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INTERACTIONS AMONG CLOUD, WATER VAPOR, RADIATION AND LARGE-SCALE CIRCULATION IN THE TROPICAL CLIMATE

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1. INTRODUCTION

Convection, clouds and water vapor have the potential to produce large positive or negative feedbacks in the climate system. Feedbacks associated with the dependence of saturation vapor pressure on temperature are strongest in the tropics, where the surface temperature is already high. These feedback processes are coupled in important ways to the large-scale circulation in the tropics, which is, in turn, influenced by the spatial gradients of sea surface temperature (SST) and the distribution of land and sea.

Climate feedback processes may respond differently to changes in mean SST and changes in the gradients of SST in the tropics. In this study, we will examine the relative roles of mean SST and SST gradients by studying a set of model experiments with imposed SST distributions with sinusoidal variations. Simulations compare reasonably to averaged observations over the Pacific, considering the simplifications applied to the model. Further results can be found in Larson and Hartmann (2002a).

Larson and Hartmann (2002b, hereafter LH1) showed numerical experiments with uniform SSTs had increased high cloud with increasing SST, though the relative humidity profile showed only a weak dependence on SST. The insensitivity of the relative humidity profile to SST was also found in the numerical experiments of Tompkins and Craig (1999). LH1 also showed the sensitivity of the radiative fluxes to SST is similar to the sensitivity inferred from observations.

2. MODEL DESCRIPTION

The essential physical processes of dynamics, cloud microphysics, convection, radiation and moisture advection are included in the NCAR/PSU mesoscale model (MM5) version 2. The MM5 has been modified as in LH1. The Community Climate Model version 3 radiation code has been implemented as well as the shallow cumulus and boundary layer parameterizations based on Grenier and Bretherton (2001), and McCaa and Bretherton (2002).

3. LARGE-SCALE CIRCULATION

The results of several sinusoidal SST gradient simulations with varying ranges and means provide insights about the role of large-scale circulation. Figure 1 shows the SST distribution for five experiments. The

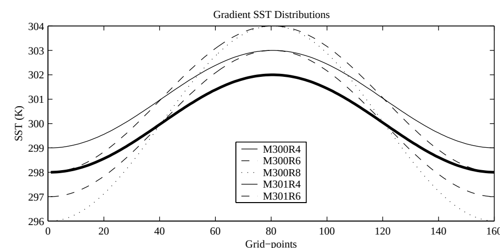

Figure 1: SST distributions for five different simulations.

experiments are named for the mean SST and for the range of SSTs in the distribution. Simulation M300R4 has a mean of 300 K and a range of 4 K. Figure 2 shows the cloud water variables for simulations with a mean of 300 K and increasing SST ranges of 4, 6 and 8 K. A

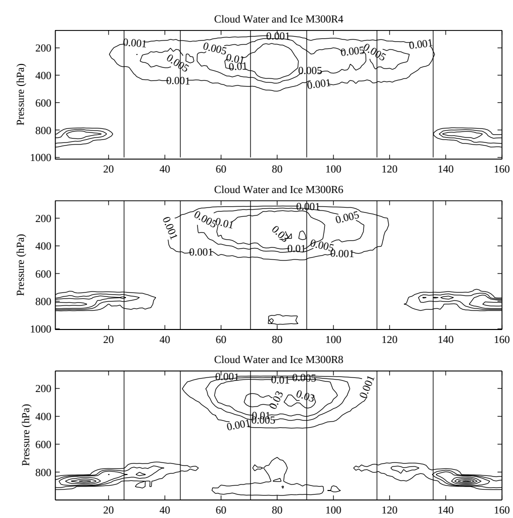

Figure 2: Cloud liquid water, and cloud ice for simulations M300R4, M300R6, M300R8. The contours of cloud ice are labeled in g kg^{-1}. The contours of cloud liquid water are 0.01 g kg^{-1}, 0.03 g kg^{-1}, 0.05 g $kg^{-1}$and every 0.05 g kg^{-1}after that. The contours of cloud ice are 0.001, 0.005, 0.01 and at 0.02 intervals until the maximum.

transition from low stratus clouds in the subsidence region over the coldest SST to high thick ice clouds over

