intercept of  $-70$ .

Figure 6a shows the warm pool, cold pool, and mean SST as a function of  $\gamma$  with cold pool cloud fraction calculated. As  $\gamma$  increases, the SST difference between the warm pool and cold pool decreases. This lowers the tropospheric stability of the cold pool and decreases the cold pool cloud fraction (Fig. 6b), increasing the shortwave radiative gain of the cold pool. However, given the insensitivity of warm pool radiative gain, the cold pool must also have the same radiative gain over  $\gamma$  [Eq. (20)] even as the cloud fraction changes. The cold pool adjusts by increasing the SST, which, in turn, increases the OLR and balances the increased shortwave radiative gain such that there is no change in net radiative gain of the cold pool. The increase in cold pool SST over  $\gamma$  is amplified by an increased downward longwave flux associated with the moistening and deepening of the CBL. This is communicated to the warm pool via advection and leads to a warming there as well. As a result, the tropical mean SST *increases* by more than 2 K as  $\gamma$  varies from 0 to 1.

If the feedback between cloud reflectivity with CBL depth is eliminated, the results are qualitatively the same, but the tropical mean SST increases by about 5.5 K as  $\gamma$  varies from 0 to 1.

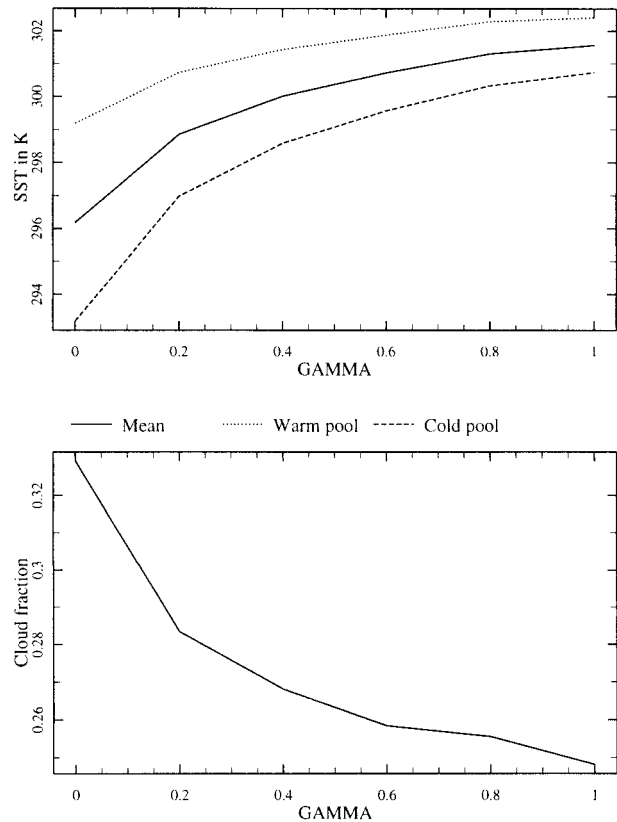


FIG. 6. (a) Warm pool, cold pool, and mean SST as a function of  $\gamma$  in the box model (in K) with variable cold pool cloud cover and (b) cold pool cloud fraction.

#### c. Summary

The primary mechanism by which the tropical climate adjusts to changes in ocean heat transport is through changes in the atmospheric heat transport. In all three cases (fixed cold pool cloud cover and variable reflectivity, fixed cold pool cloud cover and fixed cloud reflectivity, variable cold pool cloud cover), the change in evaporation required to balance the change in ocean heat transports is partly accomplished by changes in surface relative humidity, diminishing the SST change. As the ocean heat transport increases, the cooling and drying of the warm pool CBL by horizontal advection in the atmosphere decreases due to the decreased SST difference. This leads to an increase in the warm pool surface relative humidity, which decreases the evaporation and offsets the increased cooling of the ocean by heat transport. In the cold pool, as the ocean heat transport increases, the decreased SST difference reduces the static stability and leads to a drying of the CBL by increased subsidence. This decreases the cold pool surface relative humidity and allows the evaporation to increase and offset the heating by ocean heat transports. In short, these results illustrate the highly nonlocal nature of the adjustment of the Tropics to changes in the

ocean heat transport. Small changes in relative humidity, well within the range of spatial variations in the Tropics, can yield large changes in evaporation, altering the notion of local control of evaporation through SST.

In all three cases, the mean tropical SST changes over  $\gamma$  while the *total radiative gain of the Tropics remains the same*, which has to be the case since the heat transports to midlatitudes are fixed. This means that, as the ocean heat transport increases, the SST gradient decreases, and the overturning mass flux in the atmosphere must increase for the total heat transport to remain the same. The changes in mean SST are the result of the processes in the atmosphere that are affected by the SST gradient. We found three possible ways by which changes in ocean heat transport can affect the tropical mean SST with the differences depending on the degree of coupling between the cold pool CBL and the tropical climate. The possibilities are the following.

