Glacial Cooling in the Tropics: Exploring the Roles of Tropospheric Water Vapor, Surface Wind Speed, and Boundary Layer Processes*

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(Manuscript received 4 January 1999, in final form 5 August 1999)

ABSTRACT

This paper is a modeling study of possible roles for tropospheric water vapor, surface wind speed, and boundary layer processes in glacial cooling in the Tropics. The authors divide the Tropics into a region of persistent deep convection and a subtropical region with no deep convection. The regions are coupled via a radiatively driven Hadley cell and a wind-driven meridional overturning cell in the ocean. Radiation and the convective boundary layer (CBL) are treated in some detail.

The amount of tropical cooling depends on the height of the tropospheric drying and on the extent to which cloud water in the CBL is converted into rainwater. In the most realistic case where the CBL clouds precipitate, variations in CBL depth are small, and the tropical SST becomes most sensitive to drying immediately above the CBL. Reducing the relative humidity of the entire troposphere above the subcloud layer by about 10%–20% cools the tropical SST by just over 2 K. It is shown that this climate sensitivity arises from a complex balance of processes that control the depth of the CBL, its greenhouse trapping, and the albedo of boundary layer clouds. An increase in surface wind speed, such as occurs in simulations of the last glacial maximum with coupled general circulation models, substantially reduces the SST although the change in surface air temperature is less. The Milankovitch cycles are expected to cause changes in atmosphere and ocean circulation. It appears that a circulation change that causes the lower midtroposphere to dry would be an effective way to induce strong cooling of tropical climate.

1. Introduction

During the Quaternary period advances and retreats of ice sheets have been shown to be largely synchronous between the hemispheres on orbital timescales and, frequently, on sub-Milankovitch timescales (Lowell et al. 1995). This observation presents difficulties for theories of glacial cycles that rely on insolation changes in midlatitudes, associated with changes in the Earth's orbit, since these changes are often out of phase between the hemispheres at times when ice advance or retreat was in phase (Lowell et al. 1995). While it has been proposed that the oceanic thermohaline circulation is capable of organizing global climate (Broecker and Denton 1990), experiments with coupled general circulation models (GCMs) suggest that the effects of cessation of North Atlantic Deep Water formation are limited to the North Atlantic sector (Manabe and Stouffer 1988). Reductions in atmospheric CO_2 are capable of organizing global climate change. However, recent model estimates suggest that, even in the presence of continental ice sheets and reduced CO_2, the tropical oceans cool by less than 2 K (Hewitt and Mitchell 1997; Broccoli 2000).

A small tropical cooling is consistent with the SST reconstructions of Climate: Long-Range Investigation, Mapping, and Prediction (CLIMAP) (CLIMAP Project Members 1976), which were based on foraminiferal assemblages. It is also roughly in line with the sea surface temperature (SST) cooling reconstructed from alkenone data (Bard et al. 1997), which suggests a quite uniform 2-K cooling in the Tropics. However, other recent estimates, based on geochemical analyses of Sr/Ca (Guilderson et al. 1994) and noble gases (Stute et al. 1995) indicate much larger coolings of as much as 5 K. Also, while some oxygen isotope data seem to agree with the CLIMAP estimates, it has been argued that allowing for pore water ice volume (Schrag et al. 1996) and pH (Spero et al. 1997) influences brings these data more in line with the colder estimates. There is also significant physical evidence of cooling in excess of 2 K provided by indications that tropical snow lines lay almost 1000 m below their present levels (e.g., Rind and Peteet 1985).

Recently, Broecker (1997) has used observed glacial lowering of tropical snow lines, together with variations

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in $\delta^{18}O$ recorded in the Huascaran ice cap (Thompson et al. 1995), to infer that the glacial tropical free troposphere had lower specific *and* relative humidity than at present. Broecker suggests that during glacial times the tropical climate system operated in such a way that the water vapor greenhouse effect was greatly reduced relative to today causing tropical, and global, cooling. This theory suggests that the Tropics are the ringmaster organizing global climate, which would help explain hemispheric synchroneity of glacial cycles and significant tropical cooling.

Broecker (1997) was unable to suggest why the tropical free troposphere was drier with lower relative humidity (RH) than at present. Indeed, Pierrehumbert (1999a) has questioned his interpretation and drawn attention to potential pitfalls in isotopic paleoclimate proxies, which may, for example, record nothing more than changes in source vapor due to minor circulation changes. Nonetheless, the processes that control free tropospheric humidity are poorly known. The vertical profile of RH in regions of active convection is established through complex interactions between convective-scale circulations and cloud microphysics (Emanuel 1991). It has been suggested that changes in the strengths of the Walker and Hadley circulations (Salathé and Hartmann 1997), alterations in the process of reevaporation of hydrometeors (Sun and Lindzen 1993), and changes in lateral mixing (Pierrehumbert and Roca 1998; Pierrehumbert 1998) could also alter the relative and specific humidity distributions. Pierrehumbert (1999b) provides a review of ways in which subtropical humidity might be altered and the possible impacts on climate. Although we are unsure how the tropical humidity distribution can be perturbed, it is nonetheless useful to assess the extent to which such changes could *cause* climate change. Can reductions in RH explain tropical glacial cooling?

Most previous attempts to examine the sensitivity of tropical temperatures to atmospheric water vapor content have reasoned on radiative grounds (e.g., Lindzen 1990) or used simplified radiative–convective models that greatly restrict how the vertical distribution of humidity will alter in response to imposed perturbations (Shine and Sinha 1991). The results of these investigations can be understood in terms of Fig. 1. Here we show the increase in outgoing longwave radiation (OLR) in watts per square meter that occurs when the RH is reduced by 10% in a 100-mb layer centered at the pressure and latitude indicated (i.e., if the RH was 50%, then it is reduced to 40%; if it was 20%, then it is reduced to 10%, and so on). The unperturbed temperature and specific humidity fields were taken from European Centre for Medium-Range Weather Forecasts analyses and the radiative transfer schemes used derived from Ramanathan and Downey (1986) and Kiehl and Briegleb (1991). The OLR is most sensitive to reductions in RH in the midtroposphere. Drying at the tropopause has little effect because the specific humidity

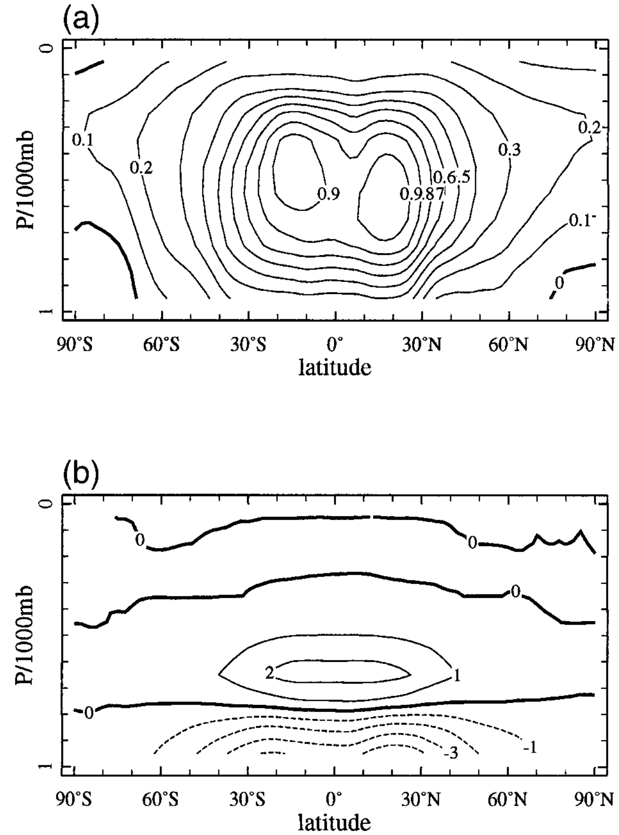

FIG. 1. (a) The sensitivity of OLR to the vertical distribution of reductions in relative humidity. The values plotted are the increase in OLR for a 10% reduction in relative humidity at that latitude in a 100-mb layer centered at the pressure indicated on the vertical axis. (b) The change in radiative flux divergence across the lowest 300 mb of the atmosphere forced by a reduction in relative humidity computed as in (a). Reductions in relative humidity immediately above the CBL increase the radiative flux divergence of the CBL, while changes at the tropopause have little effect. Drying of the CBL reduces the flux divergence.

is so small there that only a very large reduction in RH would be radiatively significant. Drying near the surface also has little effect on the OLR since almost all radiation emitted from those levels is reabsorbed higher in the atmosphere. This suggests that drying in the midtroposphere would be the most effective way to cool tropical SSTs.