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the warmest SSTs is evident. The width of the cloud ice 0.001 g kg^{-1} contour narrows as the SST range increases and the extent of the liquid cloud water increases, as shown by the 0.01 g kg^{-1} contour. Figure 3 shows the temperature tendencies due to dynamical heating (horizontal and vertical advection), moist convection and condensation (diabatic heating), and radiation for the M300R6 simulation. In Figs. 2 and 3

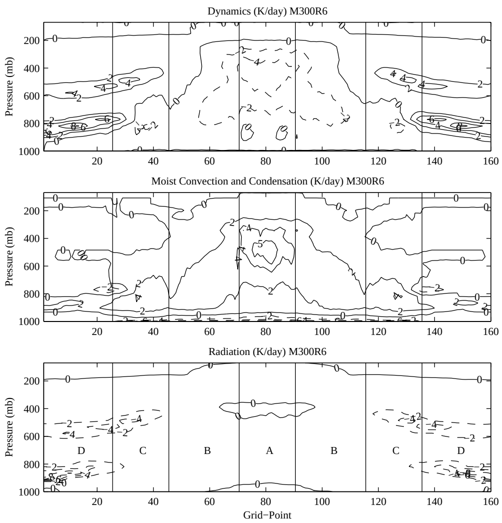

Figure 3: The temperature tendency for the dynamics, moist convection and condensation, and radiation in K day^{-1} for the experiment M300R6 . Negative contours are dashed and positive contours are solid. The contour interval is 2 K day^{-1}.

vertical lines have been drawn to identify regions of subsidence and low clouds, shallow convection, convection and intense deep convection. The boundaries are used to describe regions with different balances between radiation, dynamics and moist convection and condensation.

Figure 4 shows the zonal mass flux, rising motion in the center of the domain over the warmest SST and sinking motion over the coldest SSTs for simulation M300R8. The lower tropospheric circulation becomes

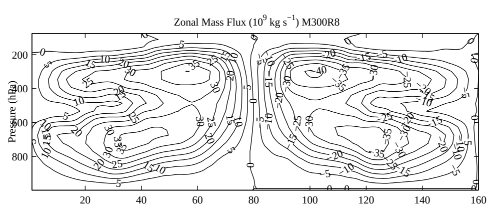

Figure 4: Zonal mass flux averaged over latitude for 30 days for the simulation M300R8.

more pronounced in the simulations with greater SST ranges and creates a double-celled circulation. Interestingly, Grabowski et al. (2000) also found a double circulation in their CRM simulation forced by a sinusoidal temperature gradient. Their double circulation was attributed to the deviation of their quasi-equilibrium temperature profile from an observed tropical temperature sounding. The temperature profile is established by the radiative heating profile, which features a smaller cooling in the upper troposphere than is found in tropical observations when the radiation interacts with the clouds in the CRM. The radiative cooling profile in these simulations is also smaller than observed in the tropical upper troposphere, contributing to the double-celled circulation.

The upper circulation is a response to the heating by deep convection, which is strongest between 300 to 500 hPa in the intense deep convection region. The strength of the heating is proportional to the warmest SST. LH1 found convective heating increased with increasing uniform SST which is consistent with increased heating with increased maximum SST. The maximum SST and the maximum zonal mass flux above the freezing level are linearly related, as shown by five representative experiments in Table 1. For every degree increase in the

Table 1: Zonal Mass Flux: Large-Scale Circulation

Name	Maximum SST (K)	Upper Maximum (10^{12} g s^{-1})	Lower Maximum (10^{12} g s^{-1})
M300R4	302	22.3	23.5
M300R6	303	32.4	33.1
M300R8	304	40.4	38.3
M301R4	303	29.3	15.6
M301R6	304	35.6	30.6

maximum SST the upper tropospheric zonal mass flux maximum increases 25% of its value for M300R6. The strength of the SST gradient has little effect on the upper zonal mass flux maximum.