- In the case where the cold pool cloud fraction is held fixed but the reflectivity is allowed to vary (Fig. 2a), the cold pool CBL deepens as the ocean heat transport increases. This increases the reflectivity of the cold pool clouds and hence the albedo of the Tropics. To balance this, the tropical SST decreases and the OLR is reduced.
- In the case where both the reflectivity and cloud fraction in the cold pool are held fixed (Fig. 2b), the deeper CBL with increased ocean heat transport increases the greenhouse capacity of the tropical atmosphere, and the mean surface temperature warms.
- When the cold pool cloud fraction is allowed to vary (Fig. 6a), the cold pool cloud cover decreases and the shortwave radiative gain increases as the ocean heat transport increases. To balance this, the tropical SST increases, and the OLR is increased.

While the magnitude of some of these feedbacks is uncertain, these results demonstrate the possibility that rather different tropical climates can be obtained through processes *internal* to the Tropics without being driven by the mid- and high latitudes.

All of these results occur within the constraint of a total amount of low-latitude poleward heat transport that is *independent* of the efficiency of ocean heat transports. The fixed total heat transport is the result of strong constraints on the clear-sky OLR in the presence of a vertical relative humidity profile that is almost constant as climate changes. The box model is idealized, and the magnitude of the feedbacks is uncertain. We will now turn to some GCM experiments in order to assess whether this constraint on the total heat transport operates in more complex models.

#### 4. The effect of ocean heat transports in GCMs

Manabe (1969) performed a set of GCM experiments that allowed quantitative examination of the effect of ocean dynamics in the global climate. He compared the

climatologies of an AGCM coupled to a mixed layer ocean to that of the AGCM coupled to a dynamically active ocean. Like our model, when the ocean can move heat, the meridional gradient in SST is diminished, and the TOA balance in the Tropics remains almost unchanged. That is, the total heat export from the deep Tropics is the same whether all of it occurred in the atmosphere or half was in the ocean. Manabe (1969) claimed that this result was due to the fact that the effect of the ocean is only felt in the lower part of the troposphere and thus would have little effect on the outgoing radiation at the TOA. This cannot be the case in the Tropics where deep convection efficiently communicates SST changes throughout the troposphere. The deep tropical troposphere was up to 4 K warmer in the uncoupled GCM where the ocean transports no heat. With no change in specific humidity this would increase the OLR by about  $12 \text{ W m}^{-2}$ . However, the relative humidity distribution was about the same in the two experiments. The higher specific humidity in the uncoupled GCM tends to decrease the OLR and opposes the direct temperature effect such that the OLR remains relatively constant. This effect operates in the box model over the whole range of ocean heat transports, and we suggest that it is the reason why the total atmosphere-ocean heat transport in the two GCM experiments remained fixed.

We performed similar experiments with the Goddard Institute for Space Studies (GISS) “*q*-flux” model (Hansen et al. 1984). This model is considerably advanced over the 1969 Geophysical Fluid Dynamics Laboratory model and includes prognostic clouds and liquid water vapor, state-of-the-art radiation and convection parameterizations, realistic land-sea distribution, orography, etc. (Del Genio et al. 1996). The control run has an ocean mixed layer and a specified ocean heat transport. This *q*-flux is diagnosed as equivalent to the net surface heating in a run with fixed modern SSTs. The model is then run with calculated SSTs imposing this ocean heat flux (Russell et al. 1985). In a second experiment, we set the *q*-flux to be zero everywhere. Figure 7 shows the net surface heat flux (i.e., the *q*-flux) in the control run. There is a cooling of the warm pool of approximately  $40 \text{ W m}^{-2}$  and a warming of about the same magnitude in the subtropics. Without ocean heat transports, the TOA balance remains nearly unchanged (Fig. 8a). That is, the atmospheric heat transport adjusts to compensate almost precisely for the change in ocean heat transport. Figure 8b shows the solar and longwave components of the TOA balance. When the ocean heat transport is set to zero, the warmest waters, convection, and associated cloud cover move onto the equator, the absorbed solar radiation decreases, but the OLR also decreases leaving the TOA balance fairly unchanged. The change in the zonally averaged atmospheric temperature between the two runs is similar to that of Manabe’s experiments, where the entire tropical free troposphere is warmer when the ocean does

