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On the nature of the atmospheric cloud radiative effect and its impact on tropical convection

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Abstract

On the nature of the atmospheric cloud radiative effect and its impact on tropical convection

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Tropical high clouds have been shown to converge radiant energy within the atmosphere. We term this phenomenon the Atmospheric Cloud Radiative Effect (ACRE). The addition of energy unequally in space and time has profound effects on tropical convection. In this thesis, we show that the extra heating by clouds serves to enhance the divergence of energy transport out of the atmospheric column. Additionally, the water vapor lofted into the upper troposphere by convection behaves in a similar fashion to the clouds, heating the atmosphere and enhancing energy export. This atmospheric moisture radiative effect accounts for as much as a fifth of the total radiative heating owing to convection. Principal component analysis of satellite-retrieved cloud data reveal offsetting changes in cloud amount, cloud optical thickness, and cloud top height that give rise to an insensitivity in the top-of-atmosphere cloud radiative effect to changes in sea surface temperature. While increasing vertical motion makes the cloud radiative effect more negative, that decrease is offset by reductions in outgoing longwave radiation owing to increases in water vapor.

The absorption of radiant energy by the clouds warms the upper tropical troposphere compared to simulations where ACRE is artificially removed. We show this increase in stability requires greater surface moist static energy to initiate convection, and hence, contracts

the intertropical convergence zone (ITCZ) equatorward where sea surface temperatures are at a maximum. The meridional gradient in ACRE requires greater poleward energy transport. Thus, despite the increase in stability, the mean meridional circulation intensifies to export more energy out of the tropics.

Finally, we show that the increase in stability owing to ACRE reduces cloud cover. ACRE influences the cloud cover in another way, however, and that is through destabilizing the cloud layer directly through absorption of longwave at cloud bottom and emission of longwave at cloud top. This cloud layer destabilization effect enhances the cloud areal coverage, offsetting some of the reduction from the tropospheric stability changes. The destabilization of the cloud layer also warms and thins the clouds, increasing the cloud radiative effect at the top of the atmosphere.

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DEDICATION

To my friends and family.

Chapter 1

INTRODUCTION

The convective regions of the tropics have long been known to be a critical component of the climate system. Trenberth and Solomon (1994) showed that the tropical West Pacific is the maximum of moist static energy (MSE) divergence within the atmosphere. The strong divergence of moist static energy out of the tropical West Pacific is related to the deep convection that persists there. Convection is not only important for its role in directly moving energy through the atmosphere, but also for its disruption of the flow of radiant energy. Clouds and water vapor are the two largest means of altering radiative transfer through the tropical atmosphere, and both are critically tied to convection. Clouds, in particular, have an interesting role in the radiative transfer process owing to counterbalancing effects on shortwave and longwave radiation. For shortwave radiation, cloud drops and ice crystals serve primarily as scatterers, and therefore reflect a large portion of the total solar irradiance back to space. For longwave radiation, however, clouds serve as approximate blackbodies, and thus converge or diverge a great deal of radiant energy within the atmosphere depending on the cloud altitude and thickness. At the top of the atmosphere, it has been shown that the average net shortwave radiation reflected is approximately equal to the amount of longwave radiation absorbed (Ramanathan et al., 1989; Harrison et al., 1990; Kiehl and Ramanathan, 1990; Kiehl, 1994; Cess et al., 2001; Tian and Ramanathan, 2002; Yuan and Hartmann, 2008). Hence the difference in the top-of-atmosphere net radiation owing to clouds, the Cloud Radiative Effect (CRE), is close to zero. While the clouds have little impact on the top of

atmosphere net radiation, they maintain a strong influence on the surface and atmospheric energy budgets, again owing to the different ways shortwave and longwave radiation interact with them (Ramanathan, 1987; Slingo and Slingo, 1988; Ramanathan et al., 1989; Tian et al., 2001; Tian and Ramanathan, 2002). Tropical convective clouds, on average, act as a large heating term for the atmosphere and a large cooling term for the surface (Ramanathan, 1987; Slingo and Slingo, 1988; Ramanathan et al., 1989; Harshvardhan et al., 1990; Sherwood et al., 1994; Tian et al., 2001; Tian and Ramanathan, 2002). In fact, both Zhang and Rossow (1997) and Tian et al. (2001) suggested that the surface cooling and atmospheric heating can be thought of as partitioning of energy from the surface to the atmosphere. The work presented within this thesis focuses on this atmospheric heating, known as the Atmospheric Cloud Radiative Effect (ACRE), and its role in convection.

Cloud radiative heating is a diabatic heating source term within the atmosphere that is spatially heterogeneous, both in the horizontal and vertical. As such, it is capable of invigorating atmospheric motions. Early studies using coarse grid general circulation models have shown the effects of cloud radiative heating within the atmosphere on the tropical circulation (see Ramanathan, 1987; Slingo and Slingo, 1988, 1991; Randall et al., 1989; Sherwood et al., 1994). These studies showed that cloud radiative heating warms the upper tropical troposphere, enhances the mass flux in the mean meridional circulation, strengthens the Hadley and Walker circulations, and increases local precipitation. Theoretical studies such as Stuhlmann and Smith (1988a,b) also suggested cloud radiative heating in the atmosphere will enhance the circulation patterns. Calculations of cloud radiative heating using International Satellite Cloud Climatology Project (ISCCP) data and European Centre for Medium-Range Weather Forecasts (ECMWF) reanalyses have found that cloud heating and cooling patterns reinforce the tropical circulations (Sohn, 1999; Bergman and Hendon, 2000b).

In this thesis, we seek an understanding of how ACRE impacts several aspects of convection. First, we want to know how much of the divergence of energy out of the tropics results

from ACRE. Is the diabatic heating from the cloud-radiation interactions removed locally or remotely? How much of the total divergence of energy can be accounted for by ACRE? Second, we ask how ACRE impacts the location of tropical convection. Does ACRE simply reinforce the location of convection or is it able to change it? Finally, we seek to understand how the clouds themselves respond to ACRE. How sensitive are the important cloud metrics for radiative transfer (cloud amount, cloud top temperature, and cloud optical thickness) to ACRE?

The above questions help us to understand and model the climate system, despite only focusing on the atmospheric component of the cloud radiative effect. The surface cloud radiative effect is also important to the atmospheric cloud radiative effect over regions of tropical deep convection because they are approximately equal in magnitude (e.g. Ramanathan, 1987). Focusing on the atmosphere only component of the cloud radiative effect simplifies the problem. It allows us to more fully understand the physical processes related to cloud heating within the atmosphere. In addition, so long as we continue to rely on atmosphere only experiments, such as the Atmospheric Model Intercomparison Project (AMIP), it is important that we understand how the clouds respond to radiation within those experiments.

In chapter 2, we investigate the relationship between ACRE and the divergence of MSE out of the atmosphere, which we will refer to as Atmospheric Heat Transport (AHT). We focus on the convectively active area of the tropical West Pacific known as the warm pool (defined here as 10 S – 10 N, 150 – 180 E). Much like the clouds, water vapor absorbs longwave radiation. We aim to quantify the effect of changing specific humidity through convection on the absorption of longwave radiation. We term this radiative process the Atmospheric Moisture Radiative Effect (AMRE). The specifics of its derivation are given in chapter 2, but we find that AMRE, on average, acts in the same way as the cloud systems; it is a net heating of the warm pool atmosphere. We find that AMRE is generally positive across the deep tropics (10 S – 10 N), and hence, enhances the effect of ACRE. We refer to the sum

of ACRE and AMRE as the Atmospheric Convective Radiative Effect (ACoRE), and find that it accounts for roughly two thirds of the climatological average AHT in the tropical warm pool. We use a combination of reanalysis data and satellite observations to show that ACoRE and AHT are strongly correlated in time, and increase at a roughly one-to-one rate with convective activity, owing to an insensitivity of surface turbulent fluxes to convection.

Additionally, we show that top-of-atmosphere (TOA) net radiation is insensitive to changes in average vertical velocity, owing to offsets in AMRE and CRE. Water vapor reduces the outgoing longwave radiation (OLR), and the OLR reduction explains most of AMRE. We use Moderate Resolution Imaging Spectroradiometer (MODIS) data (Platnick et al., 2003) to investigate which cloud changes drive the decrease in CRE associated with increasing convective activity. We find that increases in both cloud amount and in the average optical thickness are responsible for the decrease in CRE. Cloud top height (or more specifically, cloud top temperature) is roughly constant on monthly timescales, in agreement with the Fixed Anvil Temperature hypothesis (see Hartmann and Larson, 2002).

In chapter 3, we investigate the role ACRE plays in determining the location of the intertropical convergence zone (ITCZ). The ITCZ is important for precipitation, as well as the distribution and transport of energy and moisture. Climate models have notorious difficulty in accurately representing the ITCZ, with a bias of excess precipitation in the Southern Hemisphere often referred to as the double ITCZ bias (e.g. Mechoso et al., 1995; Lin, 2007; Flato et al., 2013; Li and Xie, 2014).

We employ the Clouds On-Off Climate Intercomparison Experiment (COOKIE) aqua planet simulations. The COOKIE experiments are a pair of simulations: one with cloud-radiative interactions included and one with cloud-radiative interactions removed. Both simulations use fixed sea surface temperatures, such that only the atmospheric component of the cloud-radiation interactions may be examined. By examining the differences in these simulation pairs, we are able to assess the role of ACRE on the location of the ITCZ. We find

that ACRE contracts the ITCZ toward the equator where sea surface temperatures (SSTs) are highest. In some models, this is realized as a switch from a double ITCZ to a single ITCZ, while in other models, the double ITCZ pattern is maintained while still contracting toward lower latitudes.

We show that the contraction of the ITCZ in the presence of ACRE is caused by an increase in atmospheric stability owing to the warming produced by the abundant high clouds in the tropics. The increase in stability requires higher surface moist static energy to initiate convection, thus reducing convection farther from the equator in favor of more intense convection closer to the equator. The mechanism we advocate is that presented by Landu et al. (2014) and is similar to the “upped-ante” mechanism of Neelin et al. (2003).

In addition to the contraction of the ITCZ, we also find an intensification of the mean meridional circulation in the presence of ACRE. This result may seem paradoxical at first since the atmospheric stability also increases. We show, however, that there is a strong meridional gradient in ACRE that requires additional energy transport, despite the increase in stability. The atmosphere therefore adjusts by contracting the ITCZ toward the equator while at the same time intensifying the convection so as to export more energy poleward.

In chapter 4, we use a cloud-resolving model to examine the influence ACRE has on cloud fraction and the top-of-atmosphere cloud radiative effect (CRE). For tropical high clouds, warming at cloud bottom by longwave absorption and cooling at cloud top by longwave emission destabilize the cloud layer (see Ackerman et al., 1988; Chen and Cotton, 1988; Lilly, 1988; Fu et al., 1995; Mace et al., 2006). Fu et al. (1995) showed that this radiative destabilization extended the cloud fraction in a simulated squall line storm system. Durran et al. (2009) and Dinh et al. (2010) have shown a similar radiation-driven cloud spreading effect in tropical tropopause layer cirrus.

Fu et al. (1995) noted that, had they continued their squall line simulation with cloud-radiative heating included, the atmosphere would begin to warm owing to that cloud-

radiative heating. It has been known for some time that the cloud warming can effect the stability of the atmosphere (see, for example Slingo and Slingo, 1988, 1991; Sherwood et al., 1994, as well as chapter 3). The increase in stability can also modify the cloud fraction and CRE, though not necessarily the same way the direct destabilization effect does. We show that these two ACRE-driven effects drive opposing changes in cloud fraction. The direct destabilization of the cloud layer enhances high cloud fraction, while the stabilization of the troposphere reduces high cloud fraction. The relative magnitude of these effects changes depending on the modeling setup employed.

There are several questions we address in chapter 4: (1) how does cloud fraction respond to ACRE? (2) How does the top-of-atmosphere cloud radiative effect respond to ACRE? (3) Does a large-scale circulation mitigate the stability effect of the cloud radiative heating? and (4) What are the implications for cloud feedbacks?

In summary, the following chapters bring new understanding to the role of ACRE on tropical convection. Chapter 2 examines the role ACRE plays in the energy transport, related to the intensity of convection. Chapter 3 examines the role ACRE plays in the location of convection. Chapter 4 examines the role ACRE plays in the cloud properties important for radiative transfer: cloud amount, cloud top temperature, and cloud optical thickness.

Chapter 2

THE RELATIONSHIP BETWEEN ATMOSPHERIC CONVECTIVE RADIATIVE EFFECT AND NET ENERGY TRANSPORT IN THE TROPICAL WARM POOL

2.1 Introduction and background

The warm pool (10 S – 10 N, 150 – 180 E) is an area of active convection. Climatologically, it has the highest sea surface temperatures and strongest incoming top-of-atmosphere (TOA) net radiation. Additionally, the strongest atmospheric heat transport (AHT) out of the tropics occurs in the warm pool (Trenberth and Solomon, 1994). The large-scale circulation ultimately controls the horizontal heat transport by the atmosphere, and convective heating plays an important role in determining the large-scale circulation. Convection is directly responsible for the vertical transport of moist static energy, and is also responsible for the generation of high clouds, whose radiative heating is another important contribution to the large-scale circulation.

Early studies using coarse grid general circulation model studies have shown the effects of cloud radiative heating within the atmosphere on the tropical circulation (see Ramanathan, 1987; Slingo and Slingo, 1988, 1991; Randall et al., 1989; Sherwood et al., 1994). These studies showed that cloud radiative heating warms the upper tropical troposphere, accelerates the jet stream, enhances the mass flux in the mean meridional circulation, strengthens the Hadley and Walker circulations, increases precipitation, and can also excite extra-tropical height anomalies. Theoretical studies such as Stuhlmann and Smith (1988a,b) also suggested cloud radiative heating in the atmosphere will enhance the circulation patterns. Calculations of cloud radiative heating using International Satellite Cloud Climatology Project (ISCCP)

data and European Centre for Medium-Range Weather Forecasts (ECMWF) analyses have found that cloud heating and cooling patterns reinforce the tropical circulations (Sohn, 1999; Bergman and Hendon, 2000b). Zhang and Rossow (1997) and Tian et al. (2001) noted that the heating of the clouds would, with all other things being equal, partition more of the net divergence of energy transport into the atmosphere instead of the ocean.

Clouds perturb the net radiative fluxes at the top of the atmosphere, termed the cloud radiative effect (CRE; see table 2.1 for list of acronyms), as well as the surface (SCRE) and within the atmosphere (ACRE). In the convective regions of the tropics the top-of-atmosphere net CRE is near zero (Ramanathan et al., 1989; Harrison et al., 1990; Kiehl and Ramanathan, 1990; Kiehl, 1994; Cess et al., 2001; Tian and Ramanathan, 2002; Yuan and Hartmann, 2008). The atmospheric and surface cloud radiative effects can be quite large in cloudy regions, however, owing primarily to the different ways shortwave and longwave radiation interact with clouds (Ramanathan, 1987; Slingo and Slingo, 1988; Ramanathan et al., 1989; Tian et al., 2001; Tian and Ramanathan, 2002). ACRE is a large heating term for the atmosphere, especially in regions of persistent high clouds, while SCRE is a large cooling term for the surface (Ramanathan, 1987; Slingo and Slingo, 1988; Ramanathan et al., 1989; Harshvardhan et al., 1990; Sherwood et al., 1994; Tian et al., 2001; Tian and Ramanathan, 2002). In the convective part of the tropics, ACRE and SCRE are largely offsetting, hence the near-zero top-of-atmosphere CRE (Tian et al., 2001). Tian et al. (2001) suggest that the offsetting nature of the ACRE and SCRE allows them to be considered as an indirect energy transport; energy is lost at the surface and gained in the atmosphere. This indirect heat transport, as Tian et al. (2001) refer to it, adds to the direct transport from surface turbulent fluxes and moist convection. The indirect heat transport could also be viewed as an alternative to the surface turbulent heat fluxes since reduced shortwave absorption at the surface is likely to be balanced by a decrease in surface turbulent fluxes. Sherwood et al. (1994) found that removing high cloud radiative heating reduced the Hadley circulation by

about 25% (as measured by 500 hPa vertical velocity) while the Walker circulation completely collapsed. Bergman and Hendon (2000a) suggested that the cloud radiative heating in the atmosphere contributes about 20% to the overall circulation, as measured by zonal mass flux.

The idea that ACRE helps sustain and enhance AHT was further tested by Tian and Ramanathan (2002) and then again by Tian and Ramanathan (2003). Tian and Ramanathan (2002) showed that the mean spatial patterns of ACRE and AHT align well across the tropical Pacific, in terms of both structure and magnitude. Tian and Ramanathan (2003) showed that a model forced only by observed ACRE was capable of developing quantitatively realistic looking Hadley and Walker circulations. In contrast, forcing the model with surface evaporation, created a wildly different circulation pattern compared to the real tropical atmosphere. Sohn (1999) suggested that water vapor convergence may act as a diabatic heating source like the clouds. Additionally, it was shown by Voigt et al. (2014) that shifts in the intertropical convergence zone (ITCZ) in aquaplanet simulations can be attributed to changes in the local cloud radiative heating within the atmosphere.

In this paper, we use the European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis Interim (ERA-Interim) output coupled with Clouds and the Earth’s Radiant Energy System-Energy Balanced and Filled (CERES-EBAF) data to estimate the ACRE, CRE, and horizontal energy divergence out of the tropical warm pool. Additionally, we explore the contributions to the radiative heating from convectively lofted water vapor, since its interactions with longwave radiation are similar to those of the clouds. We refer to this radiative heating anomaly as the Atmospheric Moisture Radiative Effect (AMRE). The combined ACRE and AMRE give us the Atmospheric Convective Radiative Effect (ACoRE), which is the total change in radiative cooling of the atmosphere due to the modification of the atmosphere’s radiative properties by convection. We explore the relationship between ACoRE and energy divergence or atmospheric heat transport (AHT). While our results cannot show a

causal relationship between ACoRE and AHT, we find that the two are closely related within the warm pool region. Our results confirm and elaborate on those of Tian et al. (2001) and Tian and Ramanathan (2002). We show that water vapor increases in convective regions make a substantial contribution to AHT. Our results are also consistent with Zelinka and Hartmann (2012), who showed that feedbacks within the climate system act to enhance the existing gradient in top-of-atmosphere radiative fluxes, and hence, require stronger poleward heat transport. A more recent study by Lee and Yoo (2014) similarly suggests poleward heat transport being enhanced by the clouds. Our results are also consistent with work by Wu et al. (2010) and Hwang and Frierson (2010), in which increases in poleward energy transport in general circulation models are found to directly relate to changes in the gradient of moist static energy (MSE) within these models. Since the ACRE (and ACoRE) is a means of enhancing the local MSE, it is likewise capable of enhancing poleward energy transport.

We further break down the ACoRE and AHT relationships across different thermodynamic and dynamic regimes. It is common to use sea surface temperature and pressure velocity along the 500 hPa surface as proxies (e.g. Bony et al., 1997, 2004; Yuan et al., 2008; Franklin et al., 2013). Additionally, we compare the rest of the tropics and several other regions (the West Pacific, East Pacific, Indian Ocean, and Atlantic Ocean) to the results found in the warm pool. We show that the structural relationships between ACoRE and AHT within the warm pool are common in all convective regions within the deep tropics (10 S – 10 N). Additionally, we make use of Moderate Resolution Imaging Spectroradiometer (MODIS) data (Platnick et al., 2003) to investigate what types of cloud changes are occurring with respect to changes in sea surface temperature or vertical velocity. Principal component analysis of the MODIS data reveal that offsetting changes in cloud amount and cloud optical thickness make the cloud radiative effect insensitive to sea surface temperature in the tropical warm pool. Increases in upward vertical velocity, however, show increases in both cloud amount and cloud optical thickness which decrease CRE.

2.2 *Methods and datasets*

We make use of the European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis Interim (ERA-Interim; Dee et al., 2011) dataset to get the divergence terms as well as the sea surface temperatures (SSTs). The reanalysis is run on a $0.75^\circ \times 0.75^\circ$ grid covering the entire globe, but is available at a variety of resolutions. For this study, we use the $1^\circ \times 1^\circ$ grid to match the $1^\circ \times 1^\circ$ grids of the CERES and MODIS datasets (see below for details). We use the monthly means of daily means for each variable. Haimberger et al. (2001) show that the mass fluxes in ERA-Interim are unreliable and considered spurious, and therefore we remove their effect on energy transports following the methodology of Chiodo and Haimberger (2010) (see additionally, Mayer and Haimberger, 2012). We examine the robustness of our results to these errors in section 2.4 and find that our conclusions remain unchanged.

For the radiative fluxes, we make use of Clouds and the Earth’s Radiant Energy System-Energy Balanced and Filled (CERES-EBAF) data (see Wielicki et al., 1996, 1998, for a description of the CERES instrument and algorithm). We make use of both CERES EBAF_Ed2.8 and CERES EBAF-Surface_Ed2.8 (see Loeb et al., 2009; Kato et al., 2013, respectively). The CERES data are on a $1^\circ \times 1^\circ$ grid. Despite the resolution being the same, the CERES data are on the grid centers (0.5° , 1.5° , etc.) while the ERA-Interim data are on the grid edges (0° , 1° , etc.). We re-grid the CERES data onto the ERA-Interim grid so that the data are co-located. Again, we make use of monthly averages for the top-of-atmosphere and surface fluxes.

Regional error estimates are 4 W/m^2 for TOA SW fluxes, 5 W/m^2 for LW fluxes, 2.6 W/m^2 for clear-sky SW fluxes, and 3.6 W/m^2 for clear-sky LW fluxes (Data Quality Summary). Over the March 2000 – June 2002 period when only Terra was carrying the CERES instruments, the regional error for the TOA SW fluxes was slightly higher at 5 W/m^2 . The surface fluxes for CERES have larger errors associated with the higher uncertainties inherent

to the process of generating surface fluxes from space-borne measuring systems. The errors are up to 11 W/m^2 for upwelling and downwelling SW fluxes, 13 W/m^2 for upwelling LW fluxes, and 12 W/m^2 for downwelling LW fluxes (Data Quality Summary).

To calculate errors for cloud radiative effects, we must combine errors from the individual sources. This is done as the square root of the sum of the squares, assuming the errors are uncorrelated. For example, the error in the CRE is $\sqrt{4^2 + 5^2 + 2.6^2 + 3.6^2} = 8 \text{ W/m}^2$. Similarly, we calculate errors for SWCRE and LWCRE as 5 W/m^2 and 6 W/m^2 , respectively. We are also interested in the atmospheric and surface cloud radiative effect (ACRE and SCRE, respectively). Similar to the TOA CRE calculated above, the error on the SCRE is $\sqrt{11^2 + 11^2 + 13^2 + 12^2} = 24 \text{ W/m}^2$. Likewise, the error on the ACRE is $\sqrt{8^2 + 24^2} = 25 \text{ W/m}^2$. The errors for ACRE are large, similar to those for AHT. However, ACRE is largely dictated by LWCRE (Ramanathan, 1987; Slingo and Slingo, 1988; Ramanathan et al., 1989; Tian et al., 2001). The actual errors in ACRE are likely to follow those of the LWCRE, which are only 6 W/m^2 , and thus we do not expect the large error estimate for ACRE calculated here to alter our conclusions.

We investigate the West Pacific warm pool (10 S – 10 N, 150 – 180 E). We use CERES data from March 2000 up through May 2014. The ERA-Interim dataset runs from 1979 up to the present, but we use just the March 2000 through May 2014 period to be consistent with CERES. For comparison to the warm pool, we also investigate several other regions: the whole tropics (10 S – 10 N, 0 – 360 E); the Atlantic Ocean (10 S – 10 N, 300 – 360 E); the Indian Ocean (10 S – 10 N, 40 – 100 E); and the East Pacific ITCZ (5 N – 15 N, 210 – 260 E). For all regions, we consider only grid boxes that are over ocean. The seasonal cycle is removed from both the ERA-Interim and CERES-EBAF data so that the relationships shown in the results and discussion section are not simply due to the seasonal cycle. We have also performed the same analyses with the seasonal cycle included and find it does not alter our conclusions (not shown).