There are reasons to believe that the story is not this simple. For example, if a change in the operation of deep convection attempts to dry the free troposphere, we expect that the vertical distribution of specific and relative humidity throughout the column will alter as a new climatic equilibrium is established. One way this might occur is through variations in the depth of the convecting boundary layer (CBL) that couples the lower and moister levels of the atmosphere to the ocean surface. This is likely to be important since the CBL top lies at the base of the region where the OLR is quite sensitive to moisture perturbations (Fig. 1). Changes in the atmospheric specific humidity will also alter the tropospheric radiative flux divergence, which drives the prevailing subsiding motion in the Tropics. The changes in subsidence will in turn alter the relative and specific humidity distributions.

Figure 1b shows the change in radiative flux divergence of the lower 300 mb of the atmosphere, taken to be representative of the CBL, in response to a reduction of 10% in the RH of 100-mb layers centered at the latitude and pressure shown. Drying immediately above the CBL increases the radiative cooling of the CBL. Since the CBL finds an equilibrium depth where the radiative cooling balances the warming due to entrainment across the trade inversion (with the surface sensible heat flux being much smaller) the CBL will deepen (Betts and Ridgway 1989). This increases the depth of the atmosphere that is convectively coupled to the ocean surface and therefore moist. Drying at the tropopause has little effect on the CBL flux divergence but, by allowing more upwelling radiation from lower in the atmosphere to escape to space, increases the radiative flux divergence of the entire troposphere. This requires an increase in the balancing subsidence, which allows the CBL depth to fall. Hence drying at different levels in the free troposphere causes quite different changes in the vertical structure of relative and specific humidity.

Here we present a new conceptual framework within which to examine the sensitivity of tropical SST to variations in water vapor. It includes reasonably sophisticated treatments of radiation and boundary layer convection and allows for full vertical coupling via radiative, convective, and dynamical processes, between the water vapor content at one level and that at all other levels (Betts and Ridgway 1989). On the other hand, the atmosphere and ocean dynamics are greatly simplified; there are only two regions as in Pierrehumbert (1995). This model is designed to examine the importance of climate feedbacks involving CBL processes (Clement and Seager 1999). We will impose perturbations in model parameters that introduce *tendencies* to reduce the RH at specific atmospheric levels. In nature these perturbations might occur as a result of, for example, increases in the rain out of condensate in convective towers, increases in the strength of the Hadley and/or Walker circulations, a reduction in the reevaporation of hydrometeors, or a change in lateral mixing. We then examine how these tendencies lead to a new climatic equilibrium in the model within the context of full coupling between the specific humidity distribution at all levels, the radiation, the sea surface temperature, the surface fluxes, and the atmospheric and oceanic transports of heat, moisture, and mass. Since it has been suggested that surface wind speeds were greater during the last glacial maximum (e.g., Bush and Philander 1998), we will also examine the dependence of tropical SST on wind speed.

2. The two-box model of the tropical climate

The model is a two-box extension of the Betts and Ridgway (1989) model and has been described extensively in Clement and Seager (1999). Pierrehumbert (1995), Miller (1997), and Larson et al. (1999) have presented similar models. Betts and Ridgway (1992) have used a single-box version of the model to address some problems of the ice age tropical climate. Figure 2 shows a diagram of the two-box model. One box represents the ascending region of the Hadley cell (the warm pool), where deep convection occurs over warm water, and one corresponds to the chronically subsiding regions over cooler water where there is no deep convection (the cold pool). Since deep convective towers occupy a very small fraction of the Tropics, the air is subsiding almost everywhere, in both the warm pool and the cold pool. In both regions there is a shallow CBL that couples the ocean with the lower part of the atmosphere. The model solves for the SST and humidity and temperature profiles in the CBL and free troposphere regions of each box. The atmosphere transports heat and water vapor between the two boxes and the ocean transports heat between the two boxes in a closed circulation. There is a specified heat export to high latitudes. This was chosen so that the model best reproduces the current climate. It is equivalent to a tropically averaged net surface heat flux of 30 W m^{-2}, which is in good agreement with observations. The model equilibrates such that the specified heat export is balanced by the net radiative gain at the top of the atmosphere averaged over the two boxes. The implications of this assumption will be discussed in the conclusions.

As an advance on the previous models we allow the CBL clouds to precipitate according to the parameterization of Albrecht (1993). The precipitation provides a net condensational heating of the CBL that offsets the radiative cooling of the layer and reduces both the mean depth and the spatial variations of CBL depth. Both these changes are in the direction of improving the agreement with observed CBL depths and variations. Details of the entire model are provided in the appendix.

3. Effects of varying model parameters on tropical relative humidity and SST

The model parameters are varied to assess how the climate would change in response to alterations in the operation of deep convection, the atmospheric circulation, reevaporation, lateral mixing, and the degree of convective mixing into the CBL, all of which could alter the RH of the atmosphere. In each of these experiments the heat export from the Tropics to higher latitudes is held fixed. In nature it is internally determined and could change; this is discussed further in the conclusions. Holding the export fixed isolates changes in tropical climate that are induced even in the absence of any perturbation to the net radiation budget at the top of the

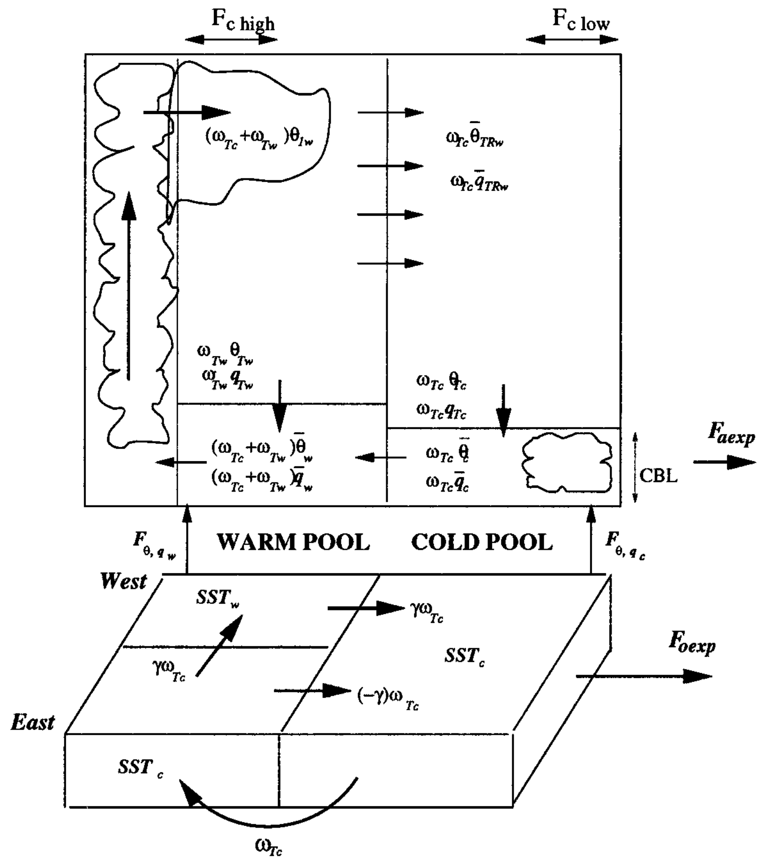

FIG. 2. Schematic representation of the box model. In the atmosphere, air ascends with the equivalent potential temperature of the subcloud layer in the warm pool, $\theta_E = \theta_{MOw}\exp(Lq_{MOw}/c_pT_{\rm sl})$, where $T_{\rm sl}$ is the saturation point temperature of the subcloud air. The air arrives at the tropopause with all of the moisture condensed out and equivalent potential temperature of $\theta_{1w} = \theta_{MOw}\exp(Lq_{MOw}/c_pT_{\rm sl})$. Some air descends locally in the warm pool through the warm pool inversion, ω_{Tw}, by which time it has acquired potential temperature θ_{Tw} and, by reevaporation of hydrometeors, specific humidity q_{Tw}. Some is advected into the cold pool where it descends into the CBL with potential temperature, θ_{Tc}, and specific humidity, q_{Tc}, and is advected back into the warm pool in the subcloud layer. Here θ_{MOc} and q_{MOc} are the potential temperature and specific humidity of the subcloud layer in the cold pool, respectively. The amount of air descending in the cold pool, ω_{Tc}, 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, with surface temperature SST_w, into the subtropics. There it is subducted, with temperature SST_c, and upwells onto the equator. The total overturning mass flux in the ocean is set equal to that in the atmosphere, ω_{Tc}. 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. Here $F_{\theta\rm exp}$ and $F_{q\rm exp}$ are specified exports of heat and moisture to high latitudes in the atmosphere and $F_{o\rm exp}$ is a specified heat export in the ocean. The $F_{\rm chigh}$ is the specified fraction of high cloud cover in the deep Tropics. The $F_{\rm clow}$ is the low-level cloud fraction over the subtropics and is either specified or predicted diagnostically.

atmosphere. We first present results from the version of the model that we believe best represents the workings of the tropical climate system. It includes drizzle and variable cloud albedo. Later we will disable certain feedbacks in order to understand the importance of the large variety of processes that are capable of influencing the sensitivity of tropical SST.