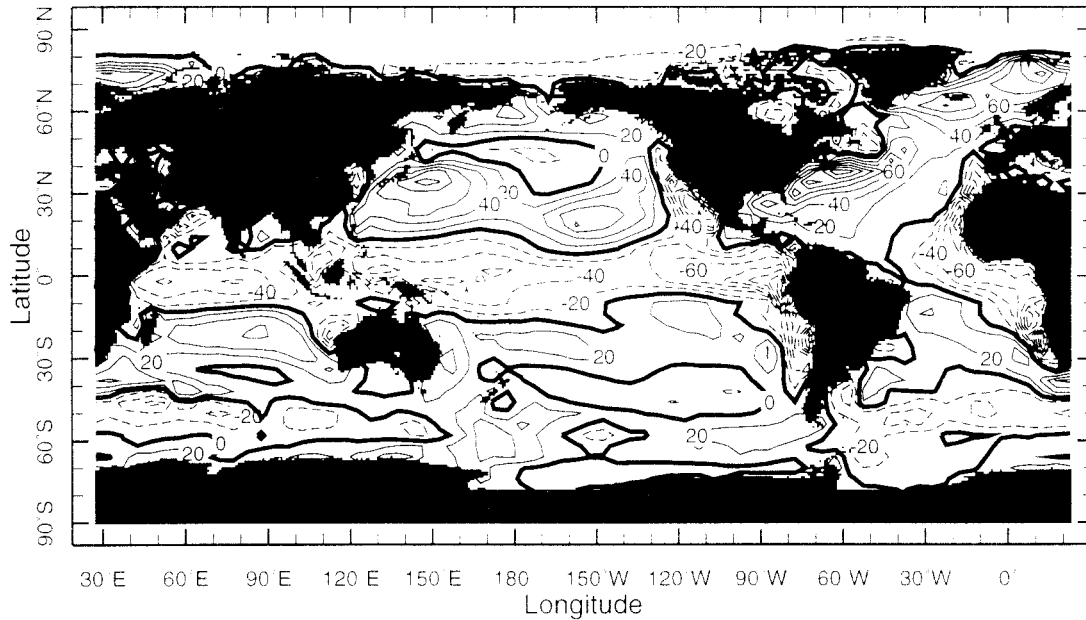


FIG. 7. The  $q$ -flux, or specified ocean heat transport, in the GISS AGCM in  $\text{W m}^{-2}$ . This  $q$ -flux is diagnosed as equivalent to the net surface heating in a run with fixed modern SSTs. The model is then run with calculated SSTs imposing this ocean heat flux (Russell et al. 1985).

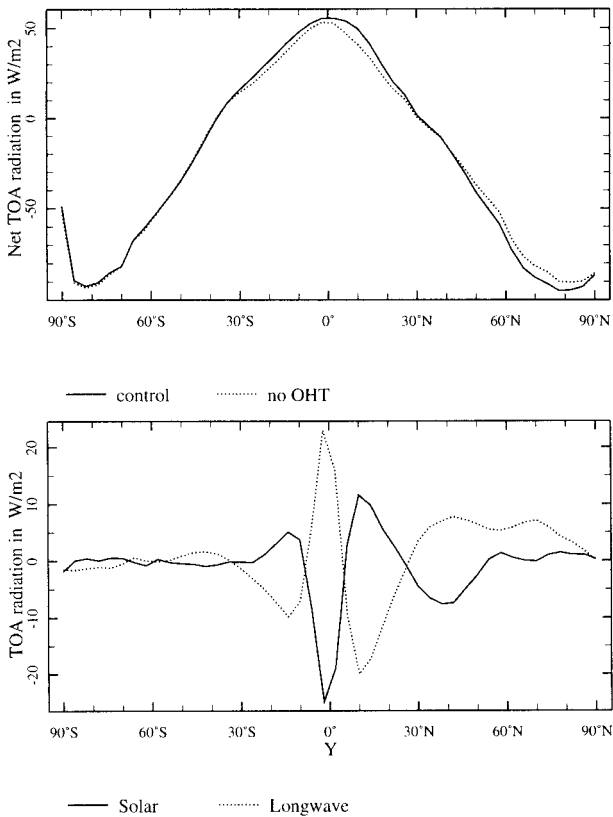


FIG. 8. (a) Net TOA balance for the control run in the GISS  $q$ -flux model (solid) and an experiment with no ocean heat transports (dotted) in  $\text{W m}^{-2}$ . (b) Change in shortwave (solid) and longwave (dotted) radiative gain for no-ocean heat transport case minus control.

not transport heat (Fig. 9a). Again, this would tend to decrease the radiative gain of the deep Tropics by increasing the OLR. The relative humidity increases by up to 10% in the deep Tropics where the temperature change is largest (Fig. 9b). The increase in specific humidity is therefore even larger than it would have been if the relative humidity remained constant, thus opposing the effect of temperature on OLR even more strongly than in the fixed relative humidity case.

Figure 10 shows the change in SST, for the no-ocean heat transport run minus the control run. With no ocean heat transports, the subtropics are considerably cooler (up to 4 K), while the warmest waters warm only slightly. The change in ocean heat transport is primarily balanced by a change in evaporation that gives a larger temperature change where the temperatures are lower. In the GCM ocean heat transports increase the mean tropical SST of the Pacific by about 0.2 K. This is most similar to the experiments with the box model where the radiative properties and cloud fraction of the subtropical marine stratus clouds are held fixed (Fig. 2b), and the mean tropical SST increases with the efficiency (and strength) of ocean heat transports. The feedback between low cloud cover and ocean heat transports found in the box model does not appear to be present in this GCM. However, the model's simulation of low cloud cover is generally unsatisfactory (Del Genio et al. 1996). Note that the spatial pattern of the temperature change in the deep Tropics is different from that of the change in ocean heat transport. This is related to a shift in the core of the trade winds that move