It is important to consider the effect convection has on the clear-sky radiative cooling profile by changing the vertical distribution of water vapor. We expect that the AHT is related to the net change in radiative heating due to convection, and hence, the combined cloud and water vapor effect. We define the Atmospheric Moisture Radiative Effect (AMRE) to be the difference between the actual clear-sky radiative heating (Rclr) and a baseline “non-convective” clear-sky radiative heating profile. Since the goal is to analyze the effect of convection on the radiation budget, we use vertical velocity as a means of discriminating when convection is active. Therefore, the “non-convective baseline” is calculated as the average radiative cooling for all points where $\omega_{500} > 0$. We select the $\omega_{500} > 0$ cutoff because there are very few clouds and the middle troposphere tends to be dry (see Fig. 2.1). Using a more strict criterion like $\omega_{500} \approx 0$, decreases the baseline cooling by about 1 W/m^2 in the warm pool. The transition from a moist, cloudy atmosphere to a dry, cloud-free is smooth in the warm pool, but the cutoff is more distinct across the whole of the tropics and occurs at $\omega_{500} = 0$ (not shown). Other baselines could have been chosen. While a different baseline would alter the mean value of AMRE, it would not change its behavior.

While the baseline criterion is the same across regions, the radiative cooling value calculated for that baseline is specific to each region. Thus, for analyses over regions outside the warm pool, the baseline is recalculated for each region of interest (the differences between regions are all less than 5 W/m^2). In the warm pool, the baseline Rclr is -132 W/m^2 . The atmospheric moisture radiative effect is reflected in changes in the top of atmosphere clear-sky radiative fluxes (mostly the outgoing longwave radiation) and there is very little sensitivity of the surface clear-sky radiative fluxes to ω_{500} (not shown). We call the combined effect of ACRE and AMRE the Atmospheric Convective Radiative Effect (ACoRE).

We have also investigated the impact of SST on Rclr and its subsequent influence on the baseline for the AMRE calculation. Radiative cooling increases with temperature, assuming constant specific humidity and all other atmospheric constituents. The baseline Rclr used for

AMRE does have a SST dependence; however, the radiative cooling weakens as temperature increases, such that the impact of increasing water vapor far outweighs any direct temperature influence. The underlying SST can, however, control the water vapor via convection, which is a likely factor in the differences in baseline for AMRE between regions.

We use the MODIS histogram of cloud frequency binned by cloud optical thickness and cloud top pressure. These MODIS data are available from July 2002 through the present as monthly averages on a $1^\circ \times 1^\circ$ grid. Like the CERES data, the MODIS data are on grid centers ($0.5^\circ, 1.5^\circ$, etc.); we re-grid the MODIS data onto ERA-Interim grid so that the data are co-located. Because the MODIS data do not cover March 2000 – June 2002, for the principal component analysis section of this paper, we only use data from July 2002 – May 2014 (the overlap period for all three datasets: ERA-Interim, CERES, and MODIS). The results derived from the ERA-Interim and CERES datasets alone are not sensitive to the change in record length.

The terms that are of most interest to us are the atmospheric radiative cooling (both all-sky and clear-sky), the top of atmosphere and surface radiative fluxes, the atmospheric cloud radiative effect, the atmospheric moisture radiative effect, the atmospheric convective radiative effect, the surface turbulent heat fluxes, and the divergence of total energy transport. The atmospheric radiative cooling, TOA and SFC radiative fluxes, and atmospheric cloud radiative effect are all taken directly from the CERES monthly data. The atmospheric moisture radiative effect and atmospheric convective radiative effect are computed as described above, and are calculated using a combination of CERES and ERA-Interim monthly data. The surface turbulent fluxes and divergence of total energy transport are directly taken from the ERA-Interim monthly data.

2.3 Results and discussion

We begin by calculating the average values for the fluxes and heat transport terms for the warm pool region. We present the results in a similar fashion to Fig. 7 of Tian et al. (2001) (see Fig. 2.2). The values in Fig. 2.2 differ from those of Tian et al. (2001) owing to measurement changes, use of reanalysis, and length of record differences. As noted in section 2.2, all of the radiative flux-dependent variables (R_{toa} , R_{clr} , R_{atm} , R_{sfc} , and $ACRE$) are calculated using CERES data, $ACoRE$ and $AMRE$ are calculated using a combination of CERES and ERA-Interim data, while the rest of the values in Fig. 2.2 are from ERA-Interim data. The top of atmosphere net radiative flux (R_{toa}) is roughly 30 W/m^2 less in our climatology than that reported by Tian et al. (2001), while the surface net radiative flux (R_{sfc}) is roughly 15 W/m^2 greater. These combine to make the net radiative cooling roughly 45 W/m^2 stronger in our climatology than that reported by Tian et al. (2001).

To understand the 45 W/m^2 difference above, we first recall that $ACRE$ is calculated as the difference between the clear-sky net radiative heating and the all-sky net radiative heating. Tian et al. (2001) use the Earth Radiation Budget Experiment (ERBE) for the top-of-atmosphere fluxes, and they use Tropical Ocean Global Atmosphere Coupled Ocean-Atmosphere Response Experiment (TOGA COARE) and Central Equatorial Pacific Experiment (CEPEX) observations for the surface fluxes. For the clear-sky radiative heating, Tian et al. (2001) use the tropical profile from Dopplnick (1979). The radiative heating profiles calculated in Dopplnick (1979), however, contain clouds (see Dopplnick, 1972, for details). Therefore, the $ACRE$ estimate provided by Tian et al. (2001) is likely an underestimate, yet it is still greater than that calculated from the CERES data. The roughly 30 W/m^2 difference between our value and that of Tian et al. (2001) results from the 45 W/m^2 difference in all-sky radiative cooling being offset by the 15 W/m^2 difference in clear-sky radiative cooling.

The divergence of total energy transport (AHT) is only about 10 W/m^2 less in our clima-

tology than the value reported by Tian et al. (2001), despite larger differences between the component terms: internal, potential, and latent. Our value for the total surface turbulent heat fluxes is about 30 W/m^2 larger than that of Tian et al. (2001), which, when combined with the difference in surface radiative fluxes, makes our oceanic energy divergence about 15 W/m^2 weaker than the value given by Tian et al. (2001).

The values that are of most interest for this study are the Atmospheric Convective Radiative Effect (ACoRE) and the net energy divergence (or atmospheric heat transport; AHT). Over the fourteen year record and within the warm pool, the ACoRE balances 66% of the total AHT, with the Atmospheric Cloud Radiative Effect (ACRE) balancing 56% of the AHT and the Atmospheric Moisture Radiative Effect (AMRE) balancing 9.6% of the AHT. The ratio of ACRE to AHT given in Fig. 2.2 is smaller than that calculated by Tian et al. (2001, roughly 88% in their data). We offset some of that difference with the inclusion of the radiative heating due to moisture. The ACoRE and the AHT vary from month to month throughout the fourteen year time period. The variations, however, are not random, but are well-correlated (significant at the 95% level; see Fig. 2.3).

Since we do not know either ACoRE or AHT with absolute certainty, we use empirical orthogonal function (EOF) analysis to minimize the perpendicular distance of the data from the linear fit instead of ordinary least-squares. Figure 2.3 shows that best fit line is roughly a 1:1 line with a y-intercept of 8.73 W/m^2 (meaning even when $\text{ACoRE} = 0$, there is still heat transport of nearly 9 W/m^2). We can compare this y-intercept value with the predicted value of AHT when $\text{ACoRE} = 0$. By energy conservation, the heat transport must equal the difference in surface turbulent fluxes and the radiative cooling of the atmosphere. We can break the radiative cooling of the atmosphere into a non-convective radiative cooling amount similar to our construction of AMRE (see above) and ACoRE. Written out this gives

$$\text{AHT} = \text{ACoRE} + (\text{LE} + \text{SH}) - R_0, \quad (2.1)$$

where LE is the surface latent heat flux, SH is the surface sensible heat flux, and R_0 is the non-convective radiative cooling of the atmosphere. From equation (2.1), when ACoRE = 0, $AHT = (LE+SH) - R_0$. Over the warm pool, $LE+SH = 149 \text{ W/m}^2$ (see Fig. 2.2) and $R_0 = 135 \text{ W/m}^2$ (equivalent to Rclr-AMRE). Thus, when ACoRE = 0, we would expect $AHT = 14 \text{ W/m}^2$, close to the 8.73 W/m^2 y-intercept for the EOF line in Fig. 2.3, considering the uncertainty on the surface radiative flux estimates.

We next bin ACoRE (and its components: ACRE and AMRE) and AHT by sea surface temperature (see Fig. 2.4, left panel) and vertical velocity along the 500 hPa surface (see Fig. 2.4, right panel). At SSTs warmer than about 301 K, ACoRE and AHT rise together at nearly a 1:1 rate with increasing SST. Most of the increase in ACoRE comes from increases in the clouds (ACRE). Over the same range of SSTs, we see that the surface turbulent fluxes (latent and sensible heat) do not vary much (within $\pm 10 \text{ W/m}^2$ of 150 W/m^2). By energy conservation, changes in ACoRE, AHT, and surface turbulent fluxes (LE+SH) must balance one another, so it is guaranteed that if ACoRE and AHT increase at exactly the same rate with increasing SST, then LE+SH must be approximately constant. When we bin the data by vertical velocity on the 500 hPa surface (ω_{500}), the relationship between ACoRE and AHT is similar to that when the data are binned by SST. As the pressure velocity decreases (stronger upward motion), ACoRE and AHT increase together, while the surface turbulent fluxes decrease slightly. Again, most of the increase in ACoRE comes from an increase in ACRE. The relationships between ACoRE and AHT can be seen elsewhere in the tropics as well. We bin the tropical average ACoRE and AHT (not including regions over land) by SST and ω_{500} (see Fig. 2.5). Figure 2.5 shows that the ACoRE and AHT relationship shares many features with the warm pool (though spanning a larger range of SSTs).

We investigate other convective regions of the tropics (see table 2.2) and find similar structures in the ACoRE and AHT relationship in these regions to those found in the warm pool (not shown). Table 2.2 shows the amount of AHT balanced by ACoRE for the warm

pool, tropical average, the Atlantic, the Indian, the East Pacific ITCZ, and the broader West Pacific regions. We note that the percentage of AHT balanced by ACoRE is roughly the same over convective regions at 66–87%. The breakdown of ACoRE into ACRE and AMRE is also roughly the same over convective regions (56–74% ACRE and 5.5–13% AMRE) suggesting this is a robust feature of tropical convection.

We next shift focus to the top of the atmosphere (TOA). Figure 2.6 shows the net TOA radiative fluxes (Rnet) and the TOA cloud radiative effect (CRE). As expected, increases in convection increase both the shortwave cloud radiative effect (SWCRE) and the longwave cloud radiative effect (LWCRE) as seen when binned by either SST or ω_{500} . CRE, however, behaves differently depending on whether we bin by SST or ω_{500} . While CRE does not have any discernible relationship with the underlying SSTs over the warm pool, when binned by ω_{500} , CRE decreases for increasing ω_{500} (see Fig. 2.6). Increases in vertical velocity are indicative of increases in convection, which one would expect to result in increases in the magnitude of both SWCRE and LWCRE. We find, however, that SWCRE increases in magnitude more rapidly with ω_{500} than does LWCRE. We will explore this result in greater detail below. Figure 2.6 (right) shows that the decrease in CRE with increasing convection is balanced by an increase in AMRE such that the net radiative fluxes at the top of the atmosphere (Rnet) are unchanged. The insensitivity between circulation strength and top-of-atmosphere fluxes was shown by Clement and Soden (2005), though they suggested the insensitivity was a cloud-driven effect while we show that water vapor is also important.

We perform principal component analysis on MODIS data to investigate the behavior of the CRE over the warm pool. Figure 2.7 shows the mean histogram of MODIS-retrieved cloud optical thickness versus cloud top pressure over the warm pool (for brevity, we refer to this as the MODIS histogram). As expected, high clouds (cloud top pressure < 440 hPa) are the most common cloud type observed by MODIS over the warm pool.

We calculate the empirical orthogonal functions (EOFs) for the MODIS histogram at all

points within the warm pool and for all months (see Fig. 2.8). The first three EOFs show the three modes of variability that explain the largest fraction of the variance in the MODIS histogram. The first EOF (53% variance explained) is an increase in high clouds across a range of optical depths and cloud top pressures similar to the mean MODIS histogram structure. The second EOF (20% of variance explained) is a shift from optically thin to optically thick clouds. The third EOF (11% of variance explained) is a vertical shift of the high clouds in pressure coordinates. We can think of these three EOFs as cloud amount, cloud optical thickness, and cloud top height, respectively. We then calculate the principal components for each EOF to get the amplitudes of each pattern across all points within the warm pool and all months of the MODIS record (see Fig. 2.9).

As the SSTs increase, PC 1 and PC 3 increase, while PC 2 decreases. As ω_{500} decreases, all three principal components increase, with PC 1 showing the largest change. We use the net cloud forcing histograms from (Hartmann et al., 2001b, Fig. 5a) to get a measure of how the cloud radiative effect changes depending on the amplitude of each EOF pattern. The change in top-of-atmosphere net cloud radiative effect for a one standard deviation increase is -3.78 W/m^2 for EOF 1, -6.63 W/m^2 for EOF 2, and $+1.36 \text{ W/m}^2$ for EOF 3. The cloud amount change associated with one standard deviation of the principal component of EOF 1 is much more than the cloud amount changes associated with one standard deviation of the principal components of EOFs 2 and 3. If we were to normalize the three leading EOFs such that the increases in clouds are all 10%, then the changes in CRE would be -2.56 W/m^2 for EOF 1, -14.0 W/m^2 for EOF 2, and $+5.20 \text{ W/m}^2$ for EOF 3. The previous calculation reveals that changes in cloud optical thickness or cloud top height have the potential to alter the CRE much more than changes in overall cloud amount, but do not vary enough over the warm pool to be the dominant factor. Cloud amount (EOF 1) has a weak negative effect on CRE.

We bin the CRE changes explained by each principal component against SST and ω_{500} as

before (see Fig. 2.10). Figure 2.10 shows that the first three PCs of the MODIS histogram do a good job of representing the total change in CRE from its mean value due to changes in SST or ω_{500} (compare Fig. 2.10 with Fig. 2.6). The overall change in CRE with increasing SST is small, as we saw before, while increasing ω_{500} leads to increases in CRE, again, as seen above. Further, Fig. 2.10 shows that the small change in CRE with increasing SST is largely due to compensation between the effects of different EOFs. As SST increases, the effect of cloud amount increases (PC1) is compensated by both cloud optical thickness decreases (PC2) and cloud top height increases (PC3). As SSTs increase, the cloud amount increases and the ΔCRE becomes more negative (at -3.78 W/m^2 per standard deviation increase). At the same time, however, increasing SSTs are associated with a shift toward optically thinner clouds providing a positive change in ΔCRE , while increasing SSTs favor higher cloud tops (also increasing ΔCRE). Since the magnitudes of these changes are comparable, little net change in CRE results from changes in SST over the warm pool. As for changes in ω_{500} , much of the variability is due to changes in cloud amount (PC 1), with cloud optical thickness being a secondary factor, and cloud top height playing virtually no role at all. As upward vertical velocities increase (more convection), the amount of clouds increase and their optical thickness distribution shifts towards thicker clouds. Since both of these features act to drive ΔCRE in the same direction, there is no cancellation effect, and the CRE becomes more negative as upward vertical velocity increases.

Figure 2.11 shows that the atmospheric heat transport responds more strongly to PC 1 (cloud amount) than either PC 2 or PC 3 (cloud optical thickness and cloud top height, respectively). Changes from optically thin to optically thick clouds (PC 2) have virtually no impact on the AHT, ACoRE, ACRE, or AMRE. Changes in cloud top height (PC 3) have a peculiar behavior with a minimum in AHT and ACRE near an amplitude of +1 for PC 3. The relationship between cloud amount (PC 1) and AHT is very similar to the relationship between ω_{500} and AHT, reaffirming that the influence of vertical velocity on

AHT and ACoRE is driven by changes in cloud amount and not cloud structure.

2.4 Additional sensitivity tests

As noted in section 2.2, several sensitivity tests were performed to test the robustness of our conclusions. In short, the structures in the relationship between atmospheric convective radiative effect (ACoRE) and atmospheric heat transport (AHT) are robust. The relationships have virtually no change whether we use an “indirect” or “direct” calculation of AHT, whether we use daily instead of monthly data, or whether we use the ERA-Interim radiative fluxes in place of CERES fluxes. Note that the “indirect” calculation of AHT uses the top-of-atmosphere and surface fluxes to determine the net energy divergence while the “direct” calculation uses the actual divergence of total energy.

We begin with errors in the Atmospheric Heat Transport (AHT). To investigate whether these errors may confound our results, we compare the AHT output from ERA-Interim (the “direct” AHT calculation) with that computed using the top-of-atmosphere and surface fluxes (the “indirect” AHT calculation). The structural relationship between ACoRE and AHT is not changed when switching between the direct and indirect AHT calculations (not shown). The magnitude of the average AHT over the warm pool drops by about 6.7 W/m^2 when using the indirect calculation compared to the direct calculation.

When using ERA-Interim radiative fluxes, the major difference comes from ACRE, with the ERA-Interim estimate being smaller than that of CERES. ERA-Interim derived ACoRE averaged over the warm pool is 41.8 W/m^2 (59% of AHT), with $ACRE = 33.1 \text{ W/m}^2$ (47% of AHT) and $AMRE = 8.67 \text{ W/m}^2$ (12% of AHT). ACRE is smaller in the ERA-Interim estimate than the CERES estimate (-6.8 W/m^2). AMRE is larger in the ERA-Interim estimate than the CERES estimate ($+1.9 \text{ W/m}^2$). The ACRE decrease is more than twice that of the AMRE increase, such that ACoRE is smaller when calculated by ERA-Interim instead of CERES (-4.9 W/m^2). As expected from the ACRE results above, there are shifts

in the magnitudes of the cloud radiative effects and net radiative fluxes, but the conclusion are unchanged (not shown).

We have tested whether our conclusions are sensitive to the monthly averaging timescale, most notably the atmospheric heat transport. We downloaded one year of daily ERA-Interim data from the time period used in the study. For this test, we make use of the ERA-Interim radiative fluxes, since they behave similarly to the CERES-EBAF fluxes as noted above. Year 2013 was chosen as it is the last full calendar year in our time period. The patterns in AHT, ACoRE, etc. are the same for both the daily and monthly data (not shown).

2.5 Conclusions

The analysis and data presented here confirm that the atmospheric cloud radiative effect (ACRE) contributes to enhanced atmospheric heat transport (AHT) in the tropics as was shown by previous studies (e.g. Sherwood et al., 1994; Sohn, 1999; Bergman and Hendon, 2000a; Tian et al., 2001; Tian and Ramanathan, 2002). Additionally, we find the atmospheric moisture radiative effect (AMRE) is important and acts in the same way as the cloud radiative heating, and we refer to their combined effect as the atmospheric convective radiative effect (ACoRE). Atmospheric radiative heating due to increases in clouds and water vapor balances 66% of the net energy divergence out of the tropical warm pool atmosphere (10 S – 10 N, 150 – 180 E) during our fourteen year time period. Tian et al. (2001) found ACRE balanced 88% of AHT during the TOGA COARE and CEPEX period. Month to month changes in the net energy divergence and the radiative heating from clouds and water vapor co-vary at roughly a 1:1 rate, shown by a linear fit, which explains 85% of the variance. The co-variations in the net horizontal energy divergence and the atmospheric radiative heating from clouds and water vapor show consistent relationships with sea surface temperature and vertical velocity. Increases in sea surface temperature lead to equal increases in both net horizontal energy divergence and atmospheric radiative heating. The same is true for

increases in upward vertical velocity. These results hold in different regions of the tropics as well, with the proportion of net horizontal energy divergence balanced by atmospheric radiative heating from clouds and water vapor ranging from 66-87%.

In the warm pool, the cloud radiative effect (CRE) at the top of the atmosphere is insensitive to sea surface temperature changes, owing to compensation between increases in all types of high clouds and a shift toward higher, thinner clouds. The longwave and shortwave cloud radiative effects (LWCRE and SWCRE, respectively) both increase in magnitude with increasing sea surface temperature in good agreement with Bony et al. (1997). Additionally, the net radiative fluxes at the top of the atmosphere are insensitive to changes in vertical velocity (in agreement with Clement and Soden, 2005). Unlike Clement and Soden (2005), we find the insensitivity of top-of-atmosphere fluxes to vertical velocity is due to compensating changes in the CRE and AMRE. Negative anomalies in CRE are balanced by enhanced water vapor greenhouse effect. The decrease in CRE with increasing vertical velocity agrees with results from Yuan et al. (2008). The introduction of AMRE, however, is new to this study. The insensitivity of top-of-atmosphere fluxes to circulation strength means horizontal transport of energy by the atmosphere can increase or decrease without any change in the top-of-atmosphere fluxes. If the atmosphere remains in steady-state, then surface fluxes must compensate the change in atmospheric heat transport. Increases in upward vertical velocity increase cloud amount as well, but also shift the cloud distribution toward more optically thick clouds, thus decreasing the cloud radiative effect. We have shown that cloud amount (EOF 1) is the dominant mode of variability for the cloud systems and accounts for the majority of the month-to-month changes in both the net horizontal energy divergence out of the column and the atmospheric convective radiative effect. It is reasonable to expect cloud amount to be tied to the mass flux. Similarly, the amount of water vapor lofted into the upper troposphere should also be tied to the mass flux. The amount of water vapor lofted is responsible for determining the atmospheric moisture radiative effect. As noted above, the

insensitivity of the top-of-atmosphere fluxes to circulation strength is only achieved by the cancellation between decreasing CRE and increasing AMRE. It is worth investigating the cause of this cancellation in future work. Additionally, it is important to ask whether the cancellation between changes in CRE and AMRE can be expected to continue in a warming climate. If not, strong feedbacks may arise due to the imbalance at the top of the atmosphere that either enhance or dampen variability in the tropical circulation.

The work presented in this paper raises further questions related to the relationship between atmospheric convective radiative effect (ACoRE) and the net divergence of energy (AHT). For instance, do models reproduce the same ACoRE-AHT relationship seen above? If yes, is that relationship preserved through a warming climate? There is no guarantee that the same ACoRE-AHT relationship will be preserved in model simulations. If the surface turbulent heat fluxes are too sensitive to changes in convection, then the relationship seen in observations may not hold up. Additionally, it has been shown difficult to obtain a maximum in precipitation coincident with a minimum in evaporation in fixed-SST experiments (see Sobel, 2003), suggesting that the surface cloud radiative effect is an important component of the ACoRE-AHT relationship through its coupling with latent heat flux. If the radiative effects of clouds are removed, do we see a drop off in AHT equivalent to the atmospheric cloud radiative effect? Does a slowdown of the tropical circulation in a warmer climate necessitate a reduction in cloud amount as is seen in the current climate? The answers to these questions will deepen our understanding of the relationship between the tropical circulation and the radiative heating due to clouds and water vapor.

The CERES EBAF data were obtained from the NASA Langley Research Center CERES ordering tool (<http://ceres.larc.nasa.gov/>). The ERA-Interim data were obtained from the ECMWF Data Server (http://apps.ecmwf.int/datasets/data/interim_full_moda/). The MODIS data used in this study were acquired as part of the NASA's Earth-Sun System Division and archived and distributed by the MODIS Adaptive Processing Sys-

tem (MODAPS). These MODIS data were obtained from the LAADS FTP server (`ftp://ladsweb.nascom.nasa.gov/`). The authors also wish to thank our anonymous reviewers for their suggestions toward improving this manuscript. Funding for this work was provided by NSF Grant # AGS-0960497.

Table 2.1: List of acronyms.

TOA	Top-of-Atmosphere
SFC	Surface
SST	Sea Surface Temperature
OLR	Outgoing Longwave Radiation
LE	Surface Latent Heat Flux
SH	Surface Sensible Heat Flux
CRE	Cloud Radiative Effect (at TOA)
SWCRE	Shortwave Cloud Radiative Effect
LWCRE	Longwave Cloud Radiative Effect
ACRE	Atmospheric Cloud Radiative Effect
SCRE	Surface Cloud Radiative Effect
AMRE	Atmospheric Moisture Radiative Effect
ACoRE	Atmospheric Convective Radiative Effect
AHT	Atmospheric Heat Transport
ERA-Interim	European Center for Medium-Range Weather Forecasts (ECMWF) Reanalysis Interim
CERES-EBAF	Clouds and the Earth's Radiant Energy System-Energy Balanced and Filled
MODIS	Moderate Resolution Imaging Spectroradiometer
EOF	Empirical Orthogonal Function
PC	Principal Component

Table 2.2: Average ACRE, AMRE, ACoRE, and AHT for various regions in the tropics. The numbers in parentheses are the fraction of AHT balanced by ACRE, AMRE, or ACoRE. All regions range from 10 S – 10 N except East Pacific which is 5 – 15 N. All averages are taken over ocean grid cells only.