In the subsidence region of Fig. 3 at about 850 hPa, large negative values of radiative cooling are balanced by subsidence warming, shown by the dynamic temperature tendency. The region of radiative cooling is related to longwave cooling at the top of the stratus clouds, which drives the lower circulation. Nigam (1997) found that lower tropospheric longwave radiative cooling from stratocumulus cloud tops produces a strong dynamical forcing that can be inferred from reanalysis. A maximum of 3-4 K day^{-1} for the radiative cooling is inferred over the east Pacific, which is comparable to values in the M300R4 experiment. The stratus longwave cooling creates a feedback that produces a rapid development of the coastal southerly surface-wind tendency and stratocumulus clouds from March to May along the equatorial South American coast (Nigam 1997). The maxima of radiative cooling above the stratus clouds produces divergence in the boundary layer, since radiation is balanced primarily by subsidence warming and the maximum subsidence occurs in the layer of maximum cooling. As the range of the SST distribution increases, the radiative cooling above the stratus clouds increases and forces an increase in the lower tropospheric zonal mass flux. The lower tropospheric zonal mass flux increases about 12% of its value in M300R6 for every degree increase in the SST range.

For similar SST gradients, when the mean temperature is increased the lower zonal mass flux maximum decreases at a rate of 14% K^{-1} for an SST range of 6 K. A decrease in the strength of the large scale circulation when the mean temperature is increased has also been found in other studies (Knutson and Manabe 1995, Larson et al. 1999). Radiative cooling increases slowly with increasing SST, but the atmospheric stability increases because of the non-linear decrease in the moist adiabatic lapse rate for increasing temperatures. The large decrease in lapse rate and increase in dry static stability forces the strength of the subsidence to decrease in the descending branch of the circulation and causes the large-scale circulation to weaken with increasing SST.

4. RAIN AND CLOUDS

In these simulations with sinusoidal SST gradients, the largest amounts of rain occur over the warmest SSTs, and the smallest amounts occur over the coldest SSTs. The average precipitation rate for five different experiments is given in Table 2. The rain fraction shows the area of the domain where the average precipitation rate is greater than 1 mm day^{-1}. The mean rain rate increases as the SST range increases and as the mean SST increases. The mean rain rate increases 0.18 mm day^{-1} for a one-degree increase in the SST range because the large-scale circulation (upper and lower) increases with the increasing SST range. The increasing circulation increases the upward mass flux which increases the rain rate. The upper tropospheric circulation also increases with increasing mean temperature. The upper circulation is responsible for the increase of 0.06 mm day^{-1} in mean rain rate for a degree increase in the mean SST.

Table 2: Rain Rate and Area, and Low Cloud Cover

Name	Maximum SST (K)	Rain Rate (mm/day)	Rain Area (%)	Low Cloud Cover (%)
M300R4	302	3.21	76.9	27.3
M300R6	303	3.71	70.6	27.5
M300R8	304	3.87	60.6	33.4
M301R4	303	3.24	86.3	22.1
M301R6	304	3.79	70.6	26.7

The rain area fraction of the simulations is defined as the percent of grid-points for which the average precipitation is greater than 1 mm day^{-1}. The rain area fraction can be thought of as the convective region of the model. Areas of smaller rain rates are the non-convective region, and often contain low clouds and subsiding motions. The rain area fraction decreases with the strength of the lower tropospheric large-scale circulation (Table 2). A stronger lower tropospheric large-scale circulation is associated with increased radiative cooling above stratus clouds which corresponds to increased non-convective area and decreased rain area fraction. The low cloud amount is roughly proportional to the lower tropospheric large-scale circulation and inversely proportional to the rain fraction.

5. RADIATION

The average outgoing longwave radiation (OLR) is not sensitive to the SST distribution, in these simulations, and the net absorbed shortwave radiation (SWI) is very sensitive. Changing the SST gradient and mean SST does not strongly affect the average outgoing longwave radiation, while a degree increase in the SST gradient decreases the SWI approximately -5.6 W m^{-2} (Table 3).