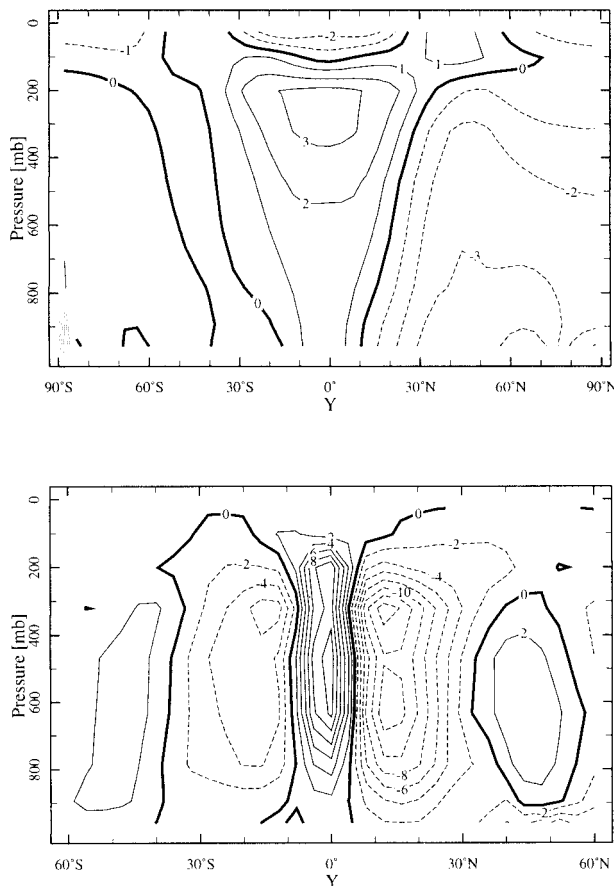


FIG. 9. (a) Change in zonally averaged temperature in the GISS model with no-ocean heat transport case minus control. (b) Change in zonally averaged relative humidity in percent.

equatorward and intensify in the case with no ocean heat transport.

The fixed total heat transport is now evident in the simple box model and in two very different GCMs. This result is underpinned by the near constancy, or positive correlation with temperature, of the free tropospheric relative humidity as climate changes. The processes that determine the humidity of the upper troposphere is one of the outstanding problems in climate change research (Intergovernmental Panel on Climate Change 1995). Numerous processes going down to the scale of the microphysics of clouds have been identified as playing a role in determining this quantity (Lindzen 1990; Sun and Lindzen 1993; Renno et al. 1994). Model treatments of these processes tend to be simple and are highly parameterized. It is unclear whether the close-to-fixed relative humidity condition in the upper troposphere is a property of the real world. However, it clearly puts a constraint on the total amount of poleward heat transport at low latitudes in these models.

**5. The role of the ocean in climate change and possible implications for GCMs**

Projections of climate change under increasing CO<sub>2</sub> have been carried out in climate models with a variety of treatments of the ocean from a mixed layer ocean (Mitchell et al. 1995), to a *q*-flux ocean with fixed ocean heat transports (Hansen et al. 1984), to fully coupled ocean atmosphere GCMs (Manabe and Stouffer 1994). Only the last is dynamically consistent. What effect does the lack of dynamical consistency have on the model's simulation of climate change? Is the sensitivity to increasing CO<sub>2</sub> the same whether or not

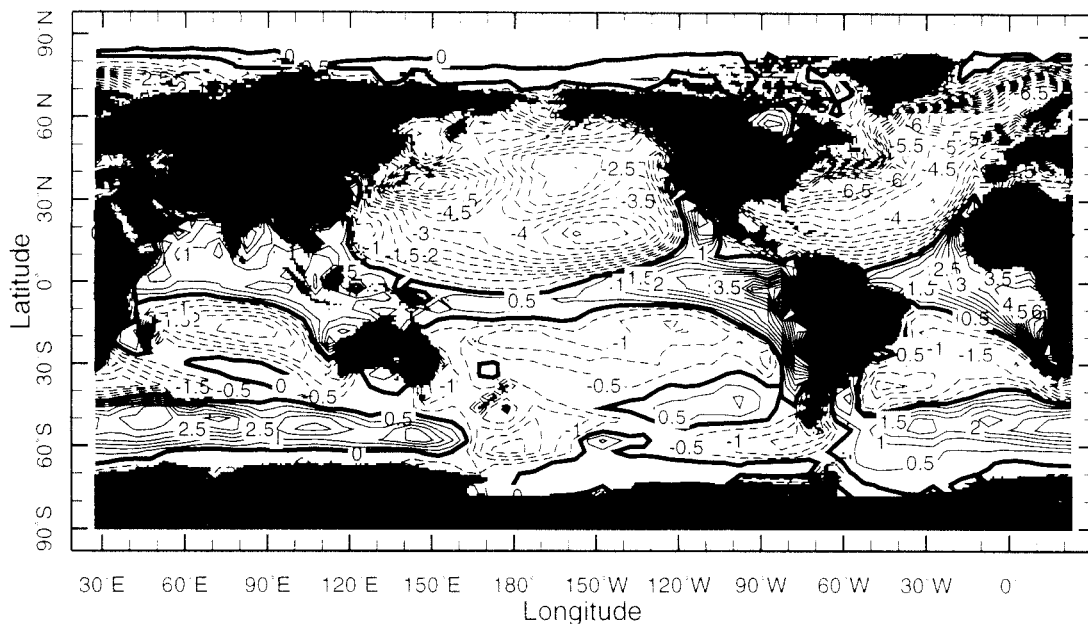


FIG. 10. Change in SST for the no-ocean heat transport minus control GISS model runs.