The distribution of free tropospheric humidity is controlled by two parameters that will be varied. The first, $T_{\rm crit}$, provides the above inversion specific humidity, q_T, according to the relation, $q_T = q_s(T_{\rm crit}, \theta_E)$ where the s subscript indicates saturation humidity and θ_E is the equivalent potential temperature of subcloud air in the warm pool that defines the tropospheric saturation moist adiabat, θ_{ES}. To derive q_T we use the relation $\theta_{ES} = \theta \exp(Lq_s/c_pT_{\rm crit})$, where $\theta = T_{\rm crit}(p_0/p)^\kappa$ and q_s is the saturation specific humidity at pressure p and temperature $T_{\rm crit}$. Knowing θ_{ES} and $T_{\rm crit}$, we can invert these relations to solve for the pressure p and then for $q_T = q_s(T_{\rm crit}, p)$. This is used to derive q_T for both regions with, for the standard case, $T_{\rm crit} = 273$ K for the warm pool and 267 K for the cold pool. Lowering $T_{\rm crit}$ reduces the above inversion humidity for a given θ_E. The other parameter is the subsaturation, $\mathcal{P}_{\rm TR} = p_{\rm TR,SL} - p_{\rm TR}$, of the tropopause, which is the difference between the tropopause pressure, $p_{\rm TR}$ and the saturation point pressure, $p_{\rm TR,SL}$ for that air. If $\mathcal{P}_{\rm TR} = 0$ the tropopause is saturated and it dries out as $\mathcal{P}_{\rm TR}$ becomes increasingly negative. In the experiments described $T_{\rm crit}$ and $\mathcal{P}_{\rm TR}$ are changed by equal amounts in each box.

Figure 3a shows two relative humidity (RH) profiles over the warm pool derived from the experiments with the precipitating CBL (the RH over the cold pool is lower). These are areal averages of the clear and cloudy portions of the CBL. The solid line shows the profile for the case that most closely reproduces the real world and corresponds to $T_{\rm crit} = 273$ K in the warm pool, $T_{\rm crit} = 269$ K in the cold pool, and $\mathcal{P}_{\rm TR} = -30$ mb in both boxes. At the surface, RH is about 80% and increases to 100% at cloud base near 960 mb. The RH decreases linearly through the cloud layer, there is a sharp drop at the inversion, and then a smooth decrease above to about 20% at the tropopause. The dotted curve corresponds to a case where $T_{\rm crit}$ was lowered by 4 K in each box. This gives an approximately 10% reduction in free tropospheric RH. The tropopause RH has to stay the same since $\mathcal{P}_{\rm TR}$ is not altered. The CBL RH is almost unchanged by the drying of the above inversion air.

Figure 3b shows RH profiles for three values of $\mathcal{P}_{\rm TR}$: the solid line is the standard case shown in Fig. 3a, the dashed line corresponds to a very moist free troposphere, and the dotted line to a drier free troposphere. Varying $\mathcal{P}_{\rm TR}$ has most effect on the upper-tropospheric humidity, very little effect on the humidity at the inversion, and essentially no effect in the CBL.

The specific humidity of the cloudy portion of the CBL is equal to its saturation value. In the clear-sky portion we assessed the effect of increasing the degree of mixing into the CBL of air from above by increasing the mixing line slope, $\beta_{\rm clr}$. Here $\beta_{\rm clr} = dp_{\rm SL}/dp$ where $p_{\rm SL}$ is the saturation point pressure. As $\beta_{\rm clr}$ becomes larger, $p_{\rm SL}$ between cloud base and cloud top becomes smaller, indicating that the clear-sky portion of the boundary layer is becoming drier. This reduces the relative humidity most near the top of the CBL. Varying the values of these three constants, $T_{\rm crit}$, $\mathcal{P}_{\rm TR}$, and $\beta_{\rm clr}$, therefore, tests the SST sensitivity to quite different per-

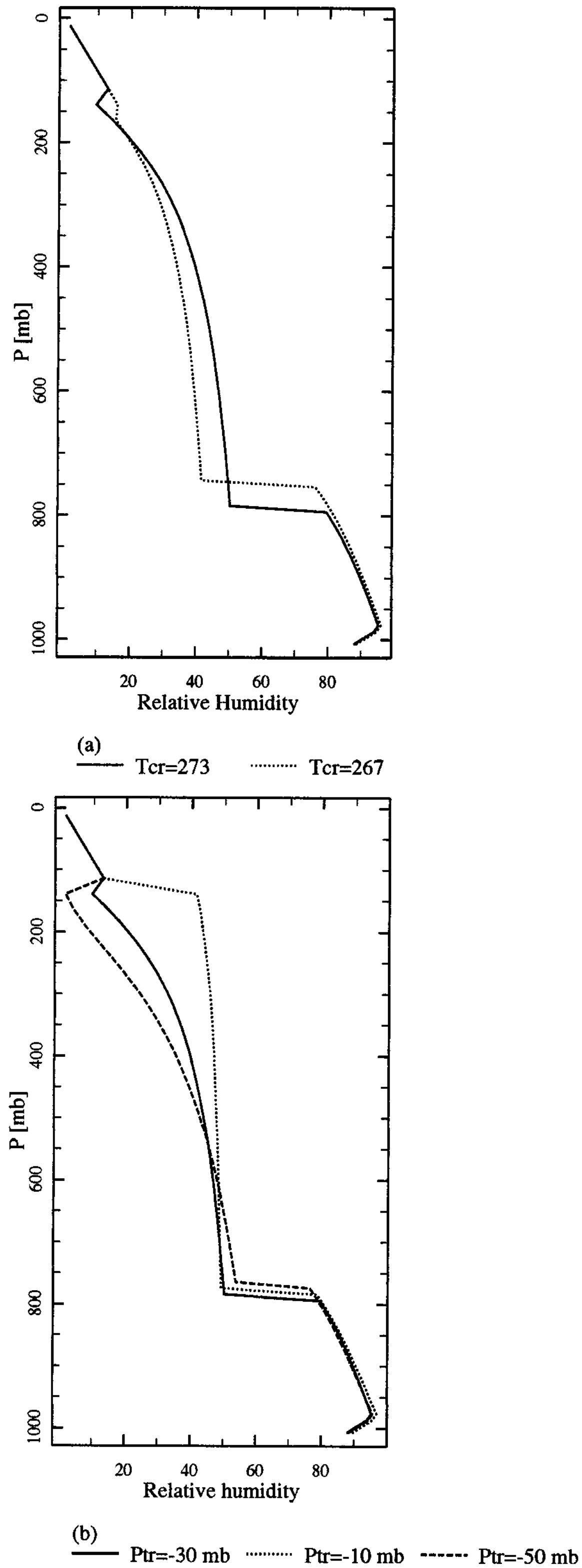

FIG. 3. Vertical profiles of relative humidity over the warm pool for different model solutions. (a) The solid line shows the relative humidity for our standard case, which most closely represents the current climate. The dotted line shows a case where we have adjusted the parameter $T_{\rm crit}$ such that the relative humidity is lower immediately above the CBL. (b) The solid line is the same as in (a), the dashed line shows a case where we assume the tropopause is close to 50% saturation and the dotted curve shows a case where we assume an extremely dry tropopause.

turbations in the atmospheric RH profile. It should be noted that changing these parameters is not the same as imposing a fixed perturbation in the specific humidity. Instead the relative and specific humidities *at all levels* adjust to parameter changes as allowed by the model's coupling of water vapor content, radiation, SST, surface fluxes, and atmospheric and oceanic transports.