	ACRE W/m ²	AMRE W/m ²	ACoRE W/m ²	AHT W/m ²
Warm Pool (150 – 180 E)	39.9 (56%)	6.81 (9.6%)	46.7 (66%)	70.7
Tropics (0 – 360 E)	23.7 (74%)	4.28 (13%)	28.0 (87%)	32.2
Indian (40 – 100 E)	30.9 (62%)	4.55 (9.2%)	35.5 (71%)	49.6
Atlantic (300 – 360 E)	14.7 (60%)	2.62 (11%)	17.3 (71%)	24.3
West Pacific (120 – 180 E)	40.3 (63%)	6.09 (9.6%)	46.4 (73%)	63.7
East Pacific (210 – 260 E)	23.6 (66%)	1.97 (5.5%)	25.6 (71%)	35.9

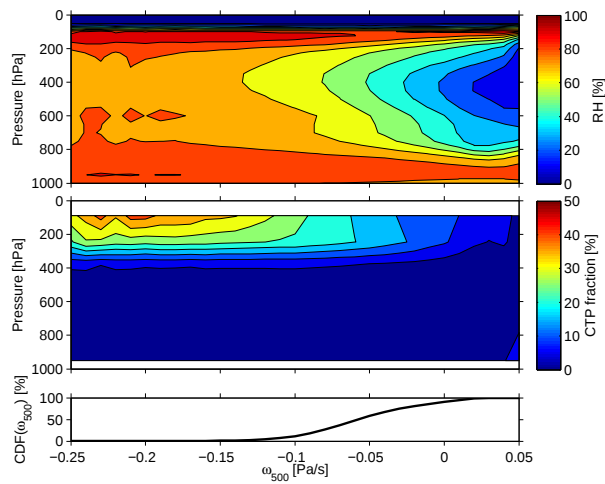


Figure 2.1: (Top) Relative humidity [%] as a function of pressure binned by ω_{500} for the warm pool region. Each bin has a width of 0.01 Pa/s. (Middle) Same as top, but for MODIS-retrieved cloud top pressure frequency. (Bottom) Cumulative distribution function of ω_{500} .

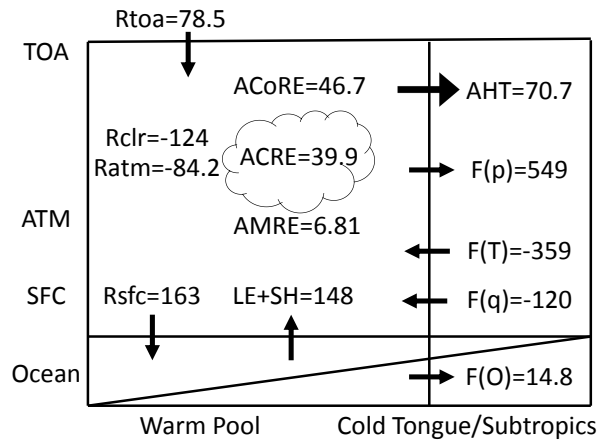


Figure 2.2: Mean values over warm pool region (10 S – 10 N, 150 – 180 E) for radiative and turbulent fluxes, energy transport terms, and radiative heating (all in W/m^2). AHT is the net divergence of moist static energy, $F(p)$ is the divergence of potential energy, $F(T)$ is the divergence of sensible energy, $F(q)$ is the divergence of latent energy, and $F(O)$ is the divergence of energy through the ocean. ACoRE is the Atmospheric Convective Radiative Effect, ACRE is the Atmospheric Cloud Radiative Effect, and AMRE is the Atmospheric Moisture Radiative Effect. R_{clr} and R_{atm} are the integrated clear-sky radiative cooling and integrated radiative cooling, respectively. R_{toa} and R_{sfc} are the top-of-atmosphere and surface radiative fluxes, respectively. $LE+SH$ is the combined latent and sensible heat fluxes at the surface.

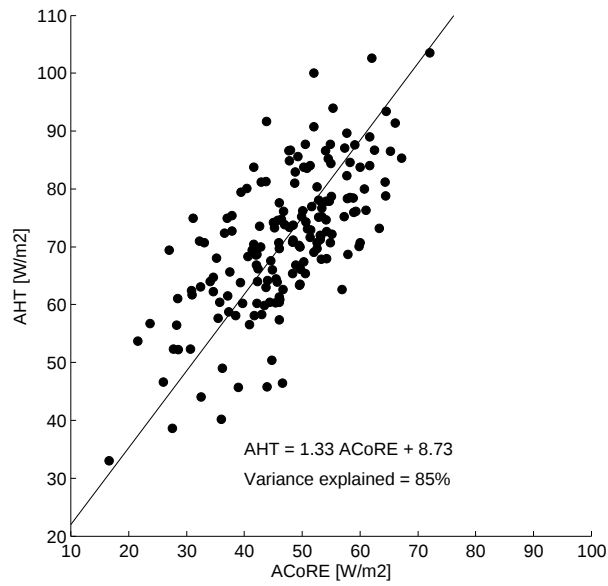


Figure 2.3: Atmospheric Convective Radiative Effect ($ACoRE = ACRE + AMRE$) versus total energy divergence (AHT) over warm pool. Each point represents one monthly averaged value over the entire warm pool region (10 S – 10 N, 150 – 180 E). The line is the linear fit to the data obtained by principal component analysis (in this case, 85% of the variance is explained by this line).

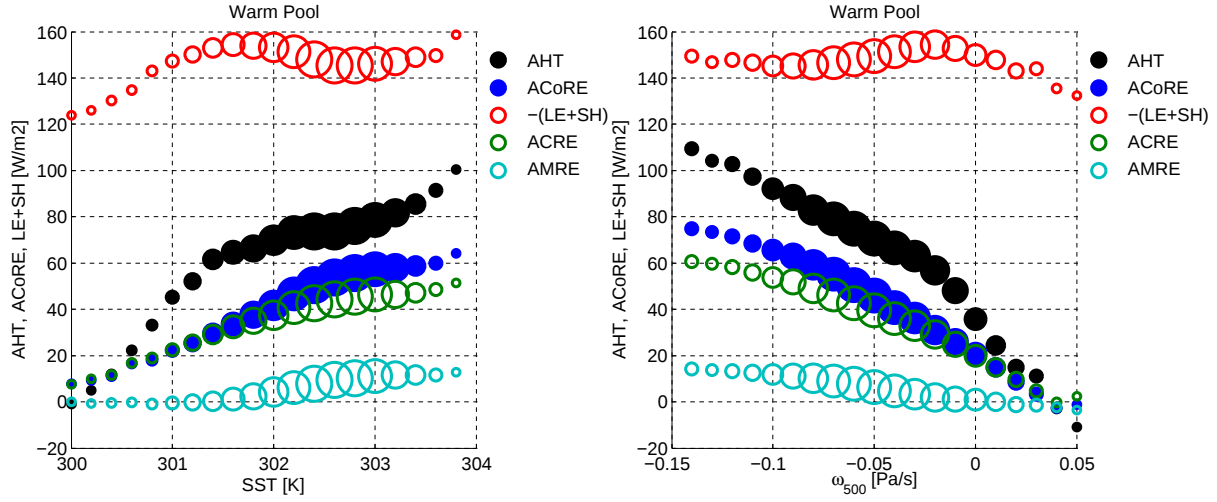


Figure 2.4: (Left) ACoRE, AHT, ACRE, AMRE, and surface turbulent fluxes (LE+SH) binned by SST (bin intervals are 0.2 K). The size of the circle is scaled by the frequency of occurrence of that SST in the warm pool region. (Right) Same as (left) only binned by vertical velocity on the 500 hPa surface (ω_{500} ; bin intervals are 0.01 Pa/s).

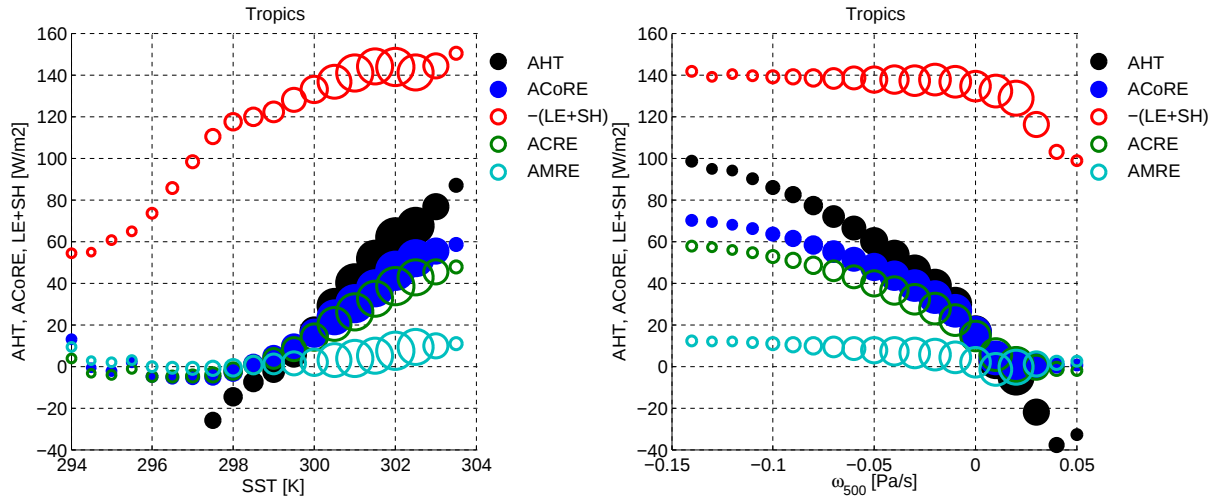


Figure 2.5: Same as Fig. 2.4 only over the entire Tropics (10 S – 10 N, 0 – 360 E); ocean grid cells only. The range of SSTs for the Tropics is larger than that used for binning the data over the warm pool.

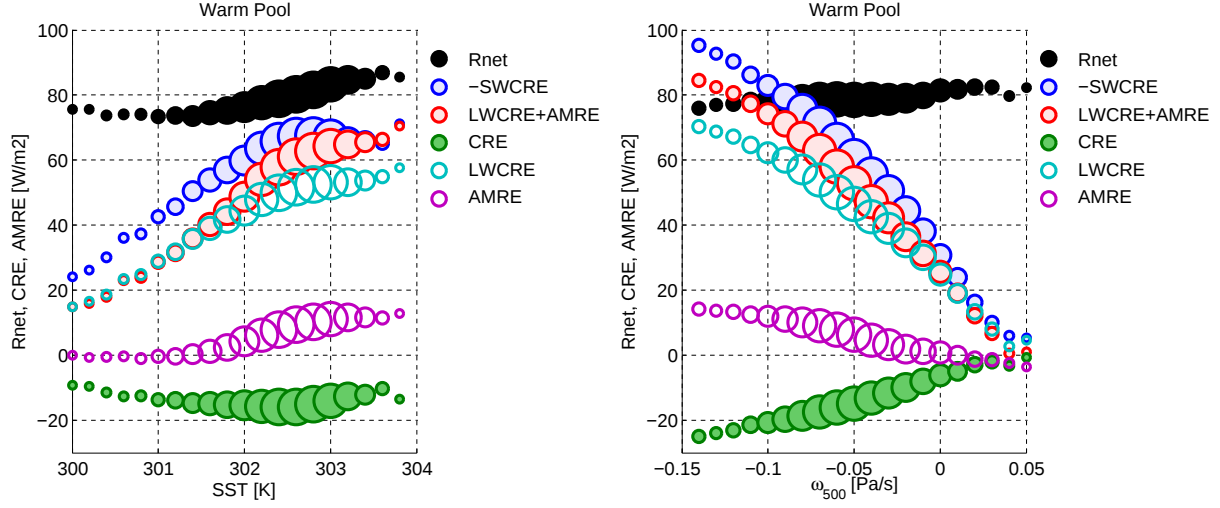


Figure 2.6: (Left) Rnet, SWCRE, LWCRE+AMRE, CRE, LWCRE, and AMRE in the warm pool binned by SST (bin intervals are 0.2 K). The size of the circle is scaled by the frequency of occurrence of that SST in the warm pool region. (Right) Same as (left) only binned by vertical velocity on the 500 hPa surface (ω_{500} ; bin intervals are 0.01 Pa/s).

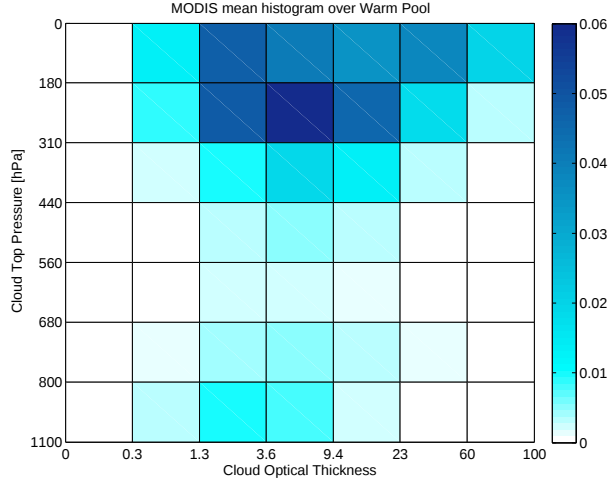


Figure 2.7: MODIS frequency histogram of cloud optical thickness versus cloud top pressure. The data are monthly mean values over the warm pool region defined above.

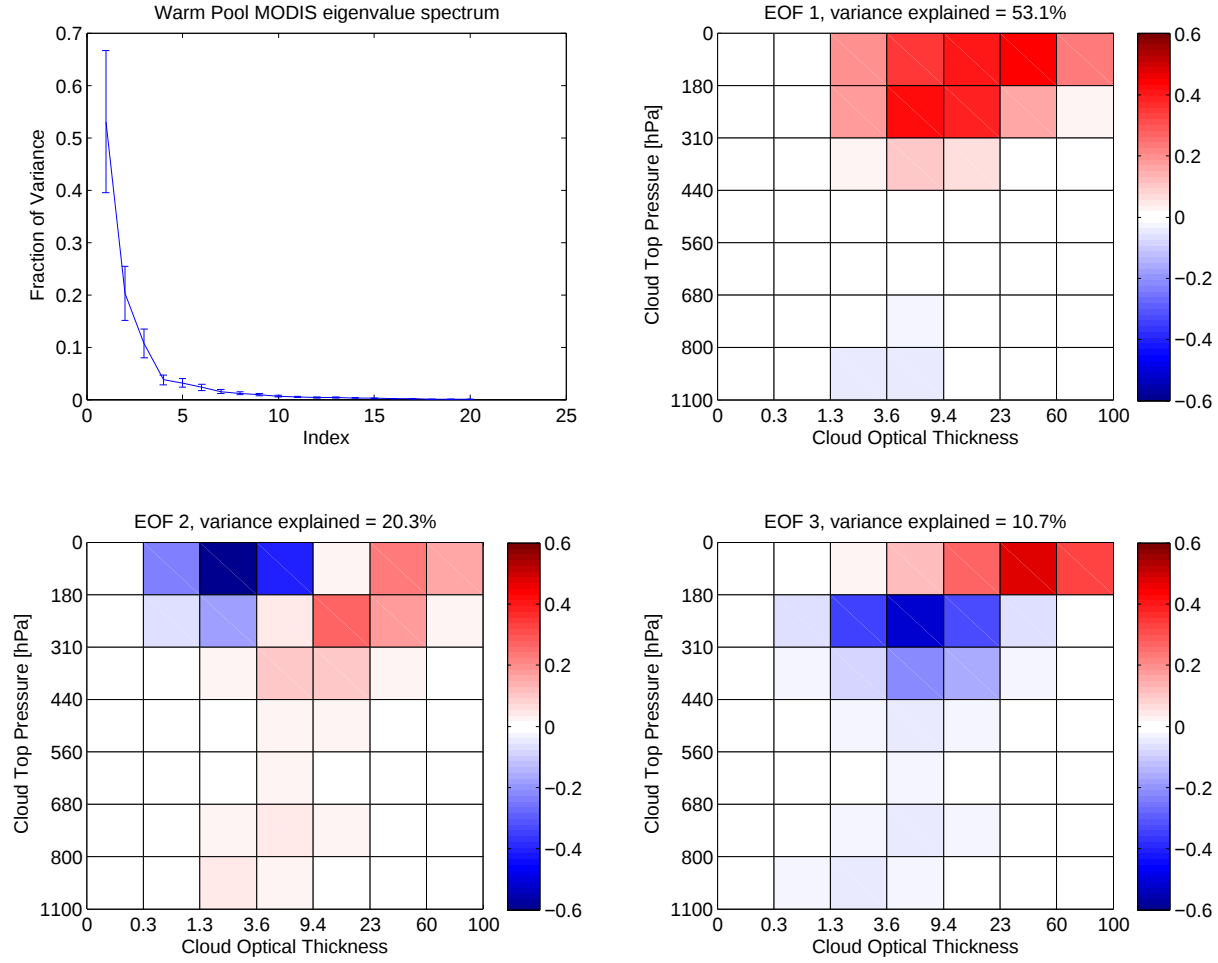


Figure 2.8: (Upper left) Eigenvalue spectrum for EOF analysis of MODIS cloud optical thickness versus cloud top pressure histogram. (Upper right, lower left, and lower right) The first three EOFs of the MODIS histogram. The change in top-of-atmosphere net cloud radiative effect for a one standard deviation increase in each of the above EOFs is -3.78 W/m^2 , -6.63 W/m^2 , and $+1.36 \text{ W/m}^2$, respectively.

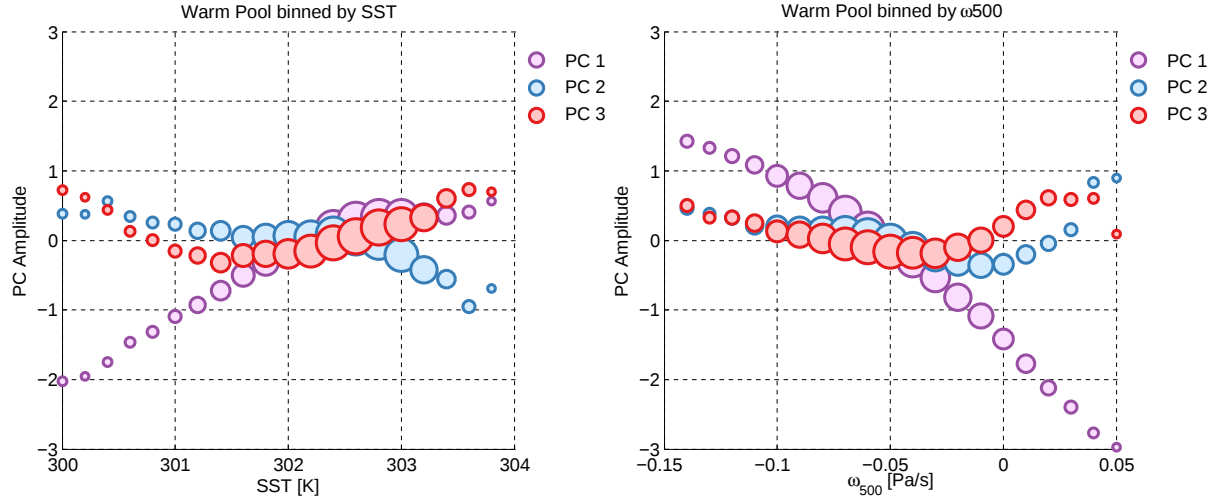


Figure 2.9: Principal components of EOFs presented in Fig. 2.8 binned by (left) SST and (right) ω_{500} . PCs 1, 2, and 3 represent changes in cloud amount, cloud optical thickness, and cloud top height, respectively.

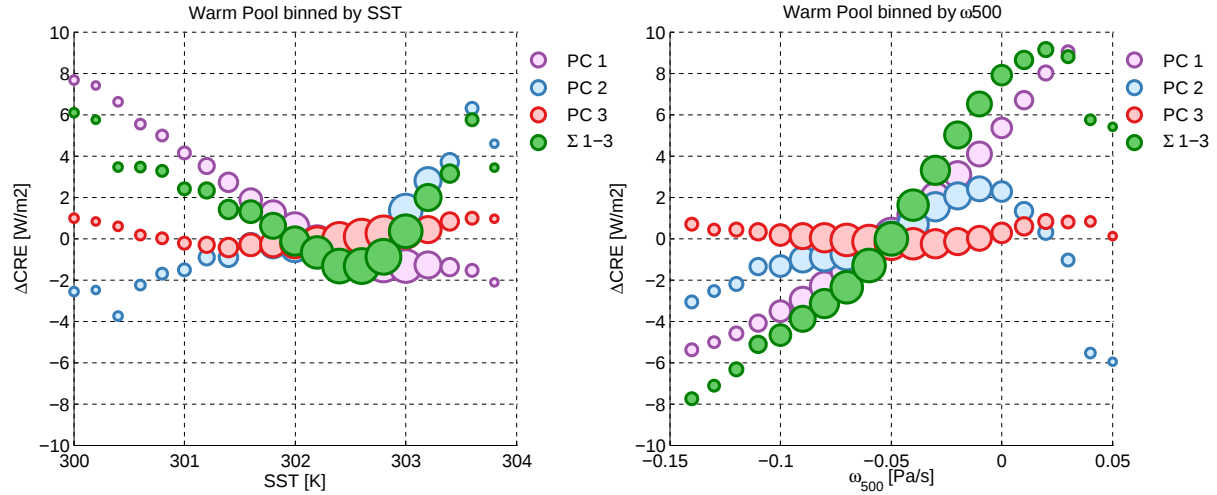


Figure 2.10: Change in CRE due to variability in cloud amount, cloud optical thickness, and cloud top height (PCs 1, 2, and 3, respectively) as binned by (left) SST and (right) ω_{500} . The green dots represent the sum of the change in CRE due to the first three principal components.

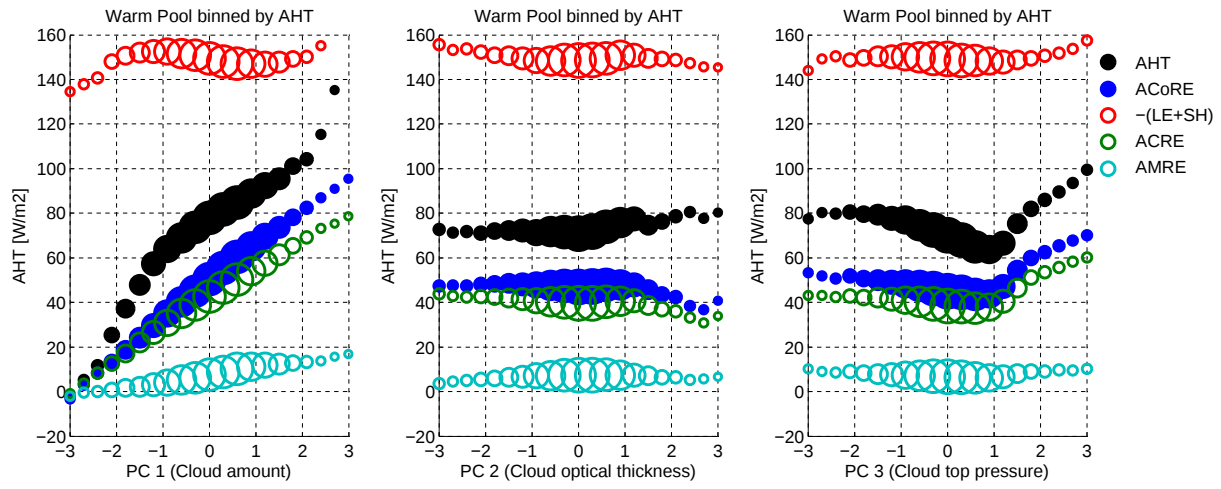


Figure 2.11: Same as Fig. 2.4 only AHT, ACoRE (ACRE+AMRE), LE+SH, ACRE, and AMRE are binned by the amplitude of (left) PC 1, (middle) PC 2, and (right) PC 3.