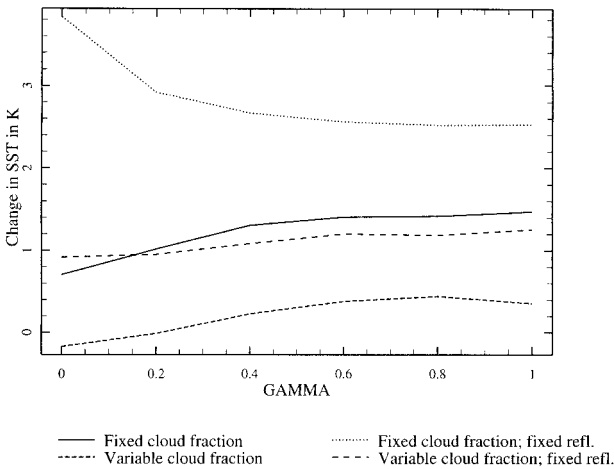


FIG. 11. Mean tropical SST change with  $2 \times \text{CO}_2$  with fixed cold pool cloud fraction (solid), with varying cloud reflectivity and fixed cloud reflectivity at  $\gamma = 0$  values (dotted), variable cold pool cloud fraction with varying cloud reflectivity (short dashes), and varying cloud fraction and fixed cloud reflectivity (long dashes).

ocean heat transports are an interactive part of the system? Does the sensitivity depend on the strength of ocean heat transports? It is difficult to answer these questions with GCMs because of the great computational expense involved in running those models. Here we attempt to address these issues within the context of the box model.

#### a. The case of fixed cold pool low cloud cover

We first perform doubled  $\text{CO}_2$  experiments with  $\gamma$  varying from 0 to 1 and fixed cold pool low cloud cover. As the climate warms, the CBL depth increases, which increases the reflectivity of the cold pool clouds acting as a negative feedback to greenhouse warming. This effect is larger at low values of ocean heat transports where the CBL depth and liquid water contents are lower (Fig. 11, solid line).

If this feedback is eliminated by holding the cold pool cloudy sky solar fluxes at their values for  $\gamma = 0$  and present-day  $\text{CO}_2$ , the mean temperature change is significantly larger for all values of  $\gamma$  (Fig. 11, dotted line) but is largest at small  $\gamma$ . Without the cloud reflectivity feedback operating, the cold pool must warm more at low values of  $\gamma$  where the temperatures are colder in order to balance the excess downward longwave radiation with increased evaporation. The large warming of the cold pool has the effect of deepening and moistening the entire tropical CBL, which enhances the water vapor greenhouse trapping and gives a larger sensitivity to doubled  $\text{CO}_2$  at low  $\gamma$ .

#### b. The case of interactive cold pool low cloud cover

Miller (1997) has suggested that low clouds in the cold pool act as a negative feedback to greenhouse

warming. Here we test whether there is any change in sensitivity when the clouds and ocean heat transports (OHTs) can interact. We perform the same  $\text{CO}_2$  change experiments as in the previous section but with clouds varying interactively according to Eq. (25) (with the intercept set to  $-70$  again). As in Miller (1997), we find that the cloud feedback is negative (Fig. 11, short dashes). Again, this relates to the insensitivity of warm pool radiative gain and total heat transport. As  $\text{CO}_2$  increases, the cold pool subsidence, or the strength of the Hadley cell, decreases in this model as in most GCMs. The reason for this is that as the Tropics move to a warmer and moister moist adiabat, the static stability increases, which reduces the subsidence required to balance the atmosphere radiative cooling (BR89). With a decreased value of  $\omega_{T_c}$ , the SST difference must increase in order to maintain the same heat transport. The increased SST difference increases the cloud cover, which reduces the radiative gain of the cold pool. This strong negative feedback in the cold pool is readily communicated to the warm pool by horizontal heat transports reducing the mean tropical doubled  $\text{CO}_2$  warming. The strength of this feedback is also a function of  $\gamma$ . At low values of  $\gamma$ , where the cold pool CBL and liquid water contents are lower, the additional negative feedback involving the increased reflectivity of clouds in the warmer climate almost prevents the mean SST from changing at all. At higher values of  $\gamma$ , where the cold pool CBL is deeper, this negative feedback diminishes and only the cloud fraction feedback operates. Nonetheless, the mean temperature change associated with doubling of  $\text{CO}_2$  is only 0.4 K. This dependence on  $\gamma$  is eliminated if the cloudy-sky solar fluxes in the cold pool are held fixed (Fig. 11, long dashes).