The variations in tropospheric RH in these experiments are consistent with observed variability. For example, Fu et al. (1997) use satellite data to show that during warm El Niño events the RH of parts of the subtropical upper troposphere is lowered by about 20%. Sun and Oort (1995), using radiosonde data, document reductions of specific humidity at 500 mb of up to 1 g kg^{-1} in some subtropical regions during the 1982–83 El Niño. This corresponds to a RH reduction of about 20%. Gaffen et al. (1991) documented increases in 850-mb RH at some tropical Pacific island stations of 6% from the mid-1970s to the early 1980s.

The dependence of the tropical mean SST on $\mathcal{P}_{TR}$ and T_{crit} is shown in Fig. 4a and the dependence on β_{clr} is shown in Fig. 4b. The SST is more sensitive to drying immediately above the inversion (smaller T_{crit}) than to drying at the tropopause (more negative $\mathcal{P}_{TR}$). Reducing T_{crit} by 6 K decreases the RH of air above the inversion by only 10%, but cools the SST by over 1.5 K. In contrast, reducing $\mathcal{P}_{TR}$ reduces the tropopause RH from about 40% to less than 10%, but cools the SST by less than 1 K. In Fig. 4b we see that the SST is also more sensitive to drying immediately above the inversion than to drying in the CBL, although the latter is certainly significant.

Although roughly similar to results that consider radiative processes alone, the SST sensitivity in the model arises through a delicate balance of processes in the CBL: drizzle, and the competing effects of greenhouse trapping and cloud reflection. For example, when the lower midtroposphere is dried, the radiative cooling of the CBL increases and the CBL depth increases in order to restore balance. However, the drizzle rate also increases and provides a net condensational heating that largely offsets the increased radiative cooling of the CBL. Thus, the CBL needs to deepen only slightly. Unfortunately, these CBL processes are poorly known. Errors in how they are modeled, or changes in how they operate as climate changes, might radically alter the climate sensitivity to water vapor perturbations. In order to estimate the change in sensitivity we perform experiments in which various feedbacks are eliminated.

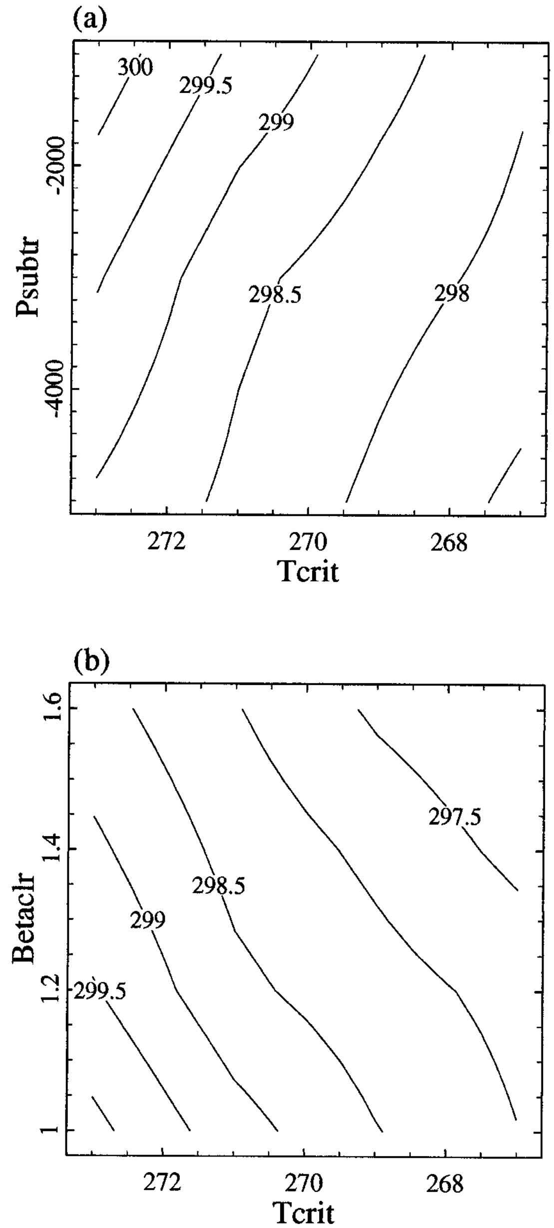

FIG. 4. The tropical mean SST as a function of the above inversion relative humidity (controlled by T_{crit}), the tropopause relative humidity (controlled by $\mathcal{P}_{TR}$), and the CBL humidity (controlled by β_{clr}, for the case of a precipitating CBL with variable cloud albedo.

4. The climatic importance of drizzle in CBL clouds

In this section we do not allow the CBL clouds to precipitate, removing a feedback that powerfully restricts the CBL depth. A deeper CBL increases the depth of the atmosphere that is convectively coupled to the ocean and moist. This increases the greenhouse trapping by water vapor. However, the clouds in the deeper CBL also have an increased liquid water content (LWC) and a higher albedo. In the model the greenhouse effect dominates the albedo effect such that a deeper CBL leads to a warmer climate and a shallower CBL leads to a cooler climate.

The results for the case without drizzle are shown in Fig. 5. By making $\mathcal{P}_{TR}$ increasingly negative the tropopause RH is reduced from 40% to less than 10% and the SST now cools by as much as 2 K. In contrast, reducing T_{crit} by 6 K decreases the above inversion RH

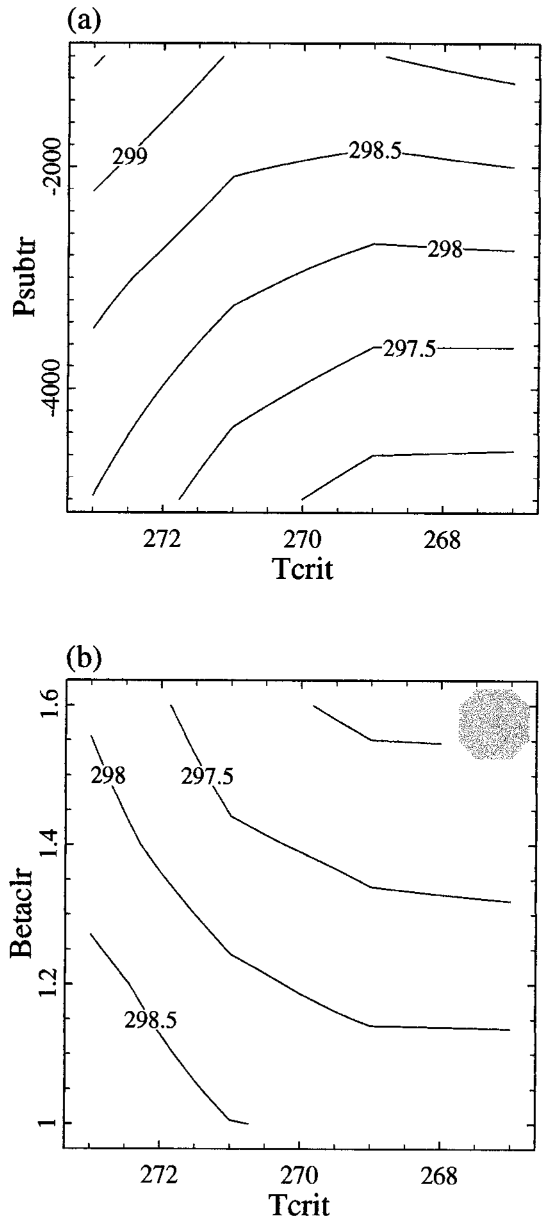

FIG. 5. The tropical mean SST as a function of the above inversion relative humidity (controlled by T_{crit}), the tropopause relative humidity (controlled by $\mathcal{P}_{TR}$), and the CBL humidity (controlled by β_{clr}, for the case of a nonprecipitating CBL with variable cloud albedo. [The gray area in (b) is a part of parameter space where we could not find a solution.]

from about 50% to about 40%, but cools the SST by only 0.5 K. Without drizzle, the CBL deepens as the RH above the inversion is reduced, decreasing the drop in SST as the greenhouse trapping increases. In contrast, the variations in CBL depth amplify the cooling as the tropopause dries. In this case the CBL shallows, and the greenhouse trapping lessens, as the increase in tropospheric radiative flux divergence forces an increase in subsidence. The CBL depth variations act as a negative feedback on lower midtropospheric drying and as a positive feedback on drying at the tropopause. Similarly, the CBL deepens when the mixing of above inversion air into the CBL is increased. We forced increased mixing by increasing the value of the mixing line parameter β_{clr}, which both dries and warms the CBL. This net effect is to increase the radiative cooling of the CBL, which therefore deepens. This provides a net warming that slightly reduces the SST sensitivity to reductions in CBL moisture relative to the case with drizzle.