Chapter 3

THE ROLE OF CLOUD RADIATIVE HEATING IN DETERMINING THE LOCATION OF THE ITCZ IN AQUA PLANET SIMULATIONS

3.1 Introduction and background

The Intertropical Convergence Zone (ITCZ) describes zonally oriented bands of heavy precipitation in the tropics where surface winds converge. The ITCZ is important for the precipitation, but it is also a critical factor in simulating the global energy and moisture distributions and transports, as well as the tropical top of atmosphere and surface fluxes. Its location has a strong impact on the trade winds and ocean circulation. On average, Earth has a single ITCZ across the tropical Pacific that is displaced into the Northern hemisphere. While a secondary ITCZ is occasionally observed in the Southern Pacific, it is a transient phenomenon occurring primarily in boreal spring (Hubert et al., 1969; Zhang, 2001). Coupled climate models, however, often show a second ITCZ in the Southern Hemisphere in the climatological mean rainfall patterns (see, for example, Mechoso et al., 1995; Lin, 2007; Flato et al., 2013; Li and Xie, 2014). This unrealistic secondary ITCZ is often referred to as the double ITCZ problem. Tian (2015) showed that the double ITCZ bias negatively correlates with ECS — ECS is higher in models with a weaker bias.

Double ITCZ features are not restricted to coupled models alone and have been shown to exist even in aqua planet simulations with fixed sea surface temperature patterns (Numaguti, 1993; Barsugli et al., 2005; Liu and Moncrieff, 2008; Liu et al., 2010; Dahms et al., 2011; Möbis and Stevens, 2012; Oueslati and Bellon, 2013a,b; Landu et al., 2014). Across many studies, a variety of factors have been shown capable of shifting a model from a sin-

gle to double ITCZ. Examples include changing the model resolution (Landu et al., 2014), removing wind contributions to surface evaporation (Numaguti, 1993), changing convective parameterizations (Numaguti, 1993; Liu et al., 2010; Möbis and Stevens, 2012; Oueslati and Bellon, 2013a), changing the SST gradient (Dahms et al., 2011; Oueslati and Bellon, 2013b) using homogenized radiative cooling (Liu and Moncrieff, 2008), and even changing the solar constant (Barsugli et al., 2005).

Chao and Chen (2004) summarized the ITCZ position as the sum of two “attractors.” The first attractor pulls the ITCZ toward the equator due to wave stability arguments, while the second attractor pulls the ITCZ poleward following arguments based on circulation strength and surface turbulent energy fluxes. Circulation strength is an important component to the surface fluxes through a positive feedback whereby stronger circulations drive faster winds, thereby picking up more latent and sensible heat from the surface to supply to the convection. Only the second of these attractors depends on model parameters. We suggest that the cloud radiative heating behaves like the first attractor, enhancing convection near the equator. Chao and Chen (2004), however, performed an experiment where the cloud-radiation interactions were removed and compared it to one where they were not and found the ITCZ locations to be roughly unchanged. Their experiments were performed in an aqua planet model with globally uniform sea surface temperature. Unlike the experiments of Chao and Chen (2004), the experiments examined in this manuscript have a SST distribution with a maximum at the equator. The SST distribution, as we will note again later, is an important component to the difference in our results and those of Chao and Chen (2004).

As noted above, we suggest that the cloud radiative heating within the atmosphere is also an attractor for the ITCZ to be closer to the equator, which is also the region of warmest sea surface temperature (SST) in the experiments described. The cloud radiative heating warms the atmosphere substantially, requiring higher near surface moist static energy (MSE) values in order to initiate deep convection. This is the same idea as the “upped-ante” mechanism

described by Neelin et al. (2003) — in a warmer atmosphere, higher boundary layer MSE is required for convection. A similar idea was presented by Landu et al. (2014) based upon arguments pertaining to CAPE, and we will use this CAPE approach in our analysis. We will perform our analysis for six models: those used in the Clouds On-Off Climate Inter-comparison Experiment (COOKIE; see Stevens et al., 2012, for details). We will use the aqua planet simulations for the COOKIE experiments. Aqua planets offer an attractive test bed for learning about the atmosphere due to their removal of Earth’s irregularities (Lorenz, 1967; Hoskins et al., 1999; Miura et al., 2005; Medeiros et al., 2008, 2015). Removing the land masses is especially attractive for exploring the relation between clouds and the tropical circulation because land-ocean temperature contrasts may exert a significant influence on the circulation patterns. Using the aqua planet simulations from the COOKIE experiments removes these complexities from consideration, yielding a simpler experiment. Crueger and Stevens (2015) showed a preference for a double ITCZ when cloud-radiation interactions are removed in the COOKIE atmosphere-only models (fixed sea surface temperature experiments with realistic geography), though they do not offer a physical mechanism for this phenomenon.

Cloud radiative heating is an important component of the tropical circulation (see Ramanathan, 1987; Slingo and Slingo, 1988, 1991; Stuhlmann and Smith, 1988a,b; Randall et al., 1989; Sherwood et al., 1994; Zhang and Rossow, 1997; Sohn, 1999; Bergman and Hendon, 2000b,a; Raymond, 2000; Tian et al., 2001; Tian and Ramanathan, 2002, 2003; Lee and Yoo, 2014; Del Genio and Chen, 2015; Crueger and Stevens, 2015; Harrop and Hartmann, 2015; Merlis, 2015). Other studies have suggested a role for local cloud radiative heating in determining the ITCZ position. Voigt et al. (2014) showed that the spread in model ITCZ shift is related to the tropical cloud radiative heating. Because climate models tend to have varied responses in how their clouds change with warming (either a prescribed SST increase or by CO₂ forcing), the corresponding change in circulation due to the new heating patterns

varies strongly from model-to-model as well. Voigt and Shaw (2015) showed similar results from aqua planet simulations. They noted that in global warming experiments, cloud radiative heating led to a latitudinal contraction of precipitation in one model while an expansion of precipitation in another.

Another example of a local tropical effect was considered by Li and Xie (2014) who found the lack of snow-radiation interactions a means for influencing the eastward overextension of the SPCZ seen in the CMIP models. In addition, Raymond (2000) showed that differential cloud radiative heating is “fundamental” for the tropical circulations. Raymond (2000) also showed that precipitation remains over the warmest SSTs when cloud radiative heating is removed — opposite to what we find here — although the distribution of precipitation widens. Wofsy and Kuang (2012) showed the warm pool (defined as the precipitating region within their cloud-resolving model) narrows as radiative cooling increases in magnitude. They explain this as a response of the increase in precipitation needed to balance the increase in radiative cooling favoring the warmest SSTs.

Despite the models used in the COOKIE experiments having different resolutions, different dynamics, different physical parameterizations, and, ultimately, different climates, their ITCZ responses to the removal of the radiative effects of clouds are consistent. Our results confirm the increase in precipitation and circulation strength when cloud radiative effects are included, at least locally. Across the whole of the tropics, however, total precipitation decreases when cloud radiative effects are included, owing to the reduced radiative cooling of the troposphere. We also show that cloud radiative effects constrain the ITCZ to be nearer the equator than when the cloud radiative effects on the atmosphere are removed. The narrowing of convection toward the warmest SSTs agrees with simulations done by Randall et al. (1989) where cloud radiative heating was turned on and off, and with simulations done by Liu and Moncrieff (2008) where radiative heating was interactive or a fixed, uniform value. Additionally, strong local precipitation rates and vertical velocities are more likely

when cloud radiative heating is present. While cloud radiative heating reduces precipitation across the tropics, it increases cloud mass (both liquid and ice), suggesting that cloud radiative heating decreases precipitation efficiency. We find that, although precipitation in the tropics is reduced, the mass circulation of the Hadley Cell increases. One reason for this is that while ACRE causes the static stability of the tropics to increase, ACRE also heats the tropical atmosphere and cools the extra-tropical atmosphere. The enhanced meridional heating gradient drives a stronger poleward heat flux, strengthening the Hadley Circulation.

3.2 Model Description and Methods

The aqua planet simulations follow the methodology of the Aqua-Planet Experiment (APE). The motivation and experimental design for APE was initially put forth by Neale and Hoskins (2000). The sea surface temperatures are fixed at the “Qobs” profile from APE. There are six companion experiments done, each with an “ACRE-on” and “ACRE-off” configuration. “ACRE-on” refers to simulations where the cloud-radiation interactions are incorporated and “ACRE-off” refers to simulations where the cloud-radiation interactions are removed. All of the models output five years of data with both ACRE-on and ACRE-off configurations. While all of the models output monthly data, only a subset output daily data. Thus, for consistency, we will use the monthly data output in this study, except where specifically noted. Because the aqua planet simulation boundary conditions have no zonal asymmetries, we zonally average all of the model outputs. Five different models are available, with one using two different physics packages (see Table 3.1 for details). The models are: HadGEM (Collins et al., 2008); CNRM (Voldoire et al., 2013); IPSL (Dufresne et al., 2013); MPI (Stevens et al., 2013); MRI (Yukimoto et al., 2012). The IPSL model is run with two different physics packages which are denoted IPSL-A and IPSL-B (see Hourdin et al., 2013a,b, respectively). We also perform another aqua planet experiment with the GFDL AM2.1 model to study the transient response to cloud effect removal.

In addition to the GCMs listed in Table 3.1, we use the System for Atmospheric Modeling (SAM) version 6.9 cloud resolving model (Khairoutdinov and Randall, 2003). The model uses the anelastic equations of motion. Our model setup has a 3072×1 horizontal domain with 1 km resolution and periodic boundary conditions. There are 96 vertical levels on a stretched grid. The upper boundary condition is a rigid lid with Newtonian damping in the upper third of the model domain to limit wave reflection. The prognostic variables in the CRM are liquid-ice moist static energy, total non-precipitating water (water vapor and cloud liquid and ice), and total precipitating water (rain, snow, and graupel). We use the Rapid and Accurate Radiative Transfer Model for GCMs (RRTMG) for all of the radiation calculations (see Mlawer et al., 1997; Iacono et al., 2008). All simulations are run to radiative-convective equilibrium. The simulations are run at the equator so there are no rotation effects (a mock-Walker circulation). We use a perpetual solar forcing with an insolation-weighted solar zenith angle following Hartmann (1994). The ozone profile used in SAM is the tropical profile used in Harrop and Hartmann (2012). We use the base SAM single-moment microphysics scheme as well as a modified microphysics scheme. The modified microphysics scheme is the “NOSEDAALIQ5” modification from Lopez et al. (2009). The “NOSEDAALIQ5” modifications remove cloud ice sedimentation, lower the threshold for ice autoconversion by a factor of 100, and increase autoconversion and accretion of liquid by factors of five. For convenience, we will refer to the base SAM microphysics as “SAM” and the modified microphysics as “NA5.” Since the cloud resolving model is a limited domain study, we do not use the same SST pattern as the COOKIE aqua planet models. Instead, we use a simple cosine distribution of SSTs with an amplitude of 1 K and a mean of 299 K for all of the simulations. This SST variation is similar in magnitude within the tropics to the Qobs profile used by the COOKIE models.

3.3 Results and Discussion

3.3.1 The influence of stability on precipitation location

Zonally and temporally averaged precipitation values show an equatorward contraction in the peak precipitation for all simulations when the cloud radiative effects are included (see Fig. 3.1). We measure the contraction in ITCZ with the precipitation-weighted latitude, calculated as follows:

$$\phi_P = \frac{\int_{0^\circ}^{\phi_{P=\min}} P \times \phi \times \cos \phi \, d\phi}{\int_{0^\circ}^{\phi_{P=\min}} P \times \cos \phi \, d\phi} \quad (3.1)$$

where P is precipitation, ϕ is latitude, and $\phi_{P=\min}$ is the latitude in the northern hemisphere where the subtropical minimum in precipitation occurs (typically somewhere between 15–30°N in these aqua planet simulations). Because the models are hemispherically symmetric, we integrate only in one hemisphere, otherwise ϕ_P would be zero in all of the models. The precipitation-weighted latitude effectively captures the spread of precipitation around the equator. The spread in ITCZ contractions is about 0.5–3.5° latitude across the models (see Table 3.2). ϕ_P in the ACRE-on simulation is uncorrelated with $\Delta\phi_P$ (the difference between ACRE-on and ACRE-off simulations).

Figure 3.2 shows the zonal mean ACRE values for the six COOKIE models. There is a very strong similarity between ACRE and precipitation; not surprising since both are linked by convection. ACRE is positive for all of these models, meaning the clouds are heating the atmospheric column in total. Positive ACRE is expected for tropical high clouds (see Slingo and Slingo, 1988), and Fig. 3.3 (left column) shows that high clouds dominate at low latitudes for these models. Figure 3.3 also shows how the cloud fraction responds to the cloud-radiative interactions (right column). Despite only four of the models reporting cloud fraction on the vertical grid, we see a lot of disparity between the model cloud responses. In general, there tend to be more clouds at low latitudes with ACRE-on, but there are a lot of changes in vertical structure that are not consistent between models. It is beyond the scope

of this paper to address the vertical structure of changes in cloud amount, but this should be investigated in future work.

Our hypothesis that ACRE contracts the ITCZ equatorward builds on the CAPE argument presented in Landu et al. (2014). The fact that some models use convective quasi equilibrium (CQE) in their convective parameterizations should not preclude the use of CAPE as a guide for the ITCZ contraction in these models. CQE, as described by Arakawa and Schubert (1974), is based on a plume model, which makes use of buoyancy. CAPE is similar to column-integrated buoyancy, so it should be an effective metric for where GCMs favor tropical convection. Landu et al. (2014) proposed that high CAPE values away from the equator favor a double ITCZ, while high CAPE values constrained near the equator favor a single ITCZ. We hypothesize that the equatorward contraction in the ITCZ is driven by changes in CAPE due to the cloud radiative heating. Specifically, the increase in upper tropospheric temperature associated with the cloud radiative heating strengthens the stability and weakens CAPE across the tropics. Therefore, we next verify that the upper tropospheric temperature increases from the ACRE-off to the ACRE-on simulations, despite the fixed sea surface temperatures (see Fig. 3.4).

As expected, the tropical upper troposphere is warmer in the ACRE-on simulations than in the ACRE-off simulations across all models because of the strong radiative warming of the high clouds. The surface air temperatures do not show much change, which is also to be expected considering that the sea surface temperatures are fixed in all of the simulations. The stratosphere is cooler in the ACRE-on simulations compared to the ACRE-off simulations. This may be due to high clouds trapping longwave fluxes emitted by the lower troposphere from making it into the stratosphere and causing warming, or it may be due to an increase in the Brewer-Dobson circulation. Indeed the radiative cooling is weaker in the tropical stratosphere with ACRE on (not shown) and the globally averaged stratospheric temperature is also colder with ACRE on, suggesting that there is a radiative impact. However, it is also

the case that the high latitude stratosphere is warmer (not shown), suggesting that there is also a dynamical component. Thus, it is likely that both the radiative and dynamic effects are important for determining the impact of ACRE on the stratosphere, but it is beyond the scope of this paper to quantify these effects. The atmospheric cloud radiative effect also acts to lower the tropopause temperature (not shown). Later we will show that ACRE has a strong latitude gradient from heating in the tropics to cooling in high latitudes, which would drive a stronger circulation.

Cloud-radiation interactions warm the upper tropical troposphere more than lower in the atmosphere and therefore they increase the stability. The ACRE-driven change in stability can be easily quantified as a change in Convective Available Potential Energy (CAPE). As noted above, CAPE is an attractive measure of how favorable the environment is for convection. We make use of George Bryan’s script to calculate CAPE (freely available from his webpage: <http://www2.mmm.ucar.edu/people/bryan/>; see also Bryan and Fritsch, 2004). When the cloud radiative effects are turned on, two things occur: non-zero values of CAPE occur at lower latitudes than in the ACRE-off configuration in all models; and CAPE decreases throughout the whole tropics in all models except CNRM (see Fig. 3.5). Sherwood et al. (1994) also found that CAPE decreased when cloud radiative effects were included in a climate model.

Temperature changes are not the only means by which CAPE can be changed. Increases in moisture within the boundary layer will increase CAPE, all else being equal. We examine the changes in specific humidity in Fig. 3.6. We find that including the cloud radiative heating increases the water vapor content of the tropical troposphere. The increase is most prominent right at the equator. Thus, we have competing influences. The clouds, through their radiative heating, act to stabilize the tropical atmosphere and decrease CAPE while at the same time increasing CAPE through increases in moisture — via increases in circulation strength as we will show later.

As shown in Fig. 3.5, including ACRE decreases CAPE for five of the six models examined in this study. That alone would suggest that the temperature increase due to ACRE is a stronger influence on the CAPE than the moisture increase, since increases in moisture would lead to an increase in CAPE, all else being equal. We want to know, however, whether the contracting CAPE profile is driven more by changes in temperature or water vapor. To answer this, we recalculate the CAPE using the temperature output from the ACRE-on experiment and the moisture output from the ACRE-off experiment, and vice versa. These terms allow us to compare changes in CAPE due to temperature or moisture alone. If temperature is driving the contraction of the ITCZ, we expect the meridional extent of the CAPE profile for “T on q off” to match that of ACRE-on and “T off q on” to match that of ACRE-off. If instead moisture is driving the contraction of the CAPE profile, the reverse should be true (i.e., the meridional extent of “T off q on” should match ACRE-on). Figure 3.5 shows temperature changes are the primary driver of the equatorward contraction of CAPE when the clouds are radiatively active. Note that while the meridional structure of the CAPE at low latitudes follows the moisture changes, the meridional extent follows the temperature changes. Since we are interested in the meridional extent of the precipitation, the meridional extent of CAPE is the important factor, and Fig. 3.5 shows this is determined by the temperature, not the moisture. Even if we scale all of the CAPE profiles to have a peak value of one (not shown), we arrive at the same conclusion that temperature is the dominant control on the meridional extent of the CAPE profile. The CNRM model has a different behavior than the other five models. Not only do the moisture changes between the ACRE-on and ACRE-off experiments in CNRM dominate the CAPE response, they also contribute some to the equatorward contraction of CAPE.

We attempted to identify a correlation between the magnitude of the cloud radiative heating in the ACRE-on experiments and the magnitude of the poleward movement of the ITCZ when the cloud radiative effect is removed, but no such relationship was significant. There

is only a weak correlation ($r=-0.6$) between the change in upper tropospheric temperature and the magnitude of the decrease in the precipitation-weighted latitude, so other factors (e.g. latent heat flux, wind speed, eddy fluxes, and mean-state temperature and humidity) must play a non-trivial role for determining any given model's exact ITCZ position. This is, of course, not surprising.

We have shown that the cloud radiative heating is responsible for keeping the ITCZ closer to the equator than it would be without that cloud radiative heating, but why is it closer? Why does the ITCZ not simply match the CAPE profile in width and structure? We will expand on the hypothesis presented in Landu et al. (2014) suggesting that the convective bands prevent moisture convergence equatorward of their latitude. In other words, the moisture convergence from the subtropics to the tropics only penetrates as far as the ITCZs. The equation for the moisture budget is

$$\frac{\partial w}{\partial t} + \nabla \cdot \frac{1}{g} \int_0^{p_s} q \vec{v} \, dp = E - P \quad (3.2)$$

where g is the gravitational acceleration constant, p_s is the surface pressure, $\vec{v}$ is the horizontal velocity, w is the precipitable water, q is the specific humidity, E is the surface evaporation, and P is the surface precipitation. The second term on the right is the moisture divergence. From Landu et al. (2014), we expect the moisture convergence to be weaker at the equator when the cloud radiative effects are turned off, and indeed that is the case. As expected, the changes in moisture convergence are strikingly similar to those seen in precipitation above (not shown). We have compared our calculations of moisture convergence using monthly data to those using daily data for the models that have daily data available and find little differences between them.

It is common in the literature for changes in surface fluxes to be responsible for switching between single and double ITCZs within models (e.g. Numaguti, 1993; Barsugli et al., 2005; Möbis and Stevens, 2012; Liu et al., 2010). While we have already shown evidence suggesting

that it is the change in temperature, not surface moisture, that leads to the equatorward contraction of high CAPE values and subsequently precipitation, it is worth examining the changes in surface latent heat flux. The models here employ fixed sea surface temperatures and use the bulk aerodynamic formula to calculate surface evaporation, so they are decoupled from the surface radiation. Therefore, the changes in surface evaporation are driven by changes in the circulation and the need for total precipitation to balance the total atmospheric radiative cooling, not changes in absorbed shortwave at the surface. Despite fixed SSTs and an enhanced mean meridional circulation owing to the cloud radiative heating, the latent heat flux does not show a consistent increase from ACRE-off to ACRE-on (Fig. 3.7). In fact, in three of the models, LHF is consistently weaker for ACRE-on compared to ACRE-off. The other three models have a more complicated meridional structure to their LHF differences. Of course, the increase in horizontal moisture convergence with ACRE is more important for the precipitation, but it is interesting that the latent heat flux fights against that increase. This conflict between moisture convergence and latent heat flux suggests that the zonal winds are slowing down, there is a decrease in eddy-driven gustiness, or there is an increase in boundary layer humidity that can account for the decrease in LHF when the cloud-radiation interactions are included. In short, the LHF often has its greatest decrease right along the equator when the cloud radiative effects are included, which should not encourage convection to move equatorward, and yet that is exactly what is observed. We show later that the mean meridional circulation intensifies with cloud radiative heating included, so the decrease in LHF is more likely due to an increase in low-level humidity (see Fig. 3.6).

Another possible factor that could influence the upper tropospheric temperature is the boundary layer Moist Static Energy (MSE) in the regions of active convection. The increase in surface moisture near the equator when cloud-radiation interactions are turned on compared to when they are off drives a similar increase in the surface MSE (see Fig. 3.8). Changes in surface air temperature are small due to the fixed sea surface temperature. We

have seen above that latent heat flux cannot explain the increase in moisture; thus it is the moisture convergence and delayed onset of convection consistent with the “upped-ante” mechanism driving the magnitude and structure of the surface MSE in the tropics in these models. The increase in surface moisture when cloud-radiation interactions are included supports the idea of enhanced latent heat release in the upper troposphere described above. The increase in boundary layer humidity likely changes the equilibrium of the boundary layer in these models and may introduce additional changes in the low clouds. The low cloud changes may, in turn, feed back on the circulation (refer to Fermepin and Bony, 2014, for details on the impact of low-cloud ACRE on circulation changes). Since the surface air temperature changes little, only the increase in mass flux owing to ACRE is capable of enhancing the vertical sensible heat flux. For most of the models, the difference in surface MSE between ACRE-on and ACRE-off is smaller than the change in upper-tropospheric MSE between ACRE-on and ACRE-off, as expected from the changes in CAPE. The only model that shows a larger change in surface MSE than upper-tropospheric MSE is the CNRM model, which is also the only model that showed an increase in CAPE with ACRE included.

3.3.2 Assessing changes in circulation

As mentioned above, the change in moisture is primarily the result of circulation changes. ACRE strengthens and focuses the mean meridional mass stream function in all of the models (see Fig. 3.9), despite simultaneously increasing the stability. There is a substantial increase in mean ascent (on the order of tens of hPa/day) in the convective regions associated with their narrowing owing to the cloud radiative heating. The subsidence change is minimal at the subtropical edge of the Hadley cell, such that the changes in mass flux are constrained to between roughly 15 S – 15 N. It is expected that the diabatic heating from clouds absorbing longwave radiation has a strong control on the circulation strength, but we want to know whether clear-sky radiatively-driven subsidence (ω_{rad} ; see Kuang and Hartmann, 2007;

Zelinka and Hartmann, 2010; Harrop and Hartmann, 2012) may contribute to the increase in circulation strength as well. Radiatively-driven subsidence is calculated as follows:

$$\omega_{\text{rad}} = \frac{Q_{\text{rad}}}{\sigma}, \quad (3.3)$$

where Q_{rad} is the clear-sky radiative cooling and σ is the static stability. The clear-sky radiative heating terms are not standard output for the COOKIE simulations, thus clear-sky radiative heating was calculated using the RRTMG model described for the cloud-resolving model in the section 3.2. ACRE enhances the clear-sky radiative cooling in the upper troposphere ($p < 500$ hPa), owing to the increase in temperature and moisture at those levels (not shown). ACRE diminishes clear-sky radiative cooling in the mid-troposphere ($700 < p < 500$ hPa) for most of the models, owing to the increases in moisture above those levels. ACRE enhances radiatively-driven subsidence in the upper troposphere ($100 < p < 300$ hPa) but diminishes it in the mid-troposphere (see Fig. 3.10). The enhancement in radiatively-driven subsidence is a combination of both changes in the radiative cooling and in the static stability, while the reduction results from the cloud radiative heating increasing static stability compared to when ACRE is removed. The circulation change does not have a strong vertical structure to it, so changes in clear-sky radiative cooling must not drive the circulation response to ACRE. Thus, it is more likely that the direct diabatic heating from convection, not the change in clear-sky radiation, contributes to the change in circulation strength between ACRE-on and ACRE-off.