In all cases, the sensitivity of the tropical SST in the box model to doubled  $\text{CO}_2$  is within the range of the models referenced in the Intergovernmental Panel on Climate Change (1995) report. The close-to-fixed relative humidity, and hence strong water vapor feedback, of all these models is mainly what produces this sensitivity.

#### c. Validity of fixed ocean heat transports in GCM climate change experiments

The results discussed above (summarized in Fig. 11) are the same if, instead, we fix the ocean heat transport at different values rather than allowing it to be determined interactively. That is, the sensitivity to doubled  $\text{CO}_2$  depends on the *strength* of the ocean heat transport, but not on whether it is determined interactively or not. With increased  $\text{CO}_2$ , the total radiative gain of the warm pool remains the same, and hence the total heat transport must remain the same. The parameter  $\gamma$  controls the partitioning of the heat transport, which does not change with increased  $\text{CO}_2$ . Thus for a given  $\gamma$  the ocean heat transport will remain the same as the  $\text{CO}_2$  changes.

What happens, however, if the warm pool radiative gain can change with climate? In this model, the constraint on the radiative gain of the warm pool relies in part on the cancelation between the shortwave and longwave forcing of high clouds. While the cancelation appears to hold for the present climate, and over interannual variations in climate associated with El Niño (Hartmann and Michelsen 1993; Chou 1994), it is unclear whether it will hold under radiative perturbations. As an example of how the coupled system operates as the total heat transport changes, we allow the liquid water content of the warm pool high clouds to increase with the warm pool tropospheric temperature. We perform experiments with the high cloud shortwave feedback for the present day  $\text{CO}_2$  and for doubled  $\text{CO}_2$  with both fixed ocean heat transport and with the ocean heat transport determined interactively by specifying a value of  $\gamma$ . The warm pool high cloud cover is set to 40% as before. The cloud cover and reflection in the cold pool are held fixed (as in the experiments shown in Fig. 2b). As the warm pool warms with increased  $\text{CO}_2$ , the liquid water content and reflectivity of the high clouds increases, thus acting as a negative feedback. This is analogous to the cirrus cloud thermostat idea suggested by Ramanathan and Collins (1991). The increase in cloud reflection decreases the radiative gain of the warm pool, which reduces the total poleward heat transport by about  $5 \text{ W m}^{-2}$ . The reduction in radiative gain is the same regardless of the value of  $\gamma$  or what the specified ocean heat transport is. That is, the strength of the negative high cloud feedback is determined locally, and the heat transports simply adjust.

In the case of interactive ocean heat transports, both the ocean and atmospheric heat transports are reduced in the doubled  $\text{CO}_2$  case, leaving the partitioning between the ocean and atmosphere the same as in the present climate. When the ocean heat transports are fixed, however, the reduction in heat transport occurs only in the atmosphere. Thus, the partitioning changes, and the ocean transports a larger percent of the total heat in the doubled  $\text{CO}_2$  case than for the present climate. As discussed in section 3, with fixed cloud reflection and cover, as the fraction of total heat transport occurring in the ocean increases, the mean temperature increases since the ocean moves heat to a region where a larger temperature change is needed to balance the ocean heat transport by evaporation (Fig. 2b). Therefore, the doubled  $\text{CO}_2$  climate is warmer when the ocean heat transports are fixed than when they are interactive because the ocean transports a larger fraction of the total heat in the former case. However, the effect on the sensitivity to doubled  $\text{CO}_2$  is small (about 10%) since the changes in heat transports are small (less than  $5 \text{ W m}^{-2}$ ). In short, if the changes in the warm pool radiative gain are small in response to changes in  $\text{CO}_2$ , the zonal mean ocean circulation plays little role in the adjustment of the tropical climate to changes in  $\text{CO}_2$ .

#### d. An ocean dynamical thermostat?

Cane et al. (1997) have suggested that the meridional overturning in the ocean should increase with increasing  $\text{CO}_2$ . They argue that, as the tropical Pacific Ocean is heated uniformly, the temperature change will initially be smaller in the eastern Pacific where upwelling strongly opposes the surface heating. This will increase the zonal SST gradient. The increase will be amplified through coupled interactions, resulting in a more zonally asymmetric ocean with increased meridional overturning (Clement et al. 1996). Seager and Murtugudde (1997) used an ocean GCM coupled to an atmospheric mixed layer in order to show that this mechanism indeed reduced the sensitivity of the tropical SST to a uniform heating. They find that the oceanic heat advection from the deep Tropics to regions of higher wind speed in the subtropics *reduces* the temperature change associated with a uniform increase in the heat flux into the ocean surface. This is because the SST change needed to balance the forcing with an increase in evaporation is smaller where the wind speeds are larger, that is, in the subtropics. The result is a cooler mean SST and a permanent increase in the equatorial gradient of SST, which, they point out, would be amplified through coupled interactions. The model used by Seager and Murtugudde, however, does not account for changes in the surface relative humidity with wind speed. Greater wind speeds tend to increase the surface relative humidity (BR89) and will reduce the sensitivity of evaporative heat flux to SST.