The total amount of cooling that can be induced by drying at all levels does not depend strongly on the rate at which cloudwater is converted to rainwater. However, drizzle changes the location in the vertical where drying is most effective at cooling the climate. If the CBL clouds did not precipitate, then the changes in CBL depth would guarantee that drying at the tropopause would cause significant cooling of the SST. The fact that CBL clouds do precipitate relocates the maximum sensitivity to drying to the lower midtroposphere.

5. The relative importance of greenhouse trapping and cloud reflection in the CBL

A warmer climate leads to a deeper CBL, which increases the greenhouse trapping within the CBL. It also increases the cloud LWC and, hence, the cloud albedo. The greenhouse trapping is a positive feedback, while the albedo effect is a negative feedback. The climate sensitivity is determined by a balance of the two. Here we examine these competing effects by not allowing the CBL clouds to precipitate and, in addition, holding the solar fluxes fixed in the cloudy portion of the CBL. This experiment isolates the role of the variations in greenhouse trapping by CBL moisture.

Figure 6a shows that the SST warms as T_{crit} is reduced, and the RH above the inversion decreases because the CBL deepens. In this case, where we do not allow the cloud reflection to alter, drying above the inversion can actually cause the climate to warm. On the other hand, as $\mathcal{P}_{TR}$ is made increasingly negative, and the tropopause RH is reduced, the SST is greatly reduced. This is because the CBL shallows as the subsidence increases to balance the increase in the tropospheric radiative flux divergence. A reduction of the tropopause RH cools the tropical SST by as much as 4 K. [This case is most similar to the results of Betts and Ridgway (1989).] Figure 6b shows that the SST cools only very slightly as the CBL dries. In this case the CBL deepens, as discussed in the previous section, and the increased greenhouse trapping almost entirely offsets the cooling tendency introduced by the reduced specific humidity of the CBL.

6. Dependence of tropical SST on surface wind speed

Bush and Philander (1998) have presented a coupled GCM simulation of the last glacial maximum in which

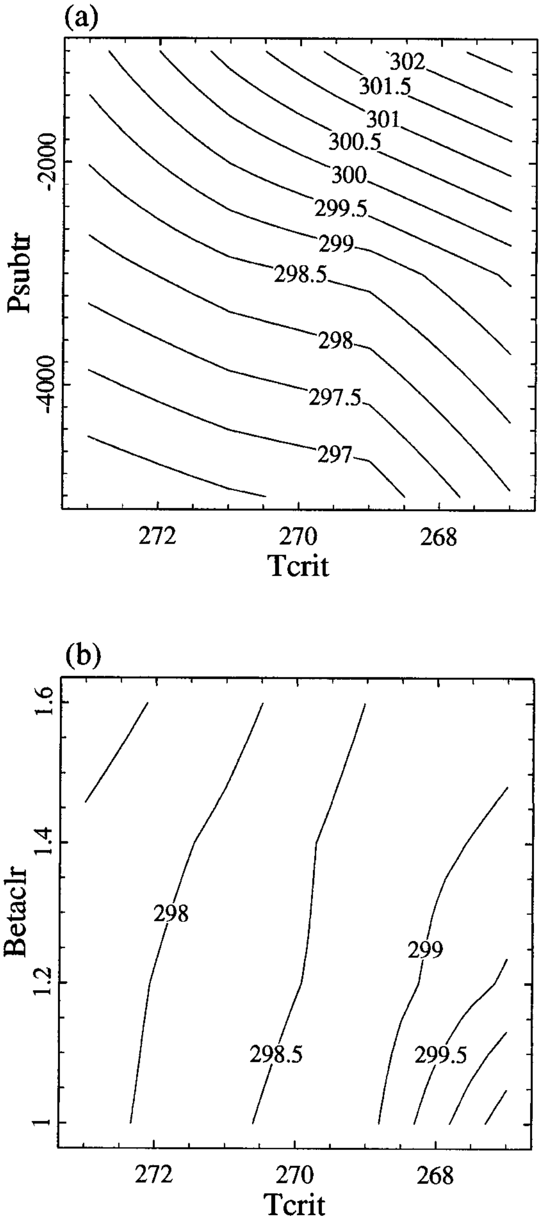

FIG. 6. The tropical mean SST as a function of the above inversion relative humidity (controlled by T_{crit}), the tropopause relative humidity (controlled by $\mathcal{P}_{TR}$), and the CBL humidity (controlled by β_{clr}, for the case of a nonprecipitating CBL with fixed cloud albedo.

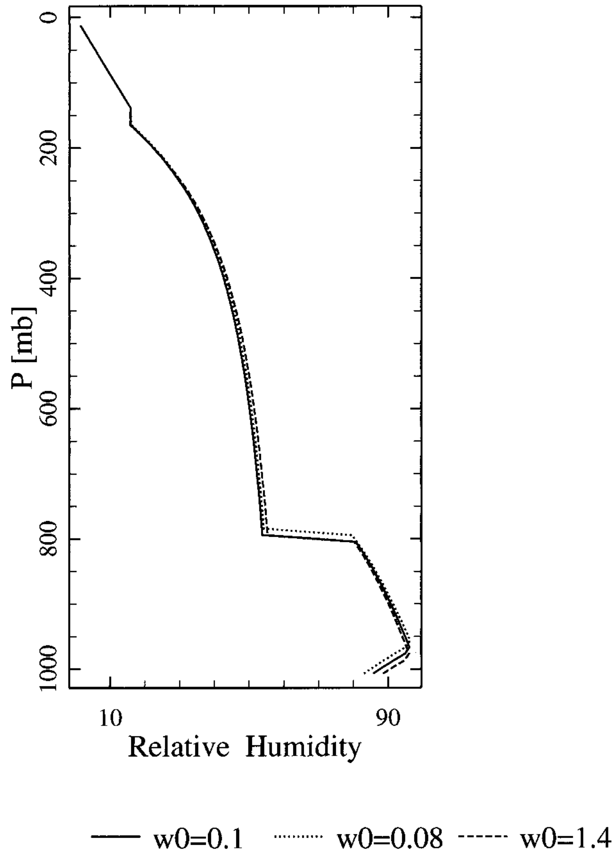

FIG. 7. Vertical profiles of relative humidity over the warm pool for model solutions with different surface wind speeds. The solid line shows the relative humidity for our standard case, which most closely represents the current climate (ω_0 = 0.1). The dotted line shows a case where the wind speed has been reduced by 20% and the dashed line shows a case where the wind speed is increased by 40%.

the surface wind speed over the tropical Pacific Ocean was dramatically increased. The paleoclimatic record of eolian deposition in the deep sea also provides evidence that glacial times might have been windier (Rea 1994). How does an increase in wind speed affect the tropical SST? The surface wind speed is specified in the model and, to look at this problem, we vary the magnitude. It might be expected that changes in surface wind speed would accompany changes in the strength of the general circulation and heat export to higher latitudes. However, to be consistent with the earlier experiments, we use the same specified heat export as before and isolate the change in tropical climate that occurs as wind speed varies, but the net radiation budget at the top of the atmosphere remains unchanged.

The RH profiles over the warm pool for model solutions with three different wind speeds are shown in Fig. 7. Variations in surface wind speed have almost no impact on the RH except in the shallow subcloud layer. In our best guess for the current climate the surface RH is just above 80%, which is close to observed. As the wind speed increases by 40%, the surface RH increases to about 90% and the cloud base drops. The change in RH in the cloud layer is small. Reducing the wind speed reduces the surface RH a few percent, but again has little effect above the subcloud layer.