The distribution of vertical velocities in the COOKIE models depends on the cloud-radiation interactions as well. The Probability Density Functions (PDFs) of vertical velocity at 500 hPa are shown in Fig. 3.11 for each model. ACRE increases the frequency of strong upward vertical velocities ($\omega_{500} < -100$ hPa/day), while also decreasing the frequency of weak upward vertical velocities ($0 < \omega_{500} < -75$ hPa/day). Because these are monthly outputs on large grid sizes, the increase in strong upward velocities for ACRE-on could be

due to the convective updraft speeds increasing, the frequency of convection within that grid increasing, or a larger area fraction of the grid box being occupied by convection. We will revisit these possibilities later when discussing results from the cloud-resolving model.

The precipitation is enhanced near the equator, but the total precipitation across the tropics (measured as the area between 30S – 30N; see Table 3.2) is less with the cloud radiative effects on owing to the weaker radiative cooling compared to the ACRE-off simulations. Thus, even though the mean circulation is more vigorous with cloud-radiation interactions turned on, there is still less precipitation. The distribution of precipitation, however, is sensitive to the cloud radiative heating and resembles the changes in vertical velocity (see Fig. 3.12). The ACRE-on simulations have much more frequent heavy precipitation (defined here as rain rates exceeding 20 mm/day) consistent with the upped-ante mechanism of Neelin et al. (2003) whereby convection is suppressed until the surface MSE is higher.

We suggest that the increase in circulation strength is due to the equator-to-pole atmospheric heating gradient being amplified by ACRE. ACRE is positive in the tropics and negative in the extra-tropics, requiring greater poleward energy transport (see figures 3.2 and 3.13). Figure 3.13 shows the multi-model mean, zonal mean ACRE and poleward heat transport for the ACRE-on experiments (left), as well as the difference in these two quantities between ACRE-on and ACRE-off (right). Note that ACRE is zero in the ACRE-off experiments by construction, so ACRE appears the same in both panels of Fig. 3.13. The poleward energy transport is calculated indirectly using the top-of-atmosphere and surface fluxes, both radiant and turbulent. The poleward energy transport is calculated separately for both ACRE-on and ACRE-off, and thus accounts for changes both in ACRE as well as latent heat flux at the surface. Figure 3.13 shows that ACRE enhances the poleward energy transport by nearly a petawatt at some latitudes — the integrated increase in poleward energy transport is roughly 15%. Because the sea surface temperatures are held fixed, the surface fluxes do not need to be balanced; thus, the ACRE gradient could be offset by

changes in surface fluxes. This does not appear to be the case in these simulations, however, and instead the atmospheric circulation intensifies to transport additional energy poleward. It is likely that since ACRE is primarily driven by longwave heating, in the tropics we would not expect ACRE heating to be taken up by the surface. Therefore, even in a coupled simulation, one might still expect an increase in poleward energy transport by the atmosphere in response to ACRE. The shortwave effect at the surface would be expected to act in the exact opposite direction to the longwave atmospheric heating, requiring weaker poleward heat transport by a combination of the atmosphere and ocean. Peters and Bretherton (2005) suggest that the atmospheric circulation response to the cloud radiative effects is negligible owing to the near cancellation of shortwave and longwave cloud radiative effects. They argue that the surface cooling from reduced shortwave absorption is offset by an equivalent reduction in latent heat flux. Note that there is a strong enhancement of poleward energy transport in the mid-latitudes consistent with an increase in eddy kinetic energy as described by Li et al. (2015).

In the original six models used for COOKIE, we see a consistent increase in integrated cloud water path when the cloud radiative effects are included (see Fig. 3.14). If we integrate the cloud water path over the tropics (30 S – 30 N), we see a consistent increase across the models ranging from 14 – 56 g/m². As noted above, over the same region, precipitation decreases when cloud radiative effects are included. A decrease in precipitation with a simultaneous increase in cloud mass would suggest a decrease in precipitation efficiency. Across the tropics, the ratio of precipitation to cloud water path increases from the ACRE-off to the ACRE-on configurations in all models. The decrease in precipitation efficiency occurs even in the deep convective regions.

3.3.3 Investigating a transient experiment

As noted above, a transient experiment was performed using the GFDL AM2.1 model in aqua planet mode. We use the “Qobs” sea surface temperature pattern as before and the SSTs remain fixed throughout the simulation. We have a two year climatology run, followed by an abrupt removal of the cloud radiative effects. The radiative relaxation time of the upper tropical troposphere is on the order of a few weeks (Hartmann et al., 2001a), so we expect the transition period from one equilibrium state to the other to occur on a similar timescale. Therefore, we use daily data in the transient experiment to capture the changes. From the hypothesis above, we expect that after the clouds are removed, the temperature in the upper tropical troposphere will begin to decrease while the ITCZ will begin moving poleward. To measure these two impacts, we will use the temperature at 224 hPa (roughly the outflow level for tropical deep convection in the real tropical atmosphere, see Houze and Betts, 1981) and the precipitation-weighted latitude. Within the model, substantial detrainment of deep convection (measured by divergence) occurs at the 157 hPa level as well. Our conclusions are not sensitive to the choice of level, so we will focus only on the 224 hPa temperature. We have chosen to average the temperature at 224 hPa over 15 S – 15 N. While the magnitude of the temperature change varies across these latitudes, we find that the zonally-averaged time rate of change in temperature is the same across 15 S – 15 N (not shown). For simplicity, we will refer to the temperature at the 224 hPa pressure level as T_{224} .

Figure 3.15 shows the time series of T_{224} and ϕ_P during the one hundred days before and after the ACRE-on to ACRE-off switch. We can see from Fig. 3.15 that the transition from the ACRE-on to the ACRE-off state occurs rapidly. The mean in T_{224} drops from 224 K to 222 K. The mean in ϕ_P increases from 7.92° to 9.43° when ACRE is turned off. Despite some low frequency variations in the precipitation signal, we see the simulation reaches its new equilibrium values for T_{224} and ϕ_P within about forty days. The main conclusion from the transient experiment is that the decrease in upper tropospheric temperature owing to the

removal of cloud-radiation interactions accompanies the poleward broadening of the ITCZ, as expected. T_{224} and ϕ_P are significantly correlated over the first forty days after the cloud radiative effects are removed, but the correlation drops off rapidly after that period such that T_{224} is not a good predictor of the oscillations seen in ϕ_P .

3.3.4 Further testing with a cloud resolving model

We have also tested our findings within a cloud resolving model with two different microphysical configurations (described in section 3.2). The main difference between the SAM and NA5 microphysics is that the NA5 microphysics produces substantially more high, thin-to-medium optical thickness clouds, in better agreement with satellite observations (Lopez et al., 2009), which exert a greater radiative heating in the atmosphere. As such, the inclusion of the cloud radiative heating increases the domain average atmospheric radiative heating by a lot more in the NA5 configuration than the SAM configuration (7.2 W/m^2 compared to 0.2 W/m^2). While both of these changes are statistically significant at the 95% level, because the change in radiative heating is so small in the SAM configuration, there is no statistically significant change in domain-average precipitation for the SAM configuration. The NA5 configuration shows a statistically significant decrease in domain-average precipitation of 0.17 mm/day . Additionally, Fig. 3.16 (top row) shows that for the NA5 configuration, the spatial extent of precipitation narrows over the warm pool part of the simulation with clouds on, similar to the ITCZ moving closer to the equator in the COOKIE simulations. The response in the SAM configuration is a slight shift of the precipitation without any change in magnitude. It is also worth noting that the mean climate is not the same between the two microphysical configurations in the CRM. The microphysical parameterizations make the temperature higher in the SAM configuration than in the NA5 configuration for ACRE-off. However, the SAM configuration is colder than NA5 for ACRE-on, as expected, owing to stronger cloud radiative heating (not shown). Similar to the COOKIE simulations, the upper

tropospheric temperatures are warmer with ACRE-on than ACRE-off for both microphysics configurations in the CRM. The warming in the SAM configuration is smaller than that of NA5, owing to less high cloud in the SAM configuration.

While the mean atmospheric radiative heating changes little between the ACRE-on and ACRE-off simulations in the SAM configuration (see Fig. 3.16, middle row), there is weaker cooling in the convective zones owing to the prevalence of high clouds and stronger cooling in the subsidence zones owing to the prevalence of low clouds there. Thus, the balance between high clouds trapping emissions and low clouds enhancing emissions causes the weak mean response of the radiative heating profile in the SAM configuration. For the NA5 configuration, the high clouds dominate the response, giving the distinct increase in radiative heating in ACRE-on compared to ACRE-off.

The mean vertical velocity at 500 hPa (w_{500}) in the region of convection is also greater with cloud-radiation interactions present than when they are removed, which agrees with the results from the GCMs. The change in w_{500} between the ACRE-on and ACRE-off simulations closely resembles the change in precipitation (not shown). Again, as in the GCMs, we see a similar circulation intensification (reflected in the precipitation changes) from ACRE-off to ACRE-on. The heating of the upper atmosphere by clouds stabilizes the atmosphere, but at the same time the gradient in that heating requires the circulation to intensify. The atmosphere responds by contracting precipitation into a narrower, more intense band over the warmest SSTs. The distribution in w_{500} for the CRMs, however, bears little resemblance to the distribution of w_{500} seen in the GCMs. The difference between the CRM and GCMs is not surprising. The distribution of w_{500} in the CRM shows the frequency of individual updraft speeds, while the distribution of w_{500} in the GCMs is an aggregation, both in time and space, of numerous updrafts. As an example of this discrepancy, the increase in w_{500} with ACRE-on for the GCMs, could be due to the convective updraft speeds increasing or it could be more frequent updrafts with the same speed as in the ACRE-off configuration.

Looking at the CRM, the distribution of w_{500} hardly changes at all between the ACRE-on and ACRE-off configurations for the SAM microphysics, while for the NA5 microphysics, the ACRE-on configuration favors weaker updrafts and downdrafts compared to the ACRE-off configuration (see Fig. 3.16, bottom row). The CRM shows that the vertical velocity of individual convective plumes is not sensitive to ACRE, and does not respond in the same fashion as the large scale circulation (as evidenced by the GCMs). If anything, the inclusion of cloud radiative heating weakens the convective updraft speeds.

For both microphysical configurations, the liquid water path (LWP) increases for ACRE-on, while at the same time the ice water path (IWP) decreases. The SAM configuration is roughly 50/50 liquid and ice, while the NA5 configuration is roughly 25/75 liquid and ice. The increase in total cloud water path (CWP) in the SAM configuration is small (+0.77 g/m²) owing to the offsetting changes in liquid and ice. In NA5, however, CWP actually decreases (-9.2 g/m²) owing to the relatively stronger influence of ice. In the GCMs, both LWP and IWP increase in all models across the tropics (not shown). The increase in LWP is greater than the increase in IWP in all of the GCMs as well.

We performed another set of experiments with the CRM where we used a slab ocean with fixed ocean heat transport in a sinusoidal pattern with an amplitude of 25 W/m² and a mean value equal to the top of atmosphere radiation imbalance. The model was run into equilibrium with the cloud radiative heating included. We then took those sea surface temperature distributions (one for each microphysics configuration) and ran the ACRE-on and ACRE-off experiments as before. In this case, we still have a pair of fixed SST simulations that we can compare with the COOKIE experiments. Experiments such as these are useful because they use a “natural” sea surface temperature distribution — natural in the sense that it represents a model-selected climate mode that has the atmosphere and ocean circulations in equilibrium with one another. The resulting changes between the ACRE-on and ACRE-off experiments are similar to those shown above with the fixed sinusoidal SST pattern described

in section 3.2. Therefore, our results from this work with the CRM are robust between the differing SST patterns.

We may expect that the strength of the precipitation contraction may also be sensitive to whether we use 2D or 3D simulations. We have chosen the 2D setup for its simplicity and cost efficiency. While 3D simulations tend to have a different organization pattern for convection compared to 2D simulations, we anticipate that both will produce the vertical heating gradient needed to test our CAPE argument. The magnitude of that heating gradient is likely to change between 2D to 3D experiments, but we do not expect 3D effects to change our basic qualitative conclusions. We have no rotation in our simulations to simplify the problem. Since the CAPE mechanism does not rely on rotation, our results should not be sensitive to whether rotation is included or not.

In short, the cloud-resolving model allows us to investigate whether the conclusions drawn from the COOKIE simulations are sensitive to the GCM framework. While all of the GCMs show a contraction of precipitation over the warmest SSTs when cloud radiative heating is included, the CRM only shows this behavior when high, thin-to-medium optical thickness clouds are prevalent. While the GCMs show a consistent increase in vertical velocity for ACRE-on, the CRM does not. The differences in vertical velocity between the GCMs and the CRM may be sensitive to the inclusion of rotation in the GCMs and the absence of rotation in the CRM. Future research will be needed to assess the combined role of rotation and cloud-radiative heating on vertical velocity. In fact, the response of vertical velocities to cloud radiative heating in the CRM depends on which microphysics is used: vertical velocities slow down in NA5, while they are insensitive to cloud radiative heating in SAM. We cannot determine whether the GCMs are simply showing the effect of more frequent convection in areas where convection is already prevalent or if they are in disagreement with the CRM. Finally, while the CWP increases robustly in the GCMs, the response of CWP to cloud radiative heating in the CRM shows a strong dependence on microphysics. The

main difference between the GCMs and the CRM is in the response of ice water path to cloud radiative heating. The GCMs show IWP increases with cloud radiative heating, while the CRM shows that IWP decreases with cloud radiative heating. Future research will be needed to clarify this difference.

3.4 *Conclusions*

In aqua planet simulations with fixed SSTs, we find the atmospheric cloud radiative effect (ACRE) results in an equatorward contraction of the ITCZ, and a substantial reduction in the double ITCZ problem in all models. The equatorward contraction in the ITCZ is consistent with the contraction of the tropical CAPE maximum (consistent with the hypothesis of Landu et al., 2014). Cloud radiative heating in the tropical upper troposphere increases the temperature there, increasing stability and decreasing CAPE (in agreement with Sherwood et al., 1994). While the atmospheric cloud radiative effect is not an effective proxy for how far the ITCZ contracts, the contraction is nonetheless robust across all models despite their differences in resolution, dynamics, and parameterizations. We have also shown that the increase in upper tropospheric temperature from the atmospheric cloud radiative effect smoothly transitions with the ITCZ in a transient simulation from the ACRE-on to the ACRE-off configuration. Our results show a robust ITCZ response across a wide range of model parameters, though none of them consider realistic land-sea geography. The equatorward contraction of the precipitation envelope when including cloud radiative heating agrees with Randall et al. (1989) and with Liu and Moncrieff (2008).

Regarding our use of CAPE (following Landu et al. 2014) as a metric for explaining how the cloud-radiative heating forces a contraction of the ITCZ, we note that CAPE is effectively an integrated measure of stability of the atmosphere. Precipitation forms where the atmosphere is most unstable to convection, i.e. where CAPE is highest. While CAPE depends on both temperature and moisture, we have already shown that it is the temperature

increase that results in the contraction of the ITCZ in these models. We argue that the cloud-radiative heating is responsible for the increase in upper atmospheric temperature. The other major heating term, latent heat release, actually decreases in the tropical average when cloud-radiative heating is included owing to a decrease in the radiative cooling of the atmosphere.

Changes in surface turbulent fluxes have been cited in the literature (e.g. Numaguti, 1993; Barsugli et al., 2005; Möbis and Stevens, 2012; Liu et al., 2010) as a cause for a change from single to double ITCZ within models. Our results suggest a change in latent heat flux is not responsible for the equatorward contraction of the ITCZ due to ACRE. In fact, in some of the models, the strongest decrease in LHF occurs on the equator, which would not be expected to draw the ITCZ equatorward. Evaporation changes are likely still important for determining the climatology of the ACRE-on simulations, and may even contribute to the magnitude of the equatorward contraction of the ITCZ when cloud-radiation interactions are included, but they are not responsible for the robust ITCZ contractions seen here. Changes in surface turbulent fluxes do offset changes in atmospheric radiative heating, as expected.

The models further show a robust increase in precipitation in convective regions in the ACRE-on compared to ACRE-off configurations, confirming results from prior studies (e.g. Slingo and Slingo, 1988, 1991; Randall et al., 1989). Across the whole of the tropics, however, the total precipitation decreases to balance the reduction in radiative cooling when the cloud-radiation interactions are present. The cloud radiative effects enhance the mean meridional circulation in all models, again confirming prior studies (e.g. Ramanathan, 1987; Slingo and Slingo, 1988, 1991; Stuhlmann and Smith, 1988a,b; Randall et al., 1989; Sherwood et al., 1994; Zhang and Rossow, 1997; Sohn, 1999; Bergman and Hendon, 2000b,a; Raymond, 2000; Tian et al., 2001; Tian and Ramanathan, 2002, 2003; Lee and Yoo, 2014; Del Genio and Chen, 2015; Crueger and Stevens, 2015). Tian and Ramanathan (2003) showed that a realistic ACRE distribution can recreate quantitatively realistic Hadley and Walker circulations. Also,

we have shown that the clear-sky radiative cooling cannot be responsible for the circulation increase. We therefore suggest that it is the enhanced meridional gradient in diabatic heating caused by ACRE (heating in the tropics and cooling in the extra-tropics) that is responsible for the circulation intensification. Despite the decrease in precipitation and increase in stability in the tropics, the enhanced energy flux poleward requires the circulation to speed up. Further research into this area needs to be done to see how robust this mechanism is.

The clouds also act to partition more of the net divergence of energy transport into the atmosphere instead of the ocean (Zhang and Rossow, 1997; Tian et al., 2001). Removing cloud radiative heating while keeping SSTs fixed increases the implied zonally-averaged values of ocean heat transport up above 100 W/m^2 along the equator (depending on the model), which is more characteristic of the cold tongue than the actual tropical mean in the real world. While fixed SSTs are a useful experiment for isolating the impact of the cloud radiative heating on the circulation, it is worth noting that the total climate response to removing cloud-radiation interactions would change in coupled atmosphere-ocean models, particularly since we expect the shortwave effect of clouds to have a strong influence on SST.

Not only does cloud radiative heating reduce tropics-wide precipitation, it also changes the distribution of precipitation intensities to favor heavier precipitation (rates exceeding 20 mm/day). Additionally, we have quantified the changes in the vertical velocity distribution across the various GCMs as well as within a pair of cloud resolving model simulations. The GCMs show a robust increase in both stronger precipitation rates and updraft speeds when cloud radiative heating is included. The CRM simulations, however, suggest that the frequency of strong updraft speeds does not actually increase in response to cloud radiative heating. Of course, a Hadley cell on a spherical Earth is different from the 2-D non-rotating circulation in the CRM, so the geometry and rotation of the GCMs may factor into the differences in vertical velocity as well. Future research should look to determine whether the absorption of radiation by clouds is capable of enhancing the updraft speed within individual

convective storms.

Cloud radiative heating and its subsequent enhancement of the circulation increase column-integrated cloud liquid and ice mass (the cloud water path: CWP). The decrease in precipitation and increase in cloud mass suggests that cloud radiative effects reduce precipitation efficiency, meaning that more high clouds are generated per unit of precipitation, including regions of deep convection. Both the liquid and ice increase across the tropics in all of the GCMs, with liquid increases being the larger component of the CWP increase. In the CRM simulations, however, only liquid water path increases in the presence of cloud radiative heating. Ice water path, shows a decrease from the ACRE-off to ACRE-on configuration in the CRM. The total change in cloud water path is sensitive to the relative abundance of liquid and ice in the mean climate, which is sensitive to the microphysical parameters. Future research should investigate the response of ice abundance to cloud radiative heating.

It is worth taking a moment to elaborate on the differences between our results and those of a few similar studies that had contrasting results. First, Raymond (2000) used a simple beta plane model with parameterized cumulus convection and he found that precipitation maximized over the warmest SSTs with and without cloud-radiation interactions included. The way he defines whether deep or shallow convection occurs is based on whether surface equivalent potential temperature exceeds some threshold. That threshold is defined as the interpolated value of saturated equivalent potential temperature taken where a surface lifted parcel has a total water mixing ratio (vapor + cloud condensate) equal to $1.1 \times$ the saturation mixing ratio. Thus, deep convection may be overly sensitive to surface moisture.

Second, Chao and Chen (2004) used aqua planet models with experiments both including and removing cloud-radiation interactions. Chao and Chen (2004) show that the position of the ITCZ within their model is not sensitive to whether the cloud-radiation interactions are active in their model, while our analyses show a robust equatorward movement of the ITCZ when cloud-radiation interactions are included. The difference is likely in the sea

surface temperature profiles. While the COOKIE experiments all use the “Qobs” sea surface temperature profile outlined in the Aqua-Planet Experiment, Chao and Chen (2004) use a globally-uniform sea surface temperature. We suspect that the simulations performed by Chao and Chen (2004) show a similar increase in upper tropospheric temperature when the cloud-radiation interactions are included. However, since the sea surface temperature has no gradient, there would be no equatorward contraction of the high CAPE region; instead CAPE would simply decrease where it was high before.

Third, Wofsy and Kuang (2012) used a cloud-resolving model to investigate the relationship between the precipitation area and radiative cooling and found that increased radiative cooling led to a narrowing of the precipitating region. In our results, the decrease in radiative cooling owing to the tropical clouds constrains the precipitation region to be closer to the equator — effectively narrowing it. The radiative cooling in the Wofsy and Kuang (2012) study was a fixed value throughout the depth of the troposphere though, so the CAPE-driven arguments resulting from the vertical distribution of cloud heating would not apply to their study.

Finally, Voigt and Shaw (2015) showed that changes in ACRE with global warming caused precipitation to contract in one aqua planet model, but expand in a different model, suggesting the role of ACRE on the meridional distribution of precipitation depends on the climate state. In their results, they break down the components of the precipitation changes into those driven by SST, ACRE, and water vapor radiative heating. They show that increasing SST increases precipitation everywhere in the tropics, but the increases are largest right along the equator, similar to a contraction of the ITCZ. The same can be seen in the vertical velocity field (see Fig. 2f in Voigt and Shaw, 2015); upward vertical velocities are enhanced right along the equator and decrease just off the equator. Because the temperature profile of the tropical atmosphere above the boundary layer is nearly moist adiabatic, the upper atmosphere warms more than the surface, creating a vertical temperature gradient similar

to that effected by ACRE. Therefore, in the case of their increasing SST experiments, their results are in line with our own. The response of precipitation to ACRE in the simulations examined by Voigt and Shaw (2015) is not due to the climatological ACRE profile, but the changes in meridional ACRE distribution brought about by warming. It is reasonable to expect that cloud feedbacks may not reflect a simple amplification of the climatological cloud patterns. While the model physics have all been developed to give reasonable mean climate states (in AMIP or CMIP type experiments, at least), their feedbacks are not nearly as well constrained. It is unclear whether the stability changes from the cloud feedbacks are the same between the two models examined by Voigt and Shaw (2015). Based on the results found here, it is likely that they are not. This same rationale applies to the circulation strength as well – cloud feedbacks that differ between models are also likely to differ on the response of the Hadley circulation to warming.

It is also worth mentioning other studies examining the impact of cloud radiative heating on the Walker circulation (e.g. Bretherton and Sobel, 2002; Sobel, 2003; Peters and Bretherton, 2005). There are enough parallels between the Hadley and Walker circulations to make this comparison useful. Bretherton and Sobel (2002) argue that cloud-radiative heating narrows the convective region in an idealized simulation of a Walker cell, which our results agree with. Work by Sobel (2003) and Peters and Bretherton (2005) showed the inclusion of the surface cloud radiative effect is shown to be important for the Walker circulation as well. Reducing shortwave absorption at the surface cools it, and reduces latent heat flux, which offsets the increase in column moist static energy from ACRE. Peters and Bretherton (2005) showed that the offsets between the surface and atmospheric cloud radiative effects led to negligible changes in convective area and circulation strength. These results highlight the importance of future simulations that are coupled such that both the atmospheric and surface cloud radiative effects may be examined.

As this work and previous work have shown, there are many processes important for

determining the climatological mean position of the ITCZ. This work emphasizes the role of local cloud radiative heating processes in constraining the ITCZ to the regions of maximum SST. Precipitation hews more closely to the maximum SST locations when cloud radiative effects are included. It is important to note that all of these simulations have been conducted with fixed SSTs, and more complicated feedback processes involving changes in the SST patterns may exist in coupled simulations. These results further highlight the importance of getting the cloud-radiation interactions correct in models — not only at the top of the atmosphere, but the amount of heating that occurs within the atmosphere is also of critical importance for the location and strength of precipitation and for atmospheric circulation.