In Table 1 we compare the mean SST change for doubled  $\text{CO}_2$  when  $\gamma$  is held fixed at a value of 0.2 to that with  $\gamma = 0.4$ , both relative to the present climate with  $\gamma = 0.2$ . Whether or not the clouds are interactive or the wind speed is higher in the cold pool than the warm pool, the mean temperature is either unchanged or actually increases with increasing heat transport. For the case of fixed cloud fraction and reflectivity, this is due to the large increase in cold pool SST with ocean heat transport, which follows from the nonlinearity of the Clausius–Clapeyron relation, together with its amplification by the greenhouse effect. With variable clouds, the increased mean SST is due to the reduction in cloud cover at higher values of ocean heat transports. It therefore seems that increasing the efficiency of ocean heat transports is likely to increase the temperature response to doubled  $\text{CO}_2$  rather than decreasing it as Cane et al. (1997) proposed. However, in terms of the twentieth-century climate change, we cannot consider the tropical climate to be in equilibrium because the forcing changes with time and because the thermocline temperature takes a finite time to adjust. Thus, the mechanism proposed by Cane et al. (1997) may operate as a transient phenomenon that temporarily reduces the tropical warming due to increased greenhouse gases. This issue can only be rigorously assessed with a coupled GCM. Cane et al. (1997) have argued that the

TABLE 1. Tropical mean SST change from present CO<sub>2</sub> to doubled CO<sub>2</sub>. The first column is for the same value of  $\gamma$  ( $\gamma = 0.2$ ), or ocean heat transport, from present CO<sub>2</sub> to doubled CO<sub>2</sub>, and the second column is for OHT ( $\gamma = 0.4$ ) in the doubled CO<sub>2</sub> climate relative to the present CO<sub>2</sub> climate with  $\gamma = 0.2$ . The first three rows show the mean SST change when the cold pool and warm pool wind speed are equal ( $0.1 \text{ Pa s}^{-1}$ ), and the last three rows are when the cold pool wind speed is double ( $0.2 \text{ Pa s}^{-1}$ ) that of the warm pool. In all cases, the equilibrium SST change is greater when the ocean heat transport is larger.

$2 \times \text{CO}_2$ tropical mean SST change	Fixed OHT ( $\gamma = 0.2$ )	Increased OHT ( $\gamma = 0.4$ )
Warm pool and cold pool wind speed equal ( $0.1 \text{ Pa s}^{-1}$ )		
Fixed cloud fraction; variable cloud reflectivity	1.3 K	1.2 K
Fixed cloud fraction; fixed cloud reflectivity	2.9 K	3.2 K
Variable cloud fraction	0.2 K	1.2 K
Higher cold pool wind speed ( $0.2 \text{ Pa s}^{-1}$ )		
Fixed cloud fraction; variable cloud reflectivity	1.5 K	1.8 K
Fixed cloud fraction; fixed cloud reflectivity	2.2 K	3.2 K
Variable cloud fraction	0.6 K	1.7 K

current generation of coupled GCMs is not up to the task because of weak dynamic coupling in the equatorial region and poor representation of the equatorial thermocline.

## 6. Discussion and conclusions

We find that changes in ocean heat transports are capable of changing the mean tropical climate by several degrees. These changes are accomplished either through the effect of the ocean on the properties of low clouds or on the total greenhouse effect of the tropical atmosphere. Atmospheric heat transports are the prime mechanism by which the tropical climate adjusts to changes in OHTs. With a sensitivity of about  $5 \text{ W m}^{-2}$  per degree change in the SST gradient, it is larger than the local adjustment through changes in the TOA radiative balance ( $1\text{--}2 \text{ W m}^{-2} \text{ K}^{-1}$  for OLR). These results illustrate the highly nonlocal nature of the determination of SST in the Tropics.

The large sensitivity of the atmospheric heat transports means that the total amount of poleward heat transport at low latitudes is independent of the partitioning between the ocean and atmosphere. This constraint on the tropical climate system implies that an increase in the SST gradient will be accompanied by a decrease in the overturning mass flux in the atmosphere, contrary to what is assumed in some simple models of the tropical climate (e.g., Hartmann and Michelsen 1993; Liu and Huang 1997). This result is subject to the assumption that the relative humidity of the warm pool does not change considerably with climate and that the longwave and shortwave effects of high clouds in the warm pool

cancel. We have found that the close-to-fixed relative humidity result holds in the box model and two rather different GCMs. However, substantial changes in relative humidity are precluded from our model by design. LH99 have made an attempt to parameterize changes in the relative humidity of the warm pool in a similar box model. A useful extension of these models would be inclusion of a deep convection scheme such that the model relative humidity was prognosed and could change as climate changes if necessary (e.g., Emanuel 1991). Upper tropospheric humidity could then be modeled outside the regions of deep convection using the method of Salathe and Hartmann (1997). Such a model would include most of what is known about how upper tropospheric humidity is determined and could be a useful tool for further research.