Table 3: TOA Radiation Quantities in W m^{-2}

Name	SWI (W m^{-2})	SW CF (W m^{-2})	OLR (W m^{-2})
M300R4	332.7	-59.4	251.3
M300R6	324.7	-67.6	256.7
M300R8	308.4	-84.0	254.0
M301R4	341.5	-50.8	253.3
M301R6	324.0	-68.5	255.3

The average optical depth of clouds in the experiment increases with SST gradient and, generally, the total cloud amount increases with increasing SST gradient. The decrease in SWI is matched by an increase in the lower tropospheric large-scale circulation. An increase in the large-scale circulation corresponds to stronger rising motion which creates more cloud water and higher surface wind speeds, which increase the evaporation at the surface. These effects are consistent with increased water content in the atmosphere, increased shortwave cloud forcing and decreased SWI. The shortwave cloud forcing and the strength of the lower large-scale circulation are both related to the radiative effect of the stratus clouds in the subsidence region.

The domain-averaged OLR is very similar for all the experiments. The cloud top temperature where the visible optical depth reaches 0.1 is approximately constant in all the simulations (Hartmann and Larson 2002b). If the SST gradient is held constant and the mean SST is increased, the longwave cloud forcing increases about 2 W m^{-2} K^{-1} and the clear-sky greenhouse effect increases about 3.5 W m^{-2} K^{-1}. The sum of these two positive feedbacks approximately cancel the longwave emission increase associated with the imposed temperature increase.

6. CONCLUSIONS

This research demonstrates the important effects of the SST gradients and large-scale circulation on tropical climate sensitivities. Simulations with imposed fixed SST gradients are similar to averaged observations over the Pacific, when the discrepancies of large-scale dynamical forcing and constant full period sinusoidal SST gradients are taken into account. The simulations with sinusoidal SST gradient give distinct upper and lower zonal mean circulations in the troposphere. The upper circulation is sensitive to the heating from deep convection over the warmest SST and the lower circulation is sensitive to radiative cooling produced by stratus clouds.

Stronger SST gradients are associated with a stronger lower and upper tropospheric large-scale circulation, a smaller extent of rain area, and larger area coverage by low clouds.

Increasing the mean SST decreases the strength of the lower tropospheric large-scale circulation, which increases the convective area and decreases the subsidence area of the simulation. In a coupled GCM with doubled carbon dioxide, Knutson and Manabe (1995) found an increase in Pacific SST. In the warm pool region they found precipitation enhanced by 15%, but a decrease in the strength of the ascending vertical motion. Those results were explained as a consequence of the moist adiabatic lapse rate decrease with increasing SST, which is also consistent with these fixed SST gradient experiments.

The absorbed shortwave radiation is found to be extremely sensitive to the SST gradient. The stronger lower tropospheric large-scale circulation produces a higher water content in the high and low clouds, increasing the absolute magnitude of the shortwave cloud forcing. The shortwave effect is larger than the small changes in the OLR, and leads to a decrease in net TOA radiation with increasing SST gradient.

Increasing the mean SST creates a positive feedback in these simulations because of the decrease in the lower tropospheric large-scale circulation and the simultaneous decrease in cloud optical depth. As the mean SST increases, low clouds cover a smaller area in the smaller subsidence region, so that the net negative cloud forcing of the low clouds is reduced, producing a net positive feedback. The increased SST decreases the high cloud OLR value and increases the water vapor amount, both of which are positive feedbacks. These effects are offset by the increase in longwave emission with increasing temperatures, but the positive feedbacks are larger.

The atmospheric large-scale circulation is responsible for the organization of tropical convection, clouds, rain, and relative humidity. The re-organization of those quantities can have strong feedbacks on the tropical climate. Increasing the mean SST decreases the strength of the lower large-scale circulation, while increasing the spatial gradients of SST produces increases in the large-scale circulation. Increasing strength of the circulation produces a reduction in the radiation balance in the current model, primarily through increased cloud shortwave forcing. The large-scale circulation causes the greatest change in the net absorbed shortwave radiation, a 25% increase in the lower maximum zonal mass flux of M300R4 leads to a 7.4 W m^{-2} decrease in the net absorbed shortwave radiation. A 0.6 K increase in the SST gradient may be able to offset a 1 degree increase in the mean SST. The response of SST gradients and the accompanying changes in large-scale circulation can be responsible for significant climate feedbacks. The effect of the large scale circulation on the cloud optical depth is very large according to these simulations and may play an important role in tropical climate sensitivities.

7. ACKNOWLEDGMENTS

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