Despite the trivially small variations in RH in these experiments the tropical SST has a strong dependence on the wind speed, as shown in Fig. 8. This figure shows the mean tropical SST as a function of surface wind speed and the warm pool value of T_{crit}, which controls the above inversion humidity. The standard value for the tropopause subsaturation was used. Taking 7 m s^{-1}

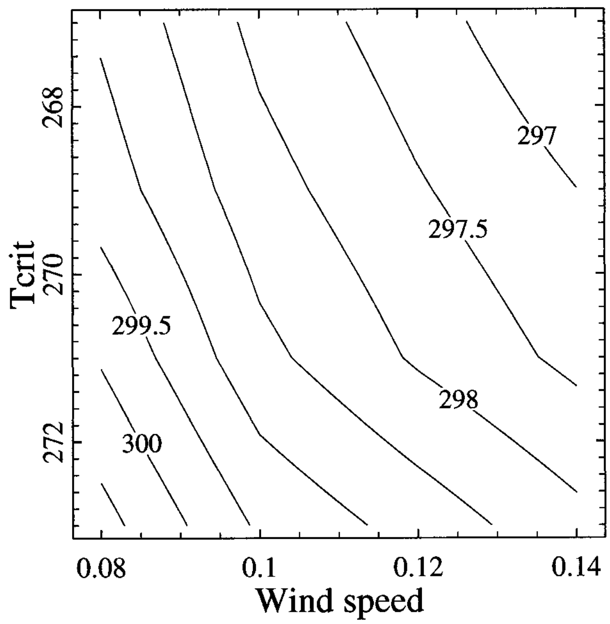

FIG. 8. The mean tropical SST as a function of the surface wind speed parameter, ω_0, and the above inversion relative humidity (controlled by T_{crit} on the vertical axis). Results are not strongly dependent on the tropopause relative humidity and are shown for a value of about 10%.

as a reasonable estimate of the current tropical mean wind speed (giving the surface wind parameter $\omega_0 = \rho g C_D V_0 = 0.1$, where ρ is the surface air density, C_D is a drag coefficient, and V_0 is the surface wind speed), a 40% increase of the wind speed would cool tropical SST by just over 1 K. The amount of cooling is not strongly dependent on either the above inversion or tropopause RH.

As the wind speed increases, the thermodynamic state of the atmosphere moves closer toward that of the ocean surface as the air–sea temperature and humidity differences are reduced. The warmer and moister atmosphere radiates more longwave radiation to space. To restore balance the SST cools. Although the 40% increase of the wind speed cools the SST by more than 1 K the air temperature cools by only 0.6 K as the air–sea temperature and humidity differences are reduced. This is important because features of the climate such as snow lines depend on the air temperature and humidity, not the SST. When the wind speed changes, caution needs to be exercised in translating the resulting SST change into a climate change.

The cooling of the SST is almost entirely accounted for by the increase in the wind speed in the cold pool. Increasing the wind speed in the warm pool alone has little effect on the SST (not shown). As the wind speed increases in the warm pool, but remains fixed in the cold pool, the warm pool moist static energy increases leading to an increase in the static stability over the cold pool. The increased cold pool stability causes the cold pool subsidence to decrease, leading to a moistening of the cold pool CBL, which increases the cold pool SST. In this case the cold pool warms as the warm pool cools and the tropical mean SST changes by little.

If only the cold pool wind speed is increased then, as the CBL warms, the static stability of the cold pool is reduced. The subsidence increases, drying the CBL and amplifying the reduction of SST induced by the increased wind speed. According to these results the tropical mean SST would be most sensitive to an increase in the wind speed in the core of the trade winds as might be caused, for example, by increases in the Hadley cell intensity associated with an increased equator-to-pole surface temperature gradient (e.g., Bush and Philander 1998).

Tropical low cloud cover may play an important role in climate change. Tropical low clouds have a significant cooling effect on the climate since their temperatures are warm and their greenhouse effect is small, but they reflect large amounts of solar radiation to space. In our model, increases in cold pool low cloud cover amplify the cooling associated with increased wind speed. The model low cloud cover is proportional to the difference between the potential temperature above the CBL and the SST (Klein and Hartmann 1993). As the air–sea temperature difference is reduced, the atmosphere warms relative to the ocean and the potential temperature difference across the CBL is increased. The low cloud cover increases. This increases the planetary albedo and amplifies the cooling of the tropical climate. In this case a 40% increase of the wind speed cools the SST by 2 K.

7. Discussion

Here we alter the parameters controlling the humidity distribution in combination to see the maximum amount of cooling that drying the atmosphere could produce. We compare this cooling with those induced by increasing the wind speed or reducing the efficiency of ocean heat transport. Clement and Seager (1999) have previously discussed how changes in the efficiency of ocean heat transport affect the mean tropical climate for the case of nonprecipitating clouds. The results in the presence of precipitating CBL clouds are very similar and will be briefly summarized. As the ocean heat transport increases, the mean SST cools. The increased transport reduces the SST gradient between the Tropics and the subtropics, which reduces the static stability over the subtropics. The subtropical CBL deepens, while the warm pool CBL shallows slightly. (This is quite different from the response to free tropospheric drying, which causes the CBL to deepen in both regions.) When the CBL is shallow and the LWC is small, which is the case for the cold pool, the cloud albedo increases strongly with the LWC (Stephens 1978). When the efficiency of ocean heat transport increases, the increased albedo in the cold pool combines with the reduced greenhouse trapping in the warm pool to cool the tropical climate.

TABLE 1. The tropical cooling induced by various perturbations in the box model. The vertical columns show the cooling of the warm pool (ΔSST_1), the cold pool (ΔSST_2), and the tropical mean (ΔSST). The lines list the perturbations. The first three rows correspond to drying at the tropopause, above the inversion, and in the CBL. The fourth line shows the cooling induced when the entire depth of the troposphere is dried. For comparison, the last two lines show the cooling induced by an increase in wind speed and for a reduction in the efficiency of ocean heat transport. The upper section of the table presents results for fixed cold pool cloud fraction and the lower portion for the case of variable cloud fraction. Control climate: $P_{TR} = -30$ mb, $T_{crit,1} = 273$, $T_{crit,2} = 269$, $\beta_{clr} = 1.2$, $\omega_0 = 0.1$, $\gamma = 0.3$.

	ΔSST_1	ΔSST_2	ΔSST
Fixed cloud cover case			
$\mathcal{P}_{TR} = -50$ mb	0.44	0.30	0.37
$T_{crit,1} = 269$ K, $T_{crit,2} = 265$ K	0.9	1.4	1.2
$\beta_{clr} = 1.6$	0.47	0.33	0.4
Combined H_2O	2.1	2.2	2.15
$\omega_0 = 0.14$	1.0	1.2	1.1
$\gamma = 0.1$	−1.2	−0.2	−0.7
Variable cloud cover case			
$\mathcal{P}_{TR} = -50$ mb	0.1	0.0	0.05
$T_{crit,1} = 269$ K, $T_{crit,2} = 265$ K	0.8	1.2	1.0
$\beta_{clr} = 1.6$	0.2	0.08	0.14
Combined H_2O	1.29	1.46	1.37
$\omega_0 = 0.14$	1.9	2.1	2.0
$\gamma = 0.1$	−0.22	1.3	0.54

The parameter settings that give the best representation of the current climate (see Table 1 for the values) give a warm pool SST of 300 K, a cold pool SST of 297 K, and a mean tropical SST of 298.5 K.

Table 1 shows the reduction of SST as various parameters are changed from these base values. The lines show the temperature reductions when the indicated perturbation is applied in isolation. The line marked ‘‘combined H_2O’’ gives the SST reductions when all the water vapor perturbations in the lines above are applied simultaneously. Because of nonlinear interactions the individual SST reductions do not add up to the total reduction. However, the agreement is sufficient to indicate the relative contributions of the different perturbations. We also show the coolings derived by increasing the wind speed and by reducing the ocean heat transport.

If we alter the parameters $\mathcal{P}_{TR}$, T_{crit}, and β_{clr} so that the RH of the entire vertical column is reduced by about 10%, then the tropical SST is cooled by 2.15 K. By comparison, a 40% increase in the wind speed cools the tropical SST by 1.1 K, which is also substantial. The climatic importance of increases in wind speed might be limited by the fact that the associated changes in SST are far larger than the changes in air temperature or humidity. In the case of fixed cloud cover, reducing the efficiency of ocean heat transport *warms* the tropical SST by 0.7 K.