Follow up research will look at cloud macrophysical responses (cloud area, cloud top temperature, etc.) to cloud radiative heating. Additional research will look into whether these effects are amplified or diminished in a warmer or CO₂ rich environment. The enhanced CO₂, for example, may reduce the amount of longwave absorbed by the clouds and limit ACRE. GCMs also tend to have a reduction in high cloud fraction at warmer temperatures, which would also reduce ACRE.

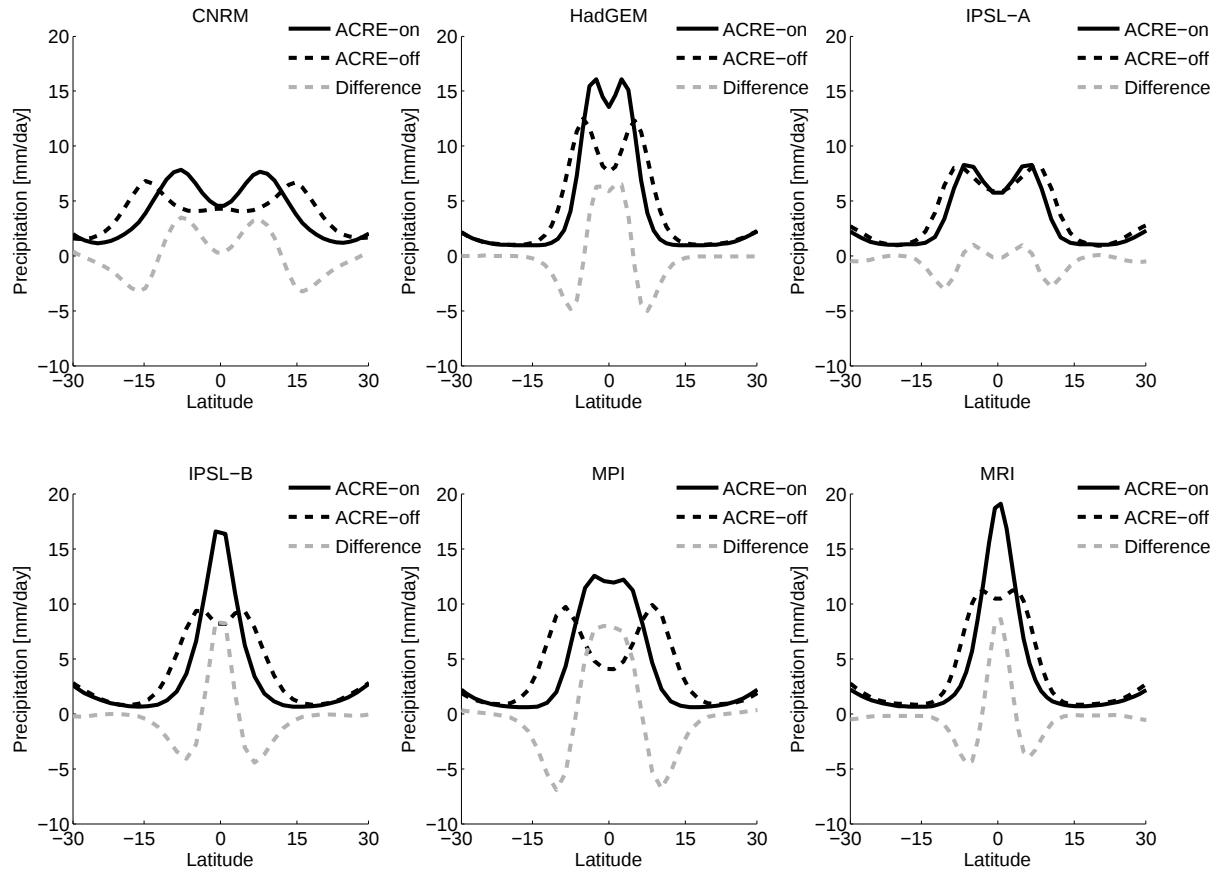


Figure 3.1: Precipitation for the six different models. The gray dashed line shows the difference between the ACRE-on and ACRE-off experiments.

Table 3.1: Model descriptions and details.

Modelling Center & Model name	Abbreviation for this study	Resolution (lon x lat)	Citations	Comments
Centre National de Recherches Meteorologiques (CNRM) CM5.1	CNRM	256×128 (1.41° × 1.40°)	Voldoire et al. (2013)	Not part of original COOKIE
Geophysical Fluid Dynamics Laboratory (GFDL) Atmospheric Model (AM2.1)	GFDL	144×90 (2.5° × 2°)	The GFDL Global Atmospheric Model Development Team (2004)	
Met Office Hadley Centre (MOHC) Hadley Global Environment Model 2 - Atmosphere (HadGEM2-A)	HadGEM	192×145 (1.875° × 1.25°)	Collins et al. (2008)	
Institut Pierre Simon Laplace (IPSL) Coupled Model (CM5A)	IPSL-A	96×96 (3.75° × 1.89°)	Dufresne et al. (2013), Hourdin et al. (2013a)	Physics package version A
Institut Pierre Simon Laplace (IPSL) Coupled Model (CM5A)	IPSL-B	96×96 (3.75° × 1.89°)	Dufresne et al. (2013), Hourdin et al. (2013b)	Physics package version B
Max Planck Institute for Meteorology (MPI-M) ECHAM6	MPI	192×96 (1.875° × 1.865°)	Stevens et al. (2013)	
Meteorological Research Institute (MRI) CGCM3	MRI	320×160 (1.125° × 1.12°)	Yukimoto et al. (2012)	

Table 3.2: Change in tropical precipitation (ACRE-on – ACRE-off), precipitation-weighted latitude for ACRE-on, and change in precipitation-weighted latitude (ACRE-on – ACRE-off) for the six COOKIE simulations and the additional run using the GFDL model. Δ Precipitation is calculated over 30S – 30N.

Model	Δ Precipitation [mm/day]	ACRE-on ϕ_P	$\Delta\phi_P$ (on – off)
CNRM	-0.061	9.58°	-3.29°
GFDL	-0.076	8.03°	-1.27°
HadGEM	-0.011	4.37°	-1.36°
IPSL-A	-0.51	7.23°	-0.44°
IPSL-B	-0.35	4.22°	-2.00°
MPI	-0.14	5.19°	-3.60°
MRI	-0.29	3.94°	-1.13°

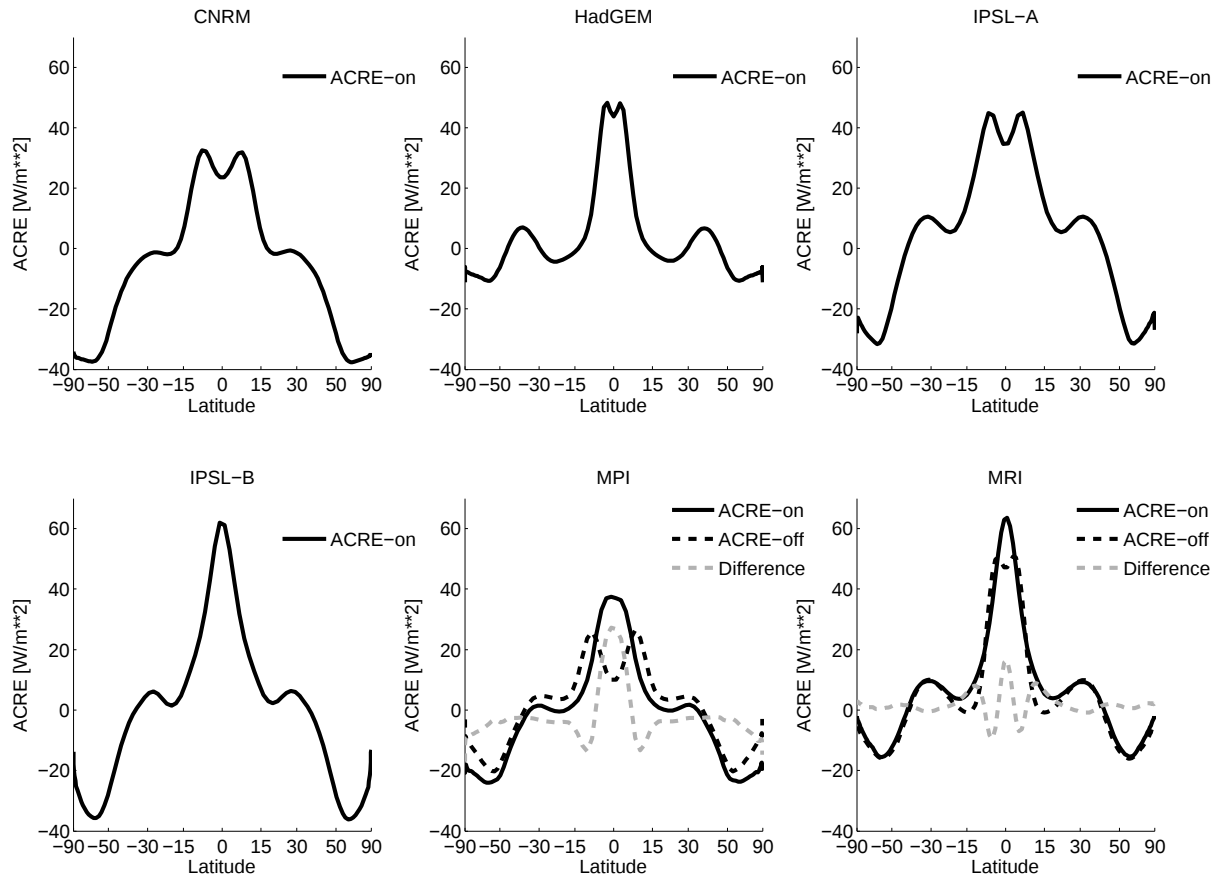


Figure 3.2: ACRE for the six different models. The gray dashed line shows the difference between the ACRE-on and ACRE-off experiments. Note that for ACRE-off, the ACRE values reported are the those that would occur if cloud-radiation interactions were included in the model. Only MPI and MRI report these ACRE-off values. Note that the latitude range for these figures extends to the poles.

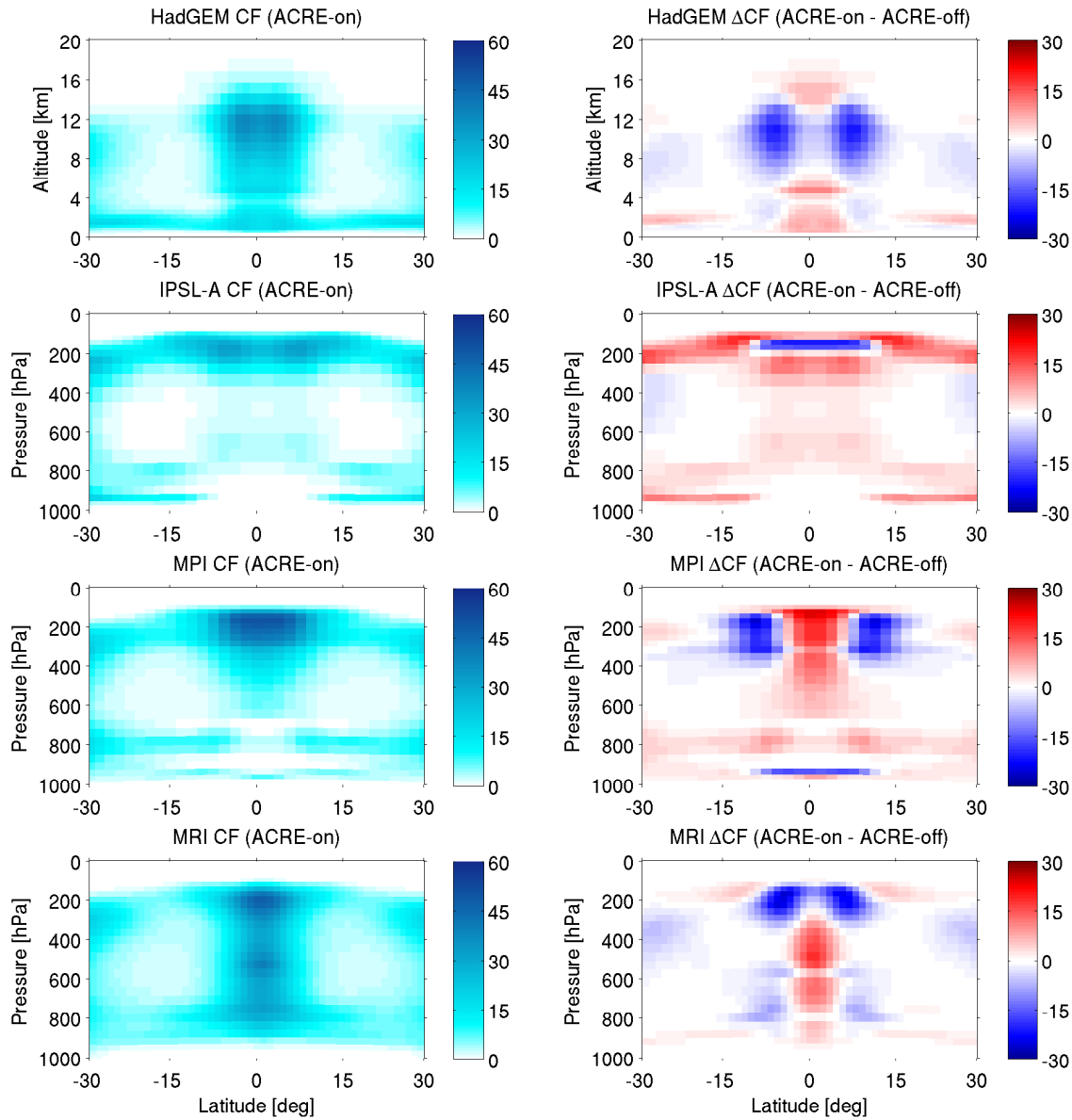


Figure 3.3: (Left column) zonal mean cloud fraction (CF) as a function of latitude and altitude/pressure for the four models that report it (HadGEM, IPSL-A, MPI, and MRI). (Right column) Differences in cloud fraction (ACRE-on – ACRE-off). Amounts and changes are given as percents.

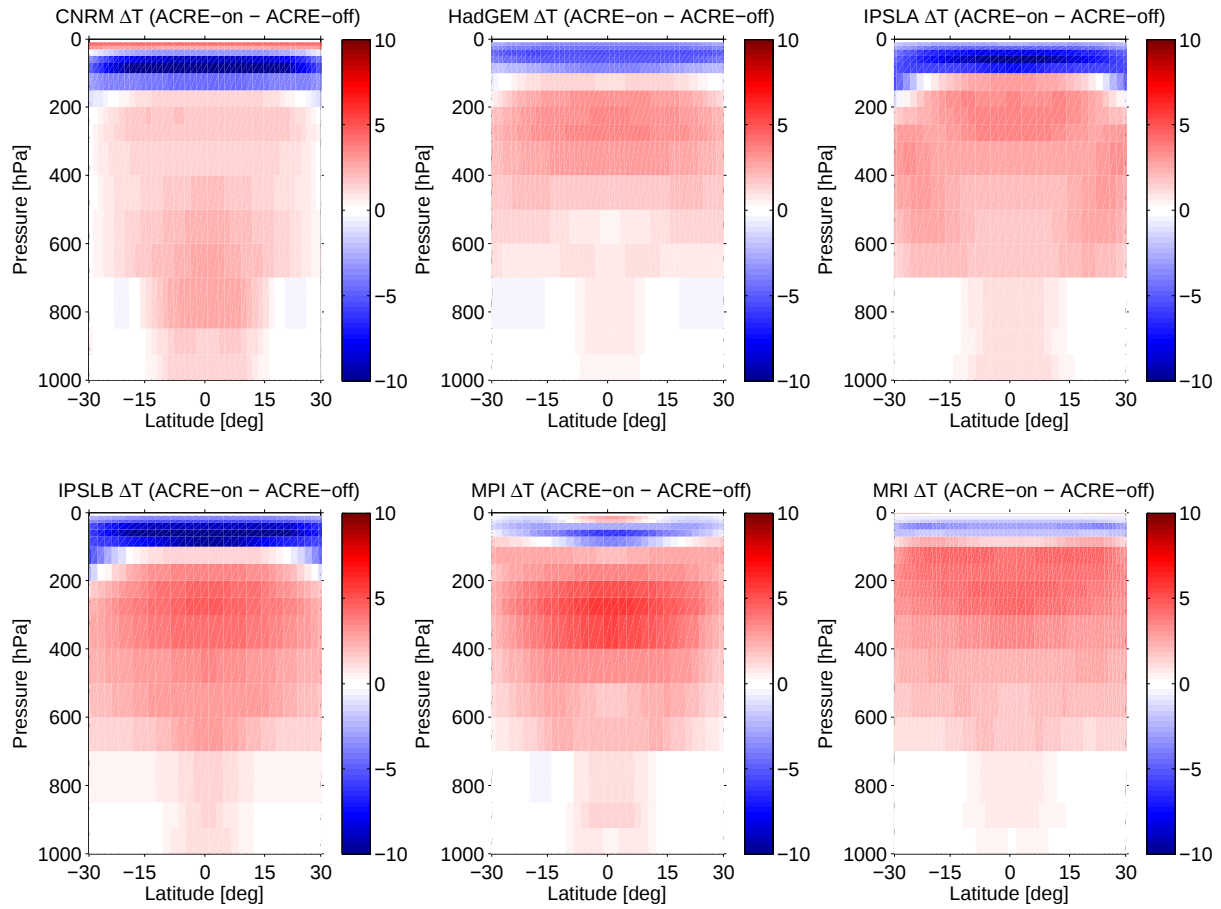


Figure 3.4: Difference in air temperature caused by the cloud radiative effect for all six COOKIE models.

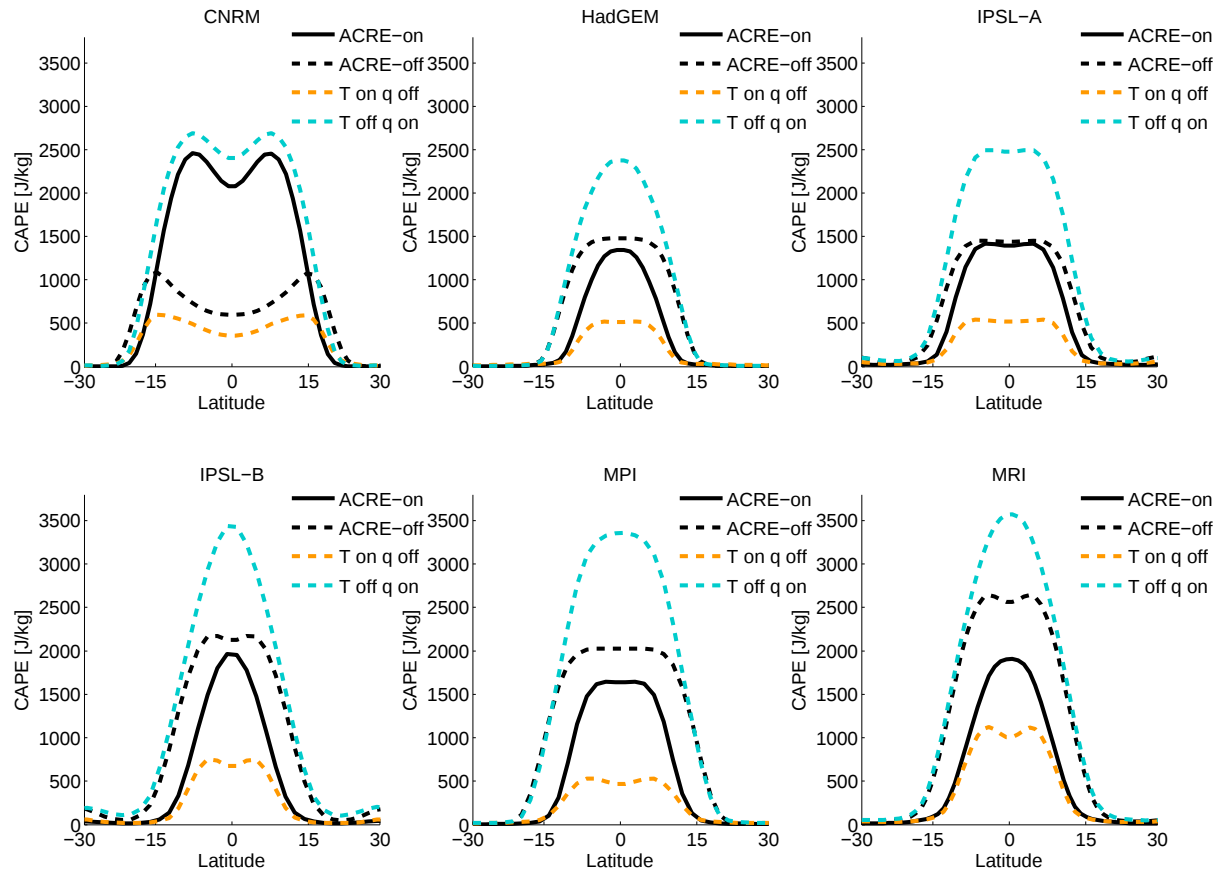


Figure 3.5: CAPE for ACRE-on T and q (solid black line), ACRE-off T and q (dashed black line), ACRE-on q and ACRE-off T (dashed teal line), and ACRE-on T and ACRE-off q (dashed orange line).

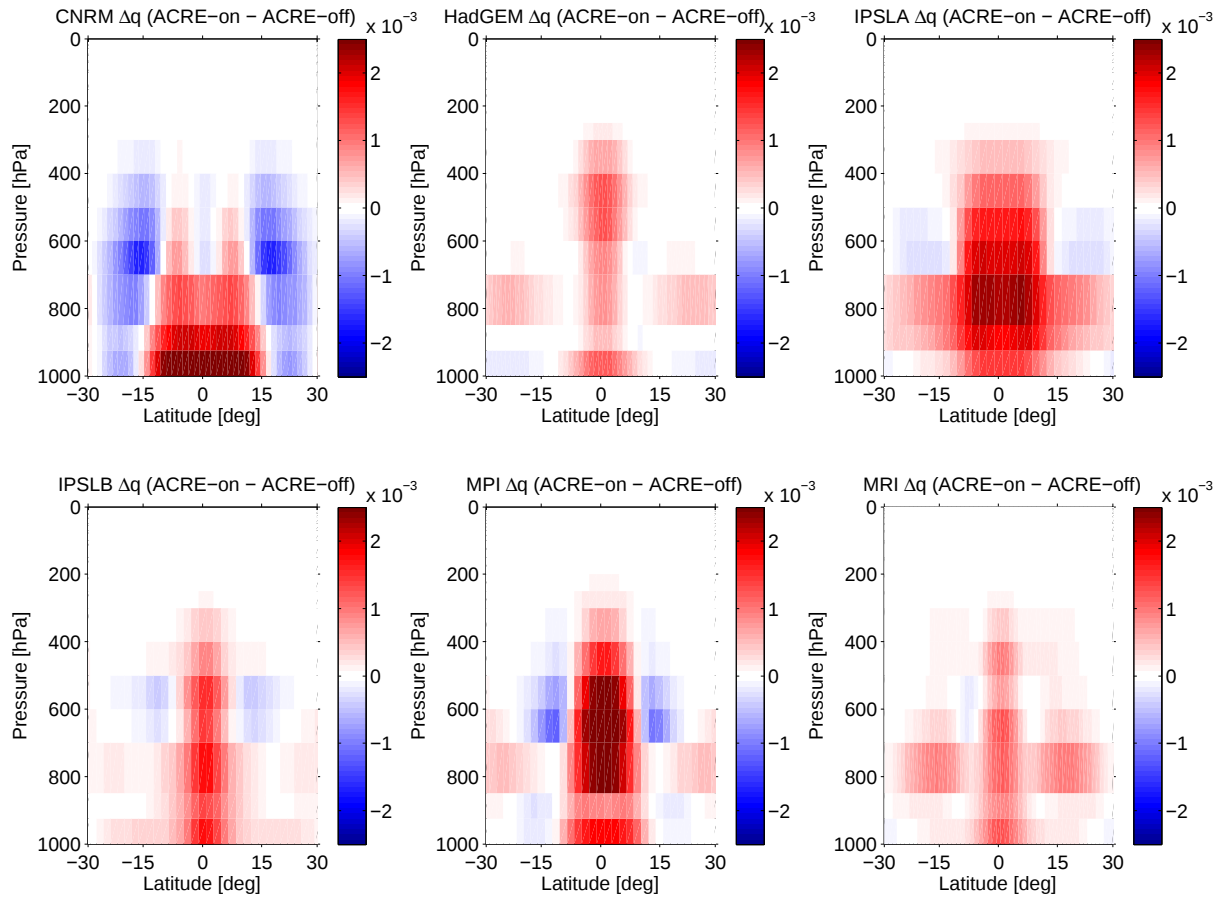


Figure 3.6: Specific humidity increase caused by the atmospheric cloud radiative effect for all six COOKIE models.