One aspect of the tropical climate that is missing from this framework is the effect of the relative areas of the warm and cold pools on the tropical climate. We have chosen to fix these areas in order to consider the effect of changes in the zonal mean ocean heat transport alone. However, as Pierrehumbert (1995), Miller (1997), and LH99 have pointed out, a change in the relative size of the warm and cold pool can also significantly affect the tropical climate. In those studies, the ratio of the area of the warm pool to cold pool was an external parameter that was varied as a sensitivity study. However, this ratio is presumably an interactive part of the tropical climate that will change as the climate changes. There is no theory for what sets the relative sizes of the warm pool in the present climate, nor for how they will change as the climate changes. It is likely to be related to the three-dimensional ocean atmosphere circulation and will need to be addressed with a coupled model that represents the details of the dynamics of the tropical climate.

We found that variations in CBL depth can significantly modify climate, both through its impact on cloud liquid water and through its effect on the total atmospheric water vapor content and, hence, the OLR. In the current climate, the CBL depth is observed to be quite uniform with cloud top at about 800 mb (Schubert et al. 1995). It seems reasonable that, as the tropospheric stability alters as climate changes, the CBL depth will vary. To date there has been almost no work done that looks at this question. Again, while believing in the signs of the model feedbacks, we have doubt about their strength. For example, in the current climate, the increase in CBL depth from the cool subtropical waters over the eastern oceans to the warmer equatorial waters is associated with a change in cloud type from stratus to trade cumulus (Albrecht et al. 1995). If the mean CBL depth alters as climate changes, will the cloud type and CBL structure also change? Such a possibility is not allowed for in the model. Schubert et al. (1995) have also shown that large gradients in CBL depth cannot be dynamically sustained in the Tropics because they would cause large gradients in temperature. Albrecht



(1993) has suggested that the onset of drizzle also restricts the CBL depth. We do not include these constraints on CBL depth in our model.

We believe that CBL processes, in particular the generation of CBL clouds, are important mechanisms whereby the climate comes into equilibrium. It is well known that low-level marine clouds are responsible for cooling the planet (e.g., Hartmann and Short 1980; Slingo 1990). It is reasonable to suppose that they are also capable of influencing climate change, and we found that they do in our model. Nonetheless, we are rather uneasy about the strengths of the low cloud feedbacks in the model. Both the liquid water and the cloud coverage feedbacks are strong. The strength suggests that the mixing line dependence of liquid water content and stability dependence of cloud cover may alter as climate changes. Understanding this will require a considerable advance in our knowledge of marine clouds. Fortunately, experiments such as the Atlantic Stratocumulus Transition Experiment (Albrecht et al. 1995) exist that may provide us with that knowledge. Generally, more research examining how CBL depth varies in space and with climate change, and how cloud liquid water and cover are determined, will be required before we can be sure of the validity of any of the CBL feedbacks suggested in the current work. It is hoped that the work presented here suggests some directions for that research.

It is found that the strength of the zonal mean ocean heat transport has an effect on the sensitivity of the tropical climate to increasing  $\text{CO}_2$ . However, the zonal mean ocean circulation plays little role in the adjustment of the tropical climate to changes in  $\text{CO}_2$ . We do not suggest that the tropical ocean will not play a role in climate change. Instead, this result points to the fact that it is the three-dimensional circulation of the ocean and atmosphere that is likely to have an effect on the sensitivity of the tropical climate to radiative perturbations. The work of Cane et al. (1997) suggests that changes in the zonally asymmetric circulation are important as  $\text{CO}_2$  changes and may act as a negative feedback to greenhouse warming. Here we point out that, when full account is taken of the atmospheric adjustment, a climate state with increased meridional overturning in the ocean may not yield a cooler climate. To properly resolve this issue, it is necessary to understand how the Walker and Hadley circulations interact, and whether a change in the ocean-atmosphere circulation can change the tropical mean SST. Furthermore, the changes in the distribution of the tropical SST that may arise from such circulation changes would have a significant impact on the global atmosphere and would thus play an important role in the response of the climate to changes in  $\text{CO}_2$  (e.g., Clement et al. 1998).

A variety of new data suggests that tropical temperatures were up to 4–5 K cooler at the last glacial maximum than in the modern climate (Guilderson et al. 1994, Stute et al. 1995; Schrag et al. 1996; Colinvaux

et al. 1996; Bard et al. 1997). A large part of this glacial cooling may have resulted from increased heat transport to mid- and high latitudes, where ice cover would have been a significant radiative sink. GCM experiments with a mixed layer ocean model that include the effects of increased high latitude albedo show only about a 2-K tropical SST change (Broccoli and Manabe 1987). Bush and Philander (1998) have shown that inclusion of ocean dynamics is essential in explaining this large glacial cooling in the Tropics. Here we have shown that significant changes in tropical SST can arise with no change in the export of heat to high latitudes but are solely due to changes internal to the Tropics. Future work will address how such changes in circulation can arise when the interaction between the Hadley and Walker circulations is accounted for. It is possible that a rearrangement of the three-dimensional tropical atmosphere and ocean circulation, driven by redistribution of incoming solar radiation associated with Milankovitch cycles, may have contributed to the glacial cooling. The current work suggests that coupling between the tropical atmosphere and ocean circulation can play an active role in climate change from the glacial past to the greenhouse future.

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