The lower part of the table shows the results when the cloud cover over the cold pool is allowed to vary according to the stability dependent parameterization of Klein and Hartmann (1993). Miller (1997) and Clement and Seager (1999) have pointed out that variations in low cloud cover are a negative feedback. In agreement, the changes in tropical SST induced by tropospheric drying are much smaller than in the case of fixed cloud cover. Reducing the efficiency of ocean heat transport cools the tropical SST by a modest 0.54 K as the cloud cover increases in response to the increase in the tropical SST gradient and the static stability in the cold pool. In contrast, variable cloud cover actually increases the sensitivity to changes in wind speed with a 40% increase in wind speed leading to a 2-K cooling of tropical mean SST.

Some change in the operation of deep convection and, more generally, the tropical climate system that is capable of drying the lower levels of the free troposphere appears to be the most effective way to cool the Tropics. If such a change could occur in tandem with an increase in surface wind speed, the total SST drop would be quite large and begin to approach the larger estimates of cooling during the last glacial maximum, even in the absence of amplifying feedbacks involving high-latitude processes. Broecker (1997) argues that the RH of the tropical free troposphere during the last glacial was about half of its current value of about 65%. A reduction of that amount would appear sufficient to explain the observed glacial cooling in the Tropics. Changes in low cloud cover have an ambiguous role. In general they will act as a negative feedback reducing the cooling. The exception is that reduced low cloud cover amplifies the cooling induced by increased surface wind speeds. This suggests another possible mechanism for glacial cooling that does not directly involve reductions in tropospheric RH.

8. Conclusions

We have examined the extent to which reductions in tropospheric water vapor content can cool the tropical climate in order to assess whether this is a feasible mechanism for driving global climate change as suggested by Broecker (1997). We use a two-box model of the tropical atmosphere and ocean that incorporates reasonably sophisticated treatments of radiation and the convecting boundary layer that couples the lower levels of the atmosphere to the ocean. The dynamics of the atmosphere and ocean are simplified, but the atmospheric and oceanic heat transports vary as required to ensure energy balance at the top of the atmosphere. As an advance on previous work, we allow the CBL to precipitate as its liquid water content increases.

We impose model parameter changes that introduce tendencies toward lower relative humidity at different levels. The model then finds the climatic state for which energy balance is achieved. The new climate involves changes in the atmospheric and oceanic circulations and heat transports, the surface fluxes, the depth of the CBL, etc. The water vapor content of the atmosphere adjusts at all levels in response to the perturbation. This is an advance over previous approaches to similar problems, which either reason entirely on radiative grounds (e.g., Lindzen 1990) or use idealized, single-column models that greatly restrict the manner in which the atmospheric water vapor content adjusts to perturbations (e.g., Shine and Sinha 1991).

The amount of cooling that occurs in response to a drying depends strongly on the location of the drying and on the CBL physics, in particular the transformation of cloud water into rainwater and precipitation. Precipitation from CBL clouds provides a net condensational heating of the CBL that offsets the radiative cooling of the layer and limits variations in the depth of the CBL. When we take the realistic step of allowing the CBL clouds to precipitate we find that the tropical SST is most sensitive to drying immediately above the inversion. This agrees with the expectations based on radiative considerations alone, but in fact it arises through a delicate balance of CBL processes.

In the case of no precipitation the SST cools most in response to drying at the tropopause. This is because drying at the tropopause increases the outgoing longwave radiation and the tropospheric radiative flux divergence. This increases the subsidence rate and drives the CBL to lower levels. The shallower CBL reduces the greenhouse trapping and amplifies the cooling effect of the initial drying. In contrast, drying immediately above the inversion increases the radiative cooling of the CBL, which, therefore, deepens and offsets the cooling induced by the initial drying. The CBL depth feedback increases the SST sensitivity to drying at the tropopause and decreases the sensitivity to drying immediately above the inversion.

We believe that it is more realistic to allow the CBL clouds to drizzle, in which case feedbacks involving CBL depth are weak. However, the processes that control the conversion of cloud water to rainwater are not fully understood and may alter as climate changes. How the cloud liquid water content and albedo are determined is also not well resolved. Therefore it would be premature to rule out a more significant role for feedbacks involving CBL depth in climate change.

If the cloud fraction is held fixed, and drizzle is allowed, then reducing the relative humidity of the entire atmospheric column by between 10% and 20% can cool the tropical SST by over 2 K. If the cloud fraction varies then the relative humidity reduction cools the SST by less than 1.5 K. These modeled reductions of tropical SST, for the case of quite modest reductions in tropospheric water vapor, correspond to the warmer end of tropical SST reconstructions. In combination with reductions of CO_2, and the tropical impacts of increases in high-latitude snow and ice, the more extreme observational estimates of tropical SST cooling could be reconciled. On the other hand, the halving of tropospheric water vapor that Broecker (1997) argues is indicated by the $\delta^{18}O$ record of Andean glaciers would likely be enough to explain even the most extreme estimated tropical cooling in the absence of other factors.

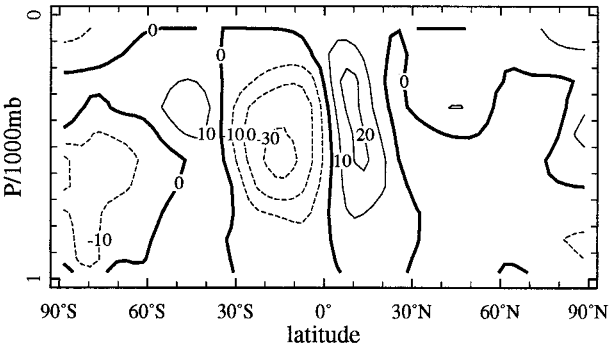

FIG. 9. The difference in relative humidity (%) between two simulations of an atmospheric general circulation model forced by zonally symmetric surface temperatures. The difference is shown for the case with a surface temperature maximum at 10°N minus the case with a maximum on the equator.

By comparison, an increase in the wind speed by 40% can cool the tropical SST by just over 1 K, if the cloud cover is held fixed, and by as much as 2 K, if the cloud cover is allowed to vary. Since there is observational evidence for increased wind speed during the last glacial (Rea 1994) we believe that this could be a means whereby tropical climate was cooled. The cooling might have been quite large, on the order of 2 K, if our hypothesized interaction between increased wind speed and increased low cloud cover is correct. The stronger pole-to-equator temperature gradient during the last glacial might be expected to cause increases in surface wind speed (Bush and Philander 1998), but it is also possible that changes internal to the Tropics, in the form of increases in the strength of the Hadley and/or Walker circulations, could also alter surface wind speeds in the Tropics.

Changes in midtropospheric water vapor are potentially potent methods for inducing climate change in the absence of changes at higher latitudes. How might the tropospheric water vapor distribution have been different in the past? Of course there are many microphysical processes that might have reduced the water vapor content in past, colder, climates. However, we think one likely cause is a change in the atmosphere and ocean circulation that alters the strength and location of major regions of subsidence.

The influence the circulation can have on relative humidity is illustrated in Fig. 9. This shows the difference in relative humidity between two, fully three-dimensional, atmospheric GCM simulations forced by zonally symmetric surface temperatures, one with a maximum at 10°N and the other at the equator. The maximum value of the SST is the same for the two experiments. This calculation was performed using the spectral dynamical core of the Geophysical Fluid Dynamics Laboratory atmospheric model and the same radiation schemes that we use in the box model. The deep and shallow convection schemes follow from Betts and Miller (1993) and contain essentially the same physics as the convective parameterizations in the box model.

The vertical diffusion of humidity is set at a trivially small magnitude. The relative humidity change within the boundary layer is small because of the convective coupling to the ocean. However, in the case where the surface temperature maximum is at 10°N, the southern Hemisphere Hadley cell strengthens (Lindzen and Hou 1988; Lindzen and Pan 1994), which greatly reduces the relative humidity in a broad band of the equatorial and southern Tropics above the boundary layer. Changes in relative humidity of this magnitude, associated with changes in the atmospheric circulation forced by the changing distribution of incident solar radiation, would appear to be a candidate for causing climate change.

While the results presented here suggest a limited role for the direct impact of changes in ocean heat transport they should not be taken to imply a limited role for the ocean in climate change. The idea that changes in atmospheric circulation may be able to induce changes in tropospheric water vapor intrinsically relies on changes in ocean circulation too. This is obvious in the case of El Niño, in which we hypothesize that coupled atmosphere–ocean dynamics, responding to Milankovitch forcing, are capable of altering the characteristics of El Niño and the mean tropical climate (Clement et al. 1999). It is also the case when we consider variations in the Hadley circulation or in the monsoons. In each case we expect that the new atmospheric circulation will be achieved through coupled dynamical processes that involve an active role for the ocean.