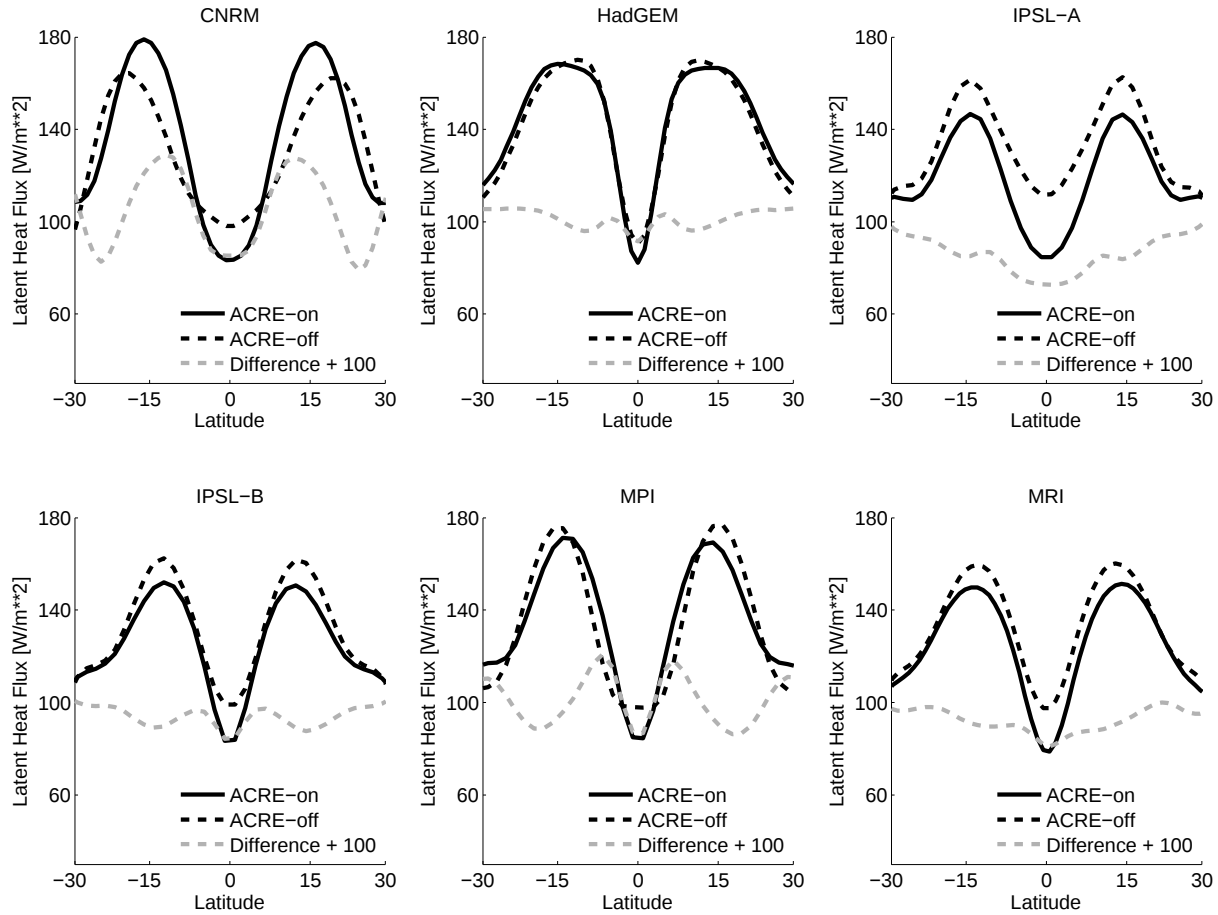


Figure 3.7: Latent heat flux for the six different models. The gray dashed line shows the difference between the ACRE-on and ACRE-off experiments.

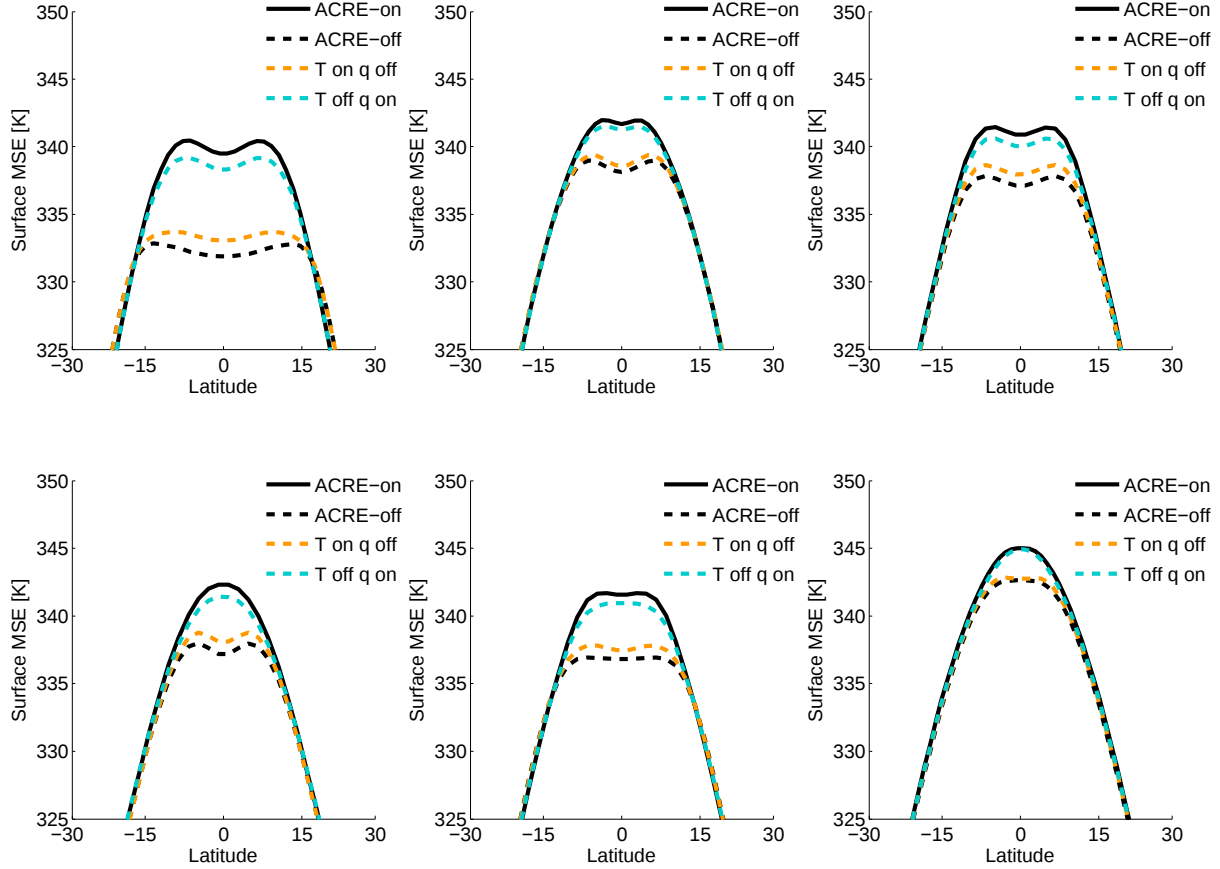


Figure 3.8: Surface Moist Static Energy (MSE) for the six different models. The black solid line is the ACRE-on simulation while the black dashed line is the ACRE-off simulation. The orange dashed line is the surface MSE calculated using the air temperature from the ACRE-on simulation and the humidity from the ACRE-off simulation. The teal dashed line is the surface MSE calculated using the air temperature from the ACRE-off simulation and the humidity from the ACRE-on simulation. The MPI model did not include surface humidity in the model output database, so we interpolated it using the full 3D humidity field and the surface pressure field.

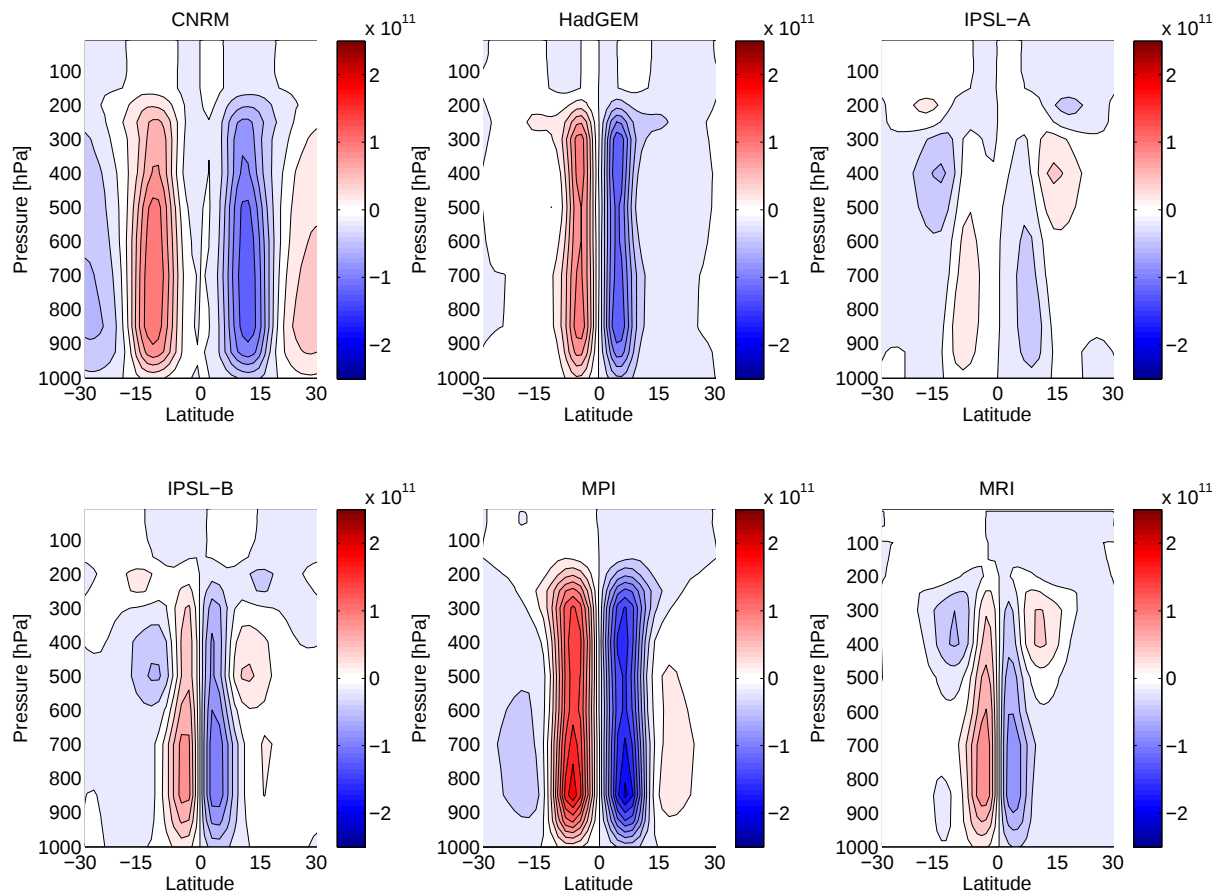


Figure 3.9: Difference in mass stream function (in units of kg/s) between ACRE-on and ACRE-off simulations for all six models. Positive is a counterclockwise circulation anomaly, so that all models show an increase in tropical overturning near the equator in response to cloud radiative forcing.

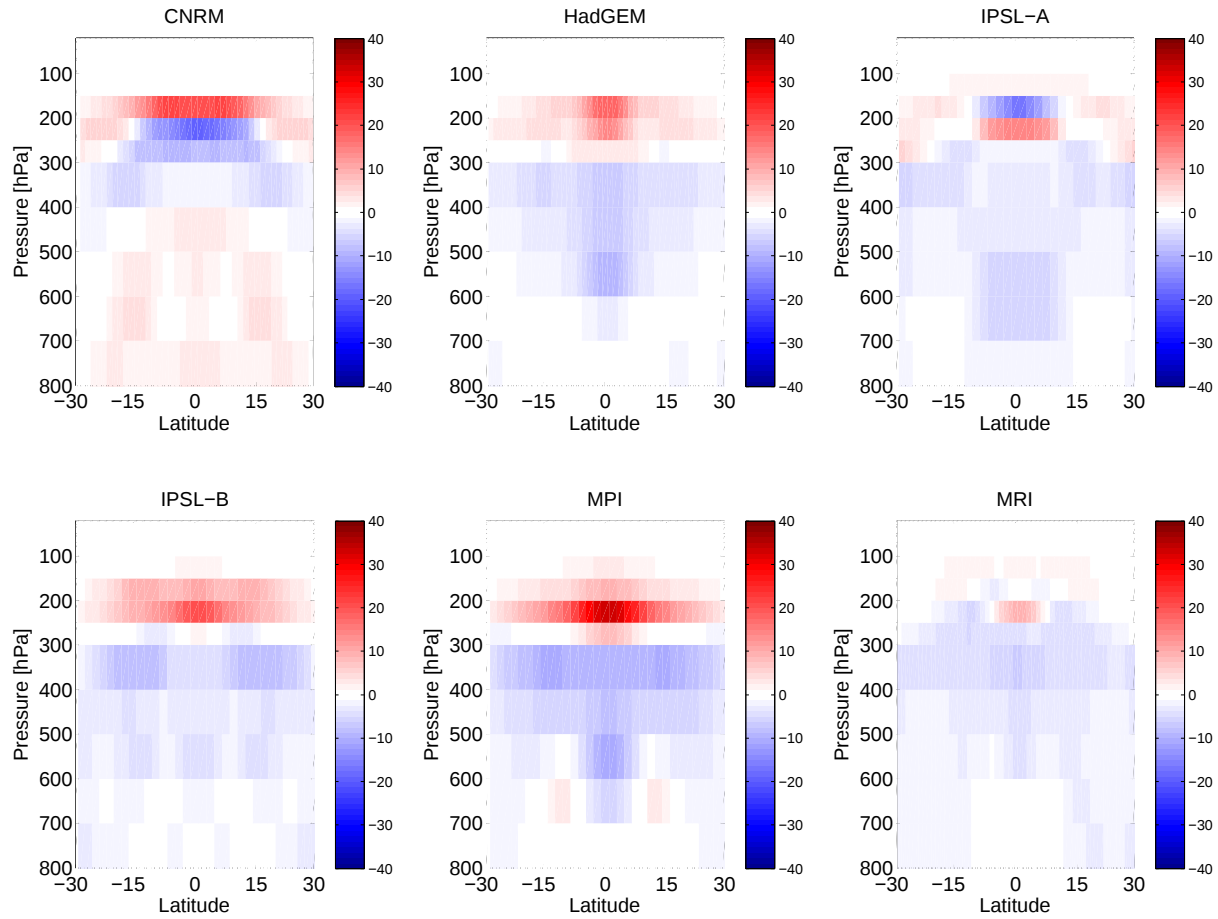


Figure 3.10: Difference in radiatively-driven subsidence above the boundary layer (ACRE-on – ACRE-off). Units are in hPa/day.

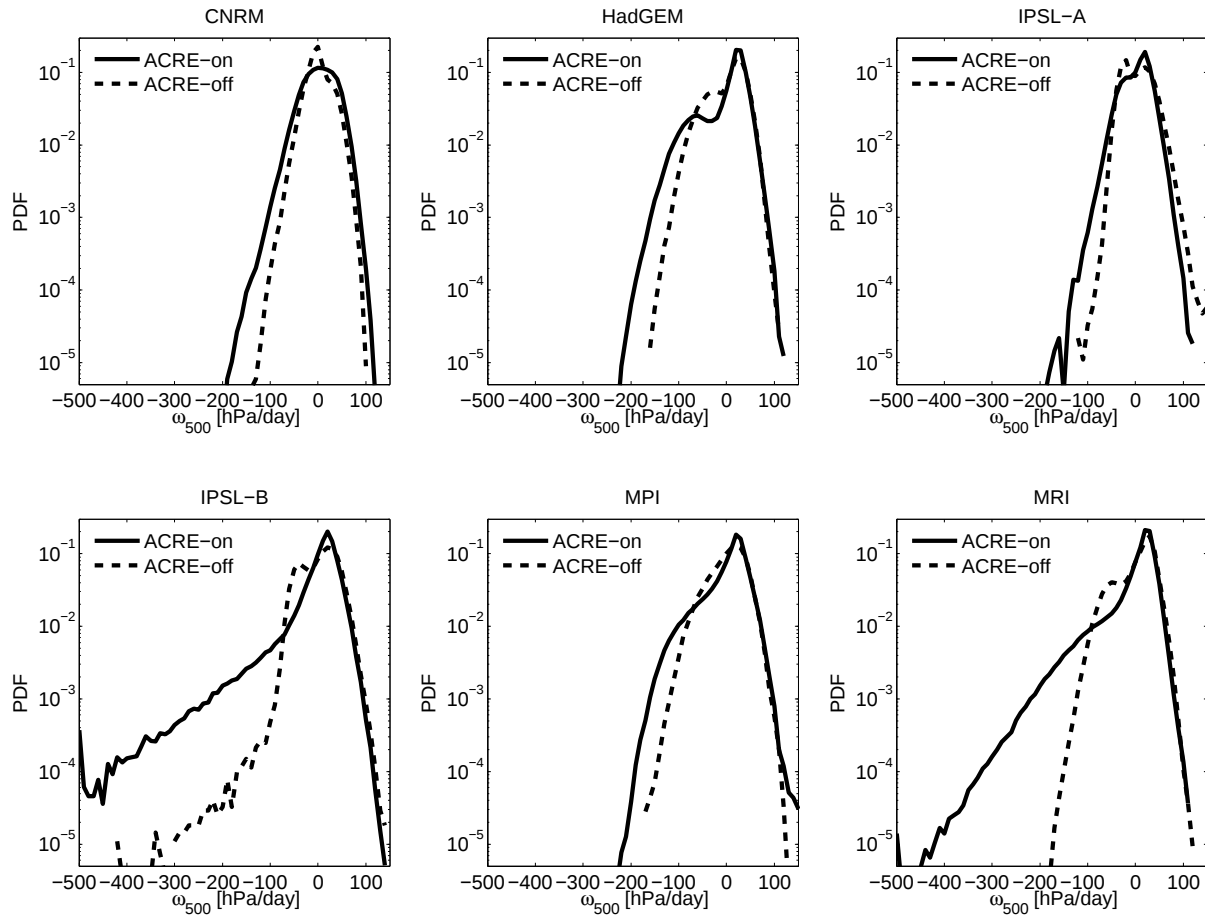


Figure 3.11: PDF of 500 hPa vertical velocity for the six different models (30S – 30N). The solid line is the ACRE-on configuration; the dashed line is the ACRE-off configuration.

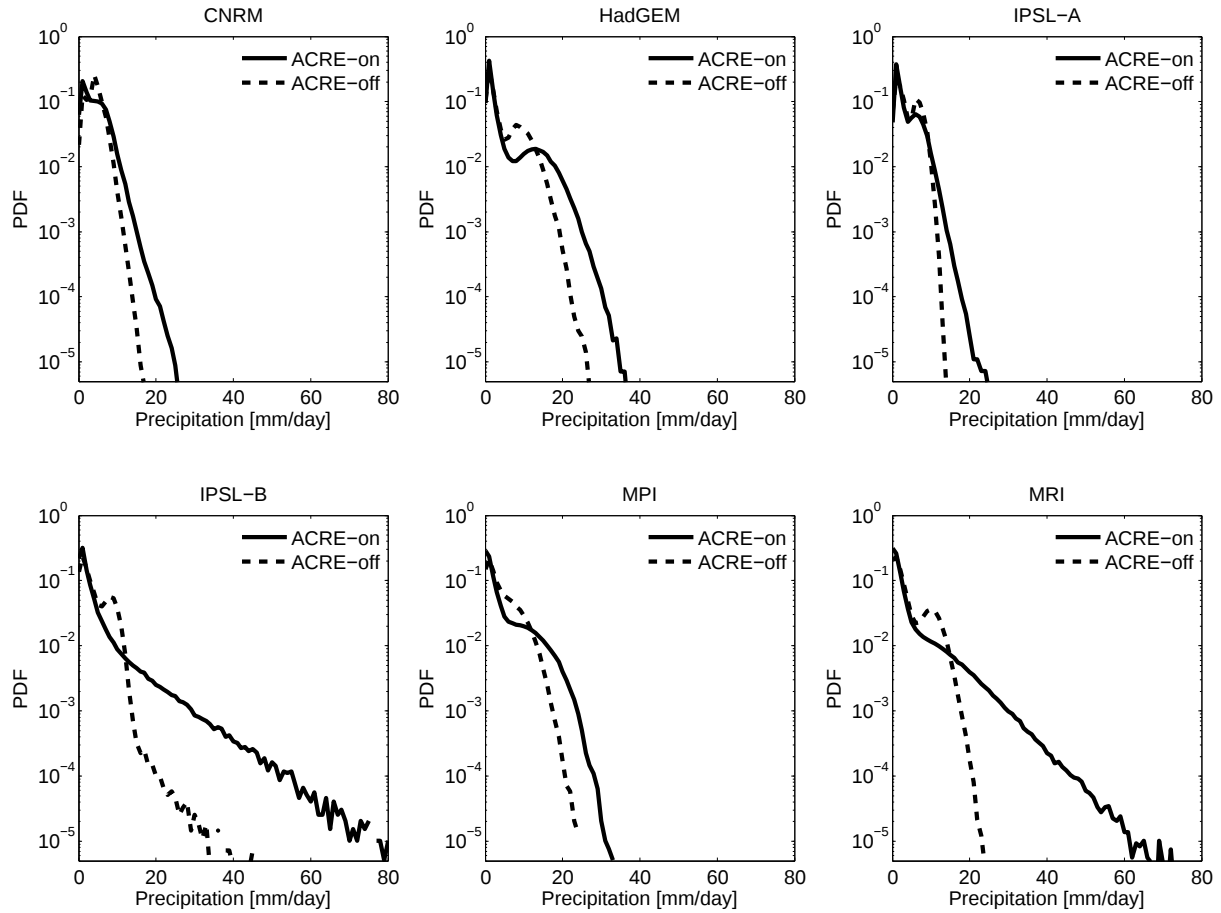


Figure 3.12: PDF of precipitation for the six different models (30S – 30N). The solid line is the ACRE-on configuration; the dashed line is the ACRE-off configuration.

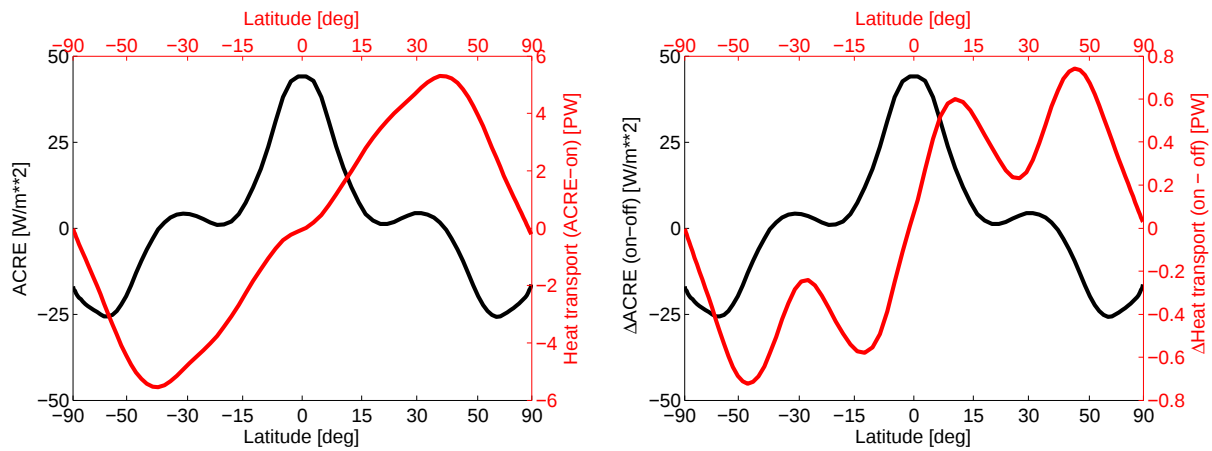


Figure 3.13: (Left) multi-model mean, zonal mean ACRE (in black) and atmospheric poleward energy transport (in red) for the six COOKIE models. (Right) same as left, only for the difference between ACRE-on and ACRE-off experiments. Note that the poleward energy transport is calculated indirectly from the top-of-atmosphere and surface fluxes, both radiant and turbulent.