While the model used here is open to criticism on many grounds, the assumption that the heat transport out of the Tropics remains fixed is the most troubling. That was the unanimous opinion of the reviewers. It would be possible to allow the heat transport to vary by making it proportional to the pole-to-equator temperature gradient and using a fixed high-latitude temperature. However, this would simply mean that changes in the heat export from the Tropics would damp changes in the tropical climate. In reality changes in heat export to high latitudes could be expected to excite an additional positive feedback involving the snow and ice cover in mid- and high latitudes. This would at least reduce the damping of tropical climate change that reactive changes in heat export would otherwise cause and, if the albedo feedback was strong enough, might conceivably amplify climate changes induced by tropical processes. We are unsure whether the assumption of fixed heat export means our estimates of tropical climate sensitivity are over- or underestimates. For now it is useful to examine how tropical climate changes subject to the constraint that heat export, and therefore the radiation budget at the top of the atmosphere, does not change. Some of the ways that heat transport by the atmosphere and ocean can play an important role in climate change have been examined by Clement and Seager (1999). They show that the total atmosphere plus ocean poleward heat transport can be quite insensitive to radical climate perturbations and explain this in terms of the tight control that the water vapor greenhouse effect exerts on the top of the atmosphere radiation budget. Examining how tropical climate changes influence changes at higher latitudes, and how that then feeds back onto the tropical climate, will be the subject of future work using atmospheric GCMs.

Coupled GCMs used in simulations of past climate are only now beginning to incorporate sufficient resolution in their equatorial ocean components to resolve the important coupled dynamics (Bush and Philander 1998). In addition, atmospheric GCMs typically show too strong a correlation between free tropospheric and surface humidity variations (Sun and Held 1996). These model limitations suggest that, in the past, the ways in which the tropical climate system might be capable, via the water vapor greenhouse effect, of providing a significant response to orbital forcing, has been overlooked. Further investigations of interrelations between coupled dynamics and the greenhouse effect of water vapor using coupled GCMs are greatly needed.

Acknowledgments. It is a pleasure to thank Ron Miller for the numerous hours of good advice. His comments on an earlier, rejected manuscript were indispensable. Alex Hall's careful reading also lead to great improvements over earlier versions. The idea of including drizzle in the CBL model came from a discussion with Dennis Hartmann and Chris Bretherton. The lead author wishes to thank them and the University of Washington for their hospitality during a recent visit. Additional conversations with Kristin Larsen and Steve Klein are gratefully acknowledged. The comments of two anonymous reviewers and Ray Pierrehumbert led to a much improved paper. RS and ACC were supported by NASA Grant NAGW-916 and NOAA Grant UCSIO-10775411D/NA47GP0188. MAC was supported by NSF Grant ATM-94-24915.

APPENDIX

Description of the Two-Box Model

Here we provide details of the two-box model. In the CBL of the two regions there is a balance between warming by the surface sensible heat flux, entrainment of warm air from above, radiative cooling, and the atmospheric heat transport. There is also a balance between surface evaporation, entrainment of dry air from above, and atmospheric moisture transport. The CBL depth is found where these terms come into balance and is also the level of neutral buoyancy for air lifted moist adiabatically from near the surface. The CBL is divided into a well-mixed subcloud layer, which extends from the surface to the lifting condensation level for near surface air, and a cloudy layer above which is bounded by an inversion at the CBL top. Profiles of humidity and temperature within the cloud layer are derived from a mixing line joining the lifting condensation levels (''saturation points'') of subcloud and above inversion air (Betts 1982). The mixing line also provides the cloud liquid water content. The degree of mixing in the CBL is controlled by the mixing line parameter, $\beta = dp_{SL}/dp$, where p_{SL} is the saturation point pressure. Here $\beta = 0$ describes a well-mixed layer. Increasing values of β indicate increased mixing of air from above the inversion into the CBL, leading to an increase in potential temperature and decrease in specific humidity, through the cloud layer. The CBL is divided horizontally into a cloudy portion and a clear-sky portion that are assumed to be mixed in different degrees, as in the earlier papers. The cloud fraction of the cold pool is either predicted from the low-level static stability, as suggested by Klein and Hartmann (1993), or held at a fixed value. There are no low-level clouds in the warm pool.

In the free troposphere above the inversion there is a balance between radiative cooling, horizontal advection of energy by the circulation, warming due to subsidence, and cooling by reevaporation of hydrometoers. The circulation and reevaporation are the sources for the humidity above the inversion. The warm pool free troposphere contains a cloud in its upper half that represents the anvil clouds associated with deep convection. The anvil cloud cover is specified and is modeled such that its net radiative effect at the top of the atmosphere is zero as observed (Pierrehumbert 1995).

The temperature profile in the free troposphere of the warm pool follows a moist adiabat through the equivalent potential temperature of subcloud air, which represents the ability of deep convection to couple the entire troposphere to the SST of the warm pool. In the Tropics the requirement for angular momentum conservation in the presence of small values of the Coriolis parameter prevents substantial horizontal temperature gradients away from the surface (Held and Hou 1980). We build this into the model by assuming the free tropospheric temperature profile of the cold pool is equal to that in the warm pool. The tropopause height is found where the warm pool moist adiabat crosses the uniform stratosphere temperature of 195 K. The humidity immediately above the inversion is equal to the saturation humidity on the warm pool moist adiabat where it crosses a critical temperature. The critical temperatures are chosen so that the warm pool free troposphere is modestly wetter, as expected given the proximity of deep convection, than the cold pool. The degree of subsaturation at the tropopause is specified in the model, and the humidity between there and the inversion varies smoothly.

The temperature, humidity, and cloud liquid water profiles are used to compute the radiative cooling of the atmosphere. The energy balance of the entire tropospheric column is used to derive the subsidence rate, which couples the circulation to the radiative cooling. Atmospheric transports of sensible and latent heat are computed using the modeled subsidence rates. The model includes reasonably detailed radiative transfer schemes for both the solar and longwave spectrum (Ramaswamy and Freidenreich 1992; Ramanthan and Downey 1986; Kiehl and Briegleb 1991). Cloud radiative properties are computed from the cloud liquid water content (Stephens 1978; Lacis and Hansen 1974). Water vapor, carbon dioxide, and ozone are the radiatively active gases.

The mass flux between the two ocean boxes is equal to that between the two atmospheric boxes, which would be the case for a zonally symmetric aquaplanet, since the frictional retardation of the atmosphere is also the frictional driving of the ocean. Water subducts in the cold pool. A fraction, γ, is assumed to upwell in the warm pool, where it cools the SST, while the remainder, $1 - \gamma$, is assumed to recirculate in the cold pool. Decreasing γ decreases the efficiency of the ocean heat transport. Varying γ allows investigations into how the degree of zonal asymmetry in the ocean impacts the climate. The SST is computed from a balance of the calculated ocean heat transport and the net surface heat flux. There is a specified export of sensible and latent heat to higher latitudes that is not varied and which is divided equally between the atmosphere and ocean. The specified heat export to higher latitudes is balance by a net radiative gain at the top of the atmosphere. The model is solved iteratively until it achieves a steady solution. The solution is in all cases unique.

In order to include precipitation from CBL clouds we follow Albrecht (1993). The precipitation is proportional to the product of the shallow convective mass flux and the cloud liquid water. The convective mass flux is derived using the sensible heat flux at cloud base. This is assumed to be proportional to the surface sensible heat flux as is common in these models (e.g., Betts and Ridgway 1989) and has some observational support. The cloud-base sensible heat flux can be parameterized in terms of the convective mass flux times the jump in potential temperature across cloud base. Therefore, estimating the temperature jump using the layers immediately above and below cloud base, we can derive the convective mass flux at cloud base. We assume the convective flux remains constant within the cloud layer, which is a reasonable approximation (e.g., Bretherton 1993). At each level the generation of precipitation is proportional to the product of the liquid water at that level and the convective mass flux. The vertical integral of the precipitation rate allows us to compute the net condensational heating of the CBL. The net heating and drying of the CBL by precipitation is applied to the relevant bulk thermodynamic equations for the CBL. The cloud liquid water is lowered by an amount equal to the precipitation rate times a convective timescale, taken to be a few hours.

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