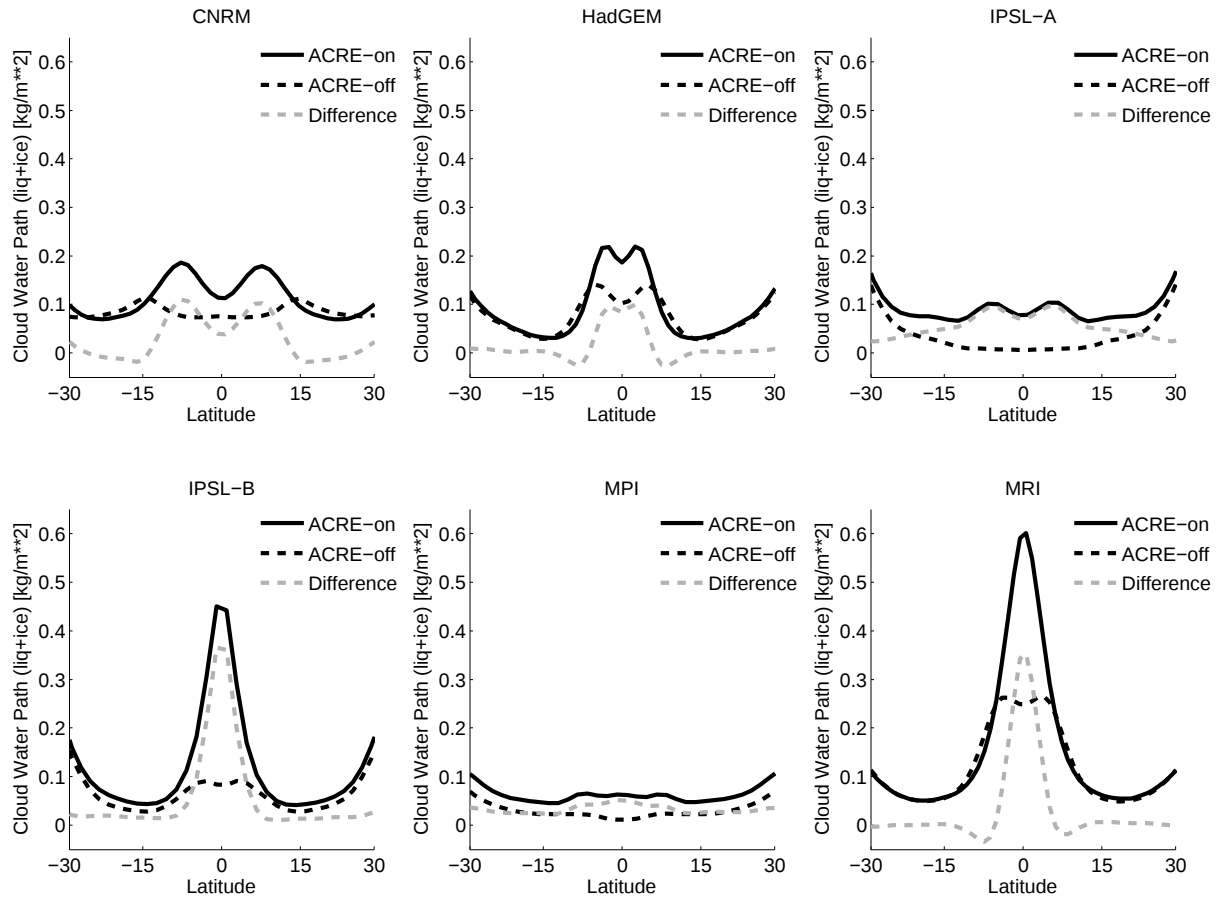


Figure 3.14: Cloud water path (liquid + ice) for the six different models. The gray dashed line shows the difference between the ACRE-on and ACRE-off experiments.

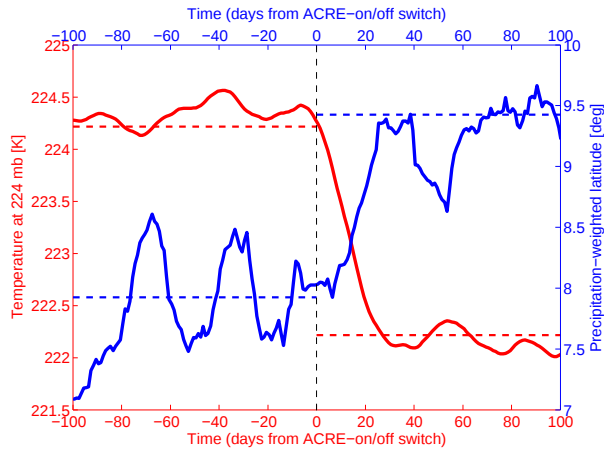


Figure 3.15: Time series of Temperature at 224 hPa averaged from 15 S – 15 N and of precipitation-weighted latitude (ϕ_P). Day 0 is the transition from radiatively-active clouds to radiatively-inactive clouds. The data are smoothed by a boxcar smoother of fifteen days. The horizontal dashed lines represent the mean values of the equilibrated simulation prior to and after the transition. Data are taken from the GFDL AM2.1 transient simulations run for this study.

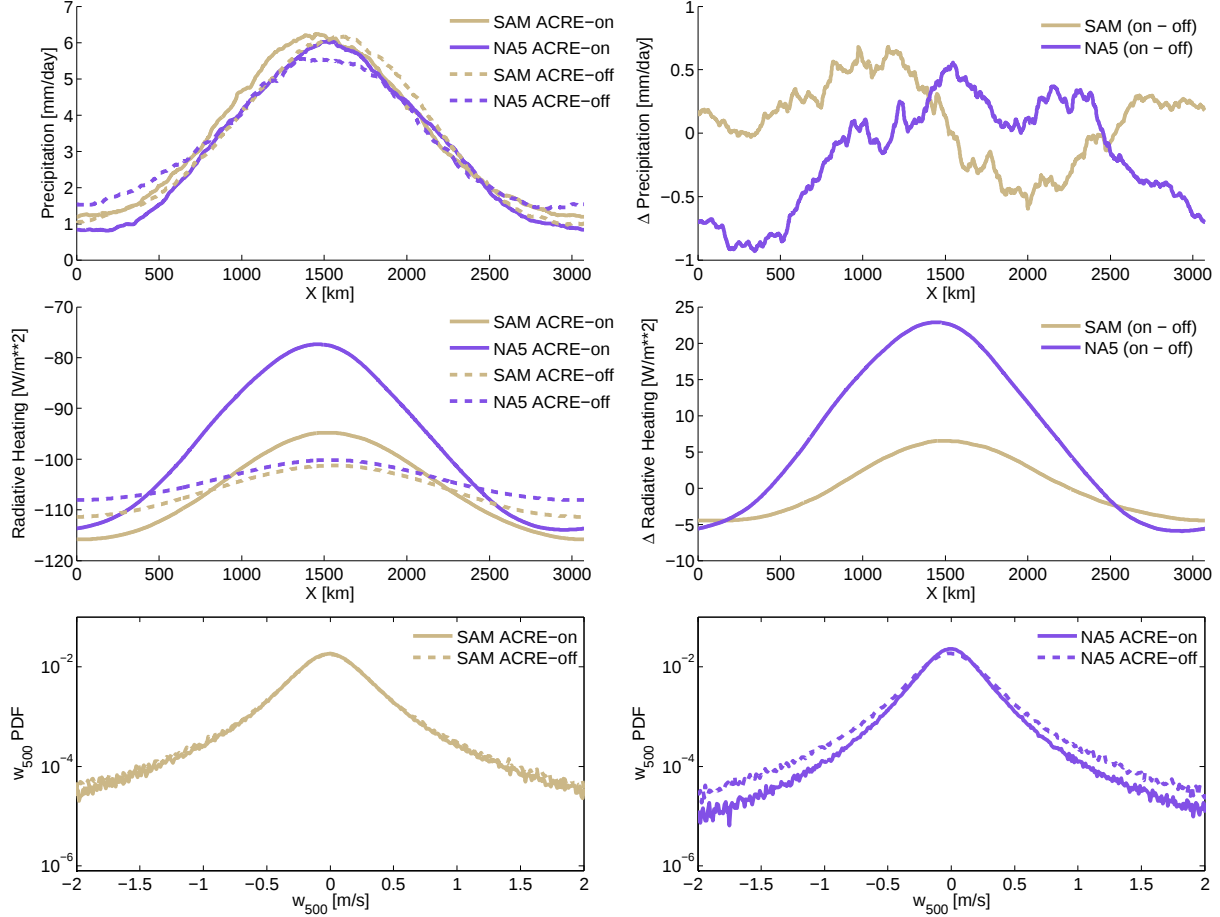


Figure 3.16: Top row: (Left) mean precipitation amounts for both microphysics settings (SAM in gold and NA5 in purple) and for both cloud-radiation configurations (ACRE-on and ACRE-off); (right) difference (ACRE-on – ACRE-off) in precipitation. Middle row: (Left) mean radiative heating rate for both microphysics (SAM and NA5) and both cloud-radiation configurations (ACRE-on and ACRE-off); (right) differences (ACRE-on – ACRE-off) in cloud water path. Bottom row: (Left) PDFs of vertical velocity at 500 hPa for both cloud-radiation configurations with the SAM microphysics; (right) same as (left) only for the NA5 microphysics.

Chapter 4

THE ROLE OF CLOUD HEATING WITHIN THE ATMOSPHERE ON THE HIGH CLOUD AMOUNT AND TOP-OF-ATMOSPHERE CLOUD RADIATIVE EFFECT

4.1 *Introduction*

Clouds have a strong impact on the radiative budget of the atmosphere. They influence the amount of shortwave and longwave radiation absorbed by the atmosphere and surface. The effect of clouds on the energy balance of the atmosphere is quite different from their effect at the surface or the top of the atmosphere. On average, clouds cool the surface by reflecting shortwave radiation, and may heat it by increasing the downward flux of longwave radiation. The effect of clouds on the atmospheric energy budget, however, is less straightforward. Because the tropospheric lapse rate is positive, clouds can either warm or cool the atmosphere, depending on their vertical structure. Low clouds can cool the atmosphere by increasing the downward emission of longwave radiation, while high clouds warm it by decreasing the upward emission of longwave radiation (see, for example, Slingo and Slingo, 1988). Over regions of deep convection within the tropics, cloud-radiation interactions warm the atmosphere (Slingo and Slingo, 1988).

The cloud-radiative heating within the atmosphere (the difference in the all-sky and clear-sky radiative heating profiles) is termed the Atmospheric Cloud Radiative Effect (ACRE). It has been shown that ACRE can be used to drive more realistic tropical circulations (Hartmann et al., 1984; Tian and Ramanathan, 2003). It has also been shown that ACRE feeds back on the clouds through destabilization of the cloud layer by radiation (Ackerman et al., 1988; Chen and Cotton, 1988; Lilly, 1988; Fu et al., 1995; Mace et al., 2006). Ackerman

et al. (1988) showed that the heating profile of anvil clouds can vary by as much as 200 K/day from the bottom to the top of the anvil cloud. Powell et al. (2012) showed that the heating and cooling at cloud bottom and top, respectively, is greatest for the thickest anvils. Fu et al. (1995) showed that a simulated squall line system run with interactive cloud radiation produced greater high cloud cover than the same squall line system run without interactive cloud radiation. They also noted, however, that the cloud radiative heating would begin to stabilize the atmosphere to deep convection if they continued their experiments beyond a few hours. Similar cloud expansion can be simulated for tropical tropopause layer cirrus as shown by Durran et al. (2009) and Dinh et al. (2010).

As already noted, ACRE is positive in tropical convective regions (Slingo and Slingo, 1988), and thus, the upper troposphere is warmed by radiative heating within the clouds (see also, Slingo and Slingo, 1991). Sherwood et al. (1994) found that removing the cloud-radiation interactions in a general circulation model reduced the cloud cover over the tropical warm pool. It is unlikely, however, that their model setup was capable of resolving the within cloud destabilization process, and it is unclear whether the radiative heating gradient between cloud bottom and top would sustain clouds without this mixing. Thus it is difficult to assess the relative importance of both the cloud-layer destabilization and the atmospheric stability change from the Sherwood et al. (1994) experiments.

Lebsock et al. (2010) showed a strong correlation between tropical mean convection and cloud radiative fields on daily timescales. They showed that tropical mean precipitation correlates positively with reflected shortwave and negatively with outgoing longwave radiation. They suggested a “radiative-convective cloud feedback,” wherein precipitation enhances clouds, which decrease the radiative cooling, and hence reduce precipitation. In this paper, we include an additional pathway in which the radiation changes associated with clouds feeds back on the clouds directly.

To summarize, cloud-radiation interactions have been shown to be important for destab-

bilizing anvil clouds in tropical deep convective systems. It has also been suggested that the net heating by these clouds will stabilize the atmosphere and alter deep convection, which will change other components of the tropical system such as the OLR and the precipitation. Much of the prior work centers on case studies of individual cloud systems and their interaction with radiation (e.g. Fu et al., 1995; Durran et al., 2009; Dinh et al., 2010). Other studies have examined the change in cloud amount to the heating of the atmosphere by ACRE (e.g. Sherwood et al., 1994). To our knowledge, however, ours is the first study to examine both of these effects together to quantify the net impact of cloud-radiation interactions on the tropical clouds.

We have designed a series of equilibrium cloud resolving model experiments as a means of exploring this problem, and with which we can examine the role of cloud radiative heating on tropical convection. Equilibrium experiments allow us to collect statistics from numerous cloud life cycles instead of relying on drawing conclusions from a single cloud system. We hope to address several questions in this paper: (1) how does cloud fraction respond to ACRE? (2) How does the top-of-atmosphere cloud radiative effect respond to ACRE? (3) Does a large-scale circulation mitigate the stability effect of the cloud radiative heating? and (4) What are the implications for cloud feedbacks?

In this paper, we confirm that the cloud fraction increases owing to an increase in turbulent kinetic energy driven by cloud-radiation interactions. We also show, however, that the stability changes over the depth of the troposphere dominate the cloud fraction by stabilizing the atmosphere to deep convection and limiting both the cloud fraction and the precipitation. We show that cloud-radiative heating further changes the clouds to be both thinner and warmer, making the top-of-atmosphere cloud radiative effect (CRE) less negative. We also show that the stability effect is mitigated somewhat by having a large-scale circulation within the model that can transport energy to a region where it can be radiated to space more efficiently. Finally, we suggest that the reduction in high cloud amount over the tropics

predicted in current general circulation models may be exaggerated, owing to their inability to resolve within-cloud circulations.

4.2 *Model details and experimental design*

We use the System for Atmospheric Modeling (SAM) cloud resolving model version 6.7 (Khairoutdinov and Randall, 2003). The model uses the anelastic equations of motion, has a rigid lid, and Newtonian damping in the upper third. It has three prognostic variables: liquid water/ice moist static energy, total non-precipitating water (vapor, cloud liquid, and cloud ice), and total precipitating water (rain, snow, and graupel). The vertical grid is stretched with 96 levels. The simulations are run with doubly periodic boundary conditions on the sides and fixed SST. Monin-Obukhov similarity is used to compute the surface turbulent fluxes. The Rapid and Accurate Radiative Transfer Model for GCMs is used for the radiative transfer calculations (see Mlawer et al., 1997; Iacono et al., 2008). The simulations are performed at the equator with no rotation and perpetual insolation. The model is run to radiative convective equilibrium, which takes 50 days in these simulations. We then run an additional 50 days to collect statistics.

We first perform a suite of simulations with a horizontal grid of 96 km x 96 km with 1 km resolution. In these small-domain simulations, we use a uniform fixed sea surface temperature for the lower boundary condition. In the real tropical warm pool, energy is diverged through atmospheric motions that are not present in these small-domain simulations. Thus, we perform a similar set of simulations within a mock-Walker experiment. The mock-Walker simulations are run on a 6144 km x 32 km domain with 4 km resolution. We tested the sensitivity of the cloud areal extent and ACRE profiles to resolution in the small domain and found little change between 1 km, 2 km, 3 km, or 4 km horizontal resolutions, so we do not expect the change from 1 km to 4 km horizontal resolution to be a major difference between the two modeling setups. For the mock-Walker simulations, the sea surface temperature

(SST) is again held fixed, but in these experiments, we introduce a sinusoidal SST pattern with a peak-to-peak amplitude of 4 K.

We run the model both with the cloud-radiation interactions included (T1C1) and with the cloud-radiation interactions removed (T0C0). When the cloud-radiation interactions are removed, the clear-sky radiative heating rates are used to update the energy tendency terms, making the clouds invisible to both shortwave and longwave radiation. To separate the cloud changes associated with the direct radiative heating of clouds from those associated with the stability effect, we run another pair of simulations. In the first, we still use the clear-sky fluxes to update the energy budget, but we additionally add in the domain mean average ACRE from the T1C1 experiment. In this case, the domain mean radiative cooling profile remains the same between this experiment and the T1C1 experiment, but the direct cloud-radiation interactions are removed, so that only the stability effect is included. We refer to this experiment as T1C0. In the second experiment, ACRE is included, but we remove its domain mean profile at each time step. Again, this method has the desired effect of maintaining a radiative cooling profile that matches that of the T0C0 experiment, but still allows for the radiative heating within the clouds to occur. We refer to this experiment as T0C1. These four experiments allow us to examine the direct effect of ACRE on the clouds by differencing the T1C1 and T1C0 as well as the T0C1 and T0C0 experiments. Similarly, we can investigate the effect of ACRE stabilizing the atmosphere and thus suppressing deep convection by differencing the T1C1 and T0C1 experiments or the T1C0 and T0C0 experiments.

In the limited area domain simulations, the T1C0 and T0C1 experiments have the advantage that they preserve the domain mean integrated radiative cooling and surface precipitation to their T1C1 and T0C0 counterparts, respectively. In the mock-Walker circulation experiments, however, the precipitation is not conserved because adding/removing the domain mean-ACRE uniformly across the domain does not preserve the structure in the time-average radiative cooling distribution, so the large-scale circulation will adjust.

The single moment microphysics scheme in SAM does not produce a lot of high, thin clouds, which makes the ACRE signal small in this modeling framework. To amplify the signal, we make use of a modification to the microphysics scheme based on the work of Lopez et al. (2009), which produces more realistic anvil cloud, tuned to reproduce tropical high cloud optical depth statistics observed by MODIS. We employ their “NOSEDAALIQ5” specifications, which remove cloud ice sedimentation, lower the threshold for ice autoconversion by a factor of 100, and increase the autoconversion and accretion of liquid by factors of five. For convenience, we will refer to this perturbed scheme as the “NA5” microphysics. The NA5 microphysics produces more extensive anvil and cirrus clouds within the model, enhancing the ACRE profile within our simulations. As a result, the NA5 microphysics does a much better job of reproducing a realistic ACRE heating profile (see Fig. 4.1) compared to cloud heating rates calculated using CloudSat/CALIPSO/MODIS retrievals over the tropical warm pool (heating rate data provided by Qiang Fu). For the upper troposphere in the NA5 version of the model, the longwave heating is a little too strong compared to observations, while the shortwave heating is a little too weak, such that the net is nearly identical to observations. The model does not generate sufficient mid-level clouds, so it is missing the additional peak near 6 km. A horizontal resolution of 1 km is insufficient for resolving low clouds accurately (Bretherton et al., 2006), so we do not expect to accurately reproduce the heating rates in the lowest part of the atmosphere.

We run the NA5 microphysics for the T1C1, T0C0, T1C0 and T0C1 configurations at sea surface temperatures of both 28.5°C and 32.5°C for the uniform SST experiments. We also run two different SST profiles for the mock-Walker experiments. In the first, the peak temperature in the warm pool is 28.5°C, and in the second, the peak temperature is 32.5°C. Figure 4.2 shows the ACRE profiles for the T1C1 experiments for each sea surface temperature and domain configuration. The calculation of ACRE for the mock-Walker circulation experiments is limited to the warm pool (defined as the half of the domain where

the sea surface temperatures exceed the average). Clouds rise with warming, consistent with the Fixed Anvil Temperature (FAT) hypothesis (e.g. Hartmann and Larson, 2002; Kuang and Hartmann, 2007; Zelinka and Hartmann, 2010; Harrop and Hartmann, 2012). We also note that the ACRE profile for the warm pool in the mock-Walker circulation experiments is comparable to that of the limited domain experiments, with two notable exceptions. The mock-Walker circulation experiments produce more mid-level clouds and more low clouds. Our study, however, will focus primarily on the high clouds.

4.3 Results

4.3.1 *How does cloud fraction respond to ACRE?*

The first question we wish to address is a direct follow-up to Fu et al. (1995). Does ACRE increase the high cloud fraction? We will begin our analysis with the limited area domain simulations. Here we adopt the convention of high clouds as those clouds with a cloud top pressure less than 440 hPa. The cloud-radiative heating decreases the high cloud fraction - from 39% to 27% for SST = 28.5°C and from 36% to 25% for SST = 32.5°C. The high cloud reduction can be seen as the difference between T1C1 and T0C0 for the two different SSTs in Fig. 4.3 (top row). We expect that the reduction of cloud fraction by ACRE is the combination of the two effects of cloud radiative heating. The first is the radiative destabilization of the anvil and cirrus clouds, by heating at the bottom and cooling at the top (which we will refer to as the direct cloud heating effect). The second is the stabilizing effect on the temperature profile from the cloud heating (which we will refer to as the indirect cloud heating effect). To separately assess the direct and indirect cloud heating effects we rely on our additional experiments.

It is worthwhile to take a moment and be clear what the different experiments mean and the language we will use to describe the physical understanding gained by using them. The direct cloud heating effect mentioned above (the destabilizing of anvil clouds by heating at

the bottom and cooling at the top) can be determined by differencing experiments T1C1 and T1C0 for a given domain size and sea surface temperature. Here, the “C1” means the clouds are radiatively active and “C0” means the clouds are radiatively inactive. “T1” means the temperature profile is the one produced by radiatively-active clouds that are allowed to heat their environment. Thus, T1C1 and T1C0 share similar environmental temperature profiles, similar stability profiles (see Fig. 4.4), the same domain mean radiative cooling profile, and the same surface precipitation. Their difference only comes from the spatial heterogeneity in the radiative heating by the clouds, and therefore their difference depends only on the direct interactions between clouds and radiative heating. Differencing T0C1 and T0C0 produce the same direct cloud heating effect information, only for an environment whose temperature profile is determined by radiatively-inactive clouds. In both cases, we are determining the influence of the direct cloud heating effect by differencing “C1” experiments with their corresponding “C0” experiments.

The indirect cloud heating effect (the stabilizing of the troposphere to convection through heating of the upper troposphere by clouds) can be determined by differencing experiments T1C0 and T0C0 for a given domain size and sea surface temperature. T1, T0, C1, and C0 have the same meaning as above. In both T1C0 and T0C0 experiments, the clouds are radiatively inactive, but the mean cloud heating still heats the environment and enhances the atmospheric stability in the T1C0 experiment. As for the direct cloud heating effect calculation in the paragraph above, the indirect cloud heating effect can also be gotten by differencing the T1C1 and T0C1 experiments. Thus, we are determining the influence of the indirect cloud heating effect by differencing “T1” experiments with their corresponding “T0” experiments.

For convenience, we will refer to the combination of the direct cloud heating effect and indirect cloud heating effect as the total cloud heating effect. The total cloud heating effect is determined by the difference in the T1C1 and T0C0 experiments for a given domain size

and sea surface temperature.

Returning to Fig. 4.3, we can see the role of the direct and indirect cloud heating effects on the cloud fraction profile. The direct cloud heating effect ($T1C1-T1C0$ or $T0C1-T0C0$) always leads to an increase in the cloud fraction. The increase in cloud fraction occurs primarily between 200–400 hPa, roughly the outflow layer for anvil clouds, as expected. The total high cloud fraction is significantly larger as a result of the direct cloud heating effect (significant at the 95% level). The increase in high cloud fraction is 11–13% for the high stability profile (T1) and 27–29% for the low stability profile (T0), both for the small-domain experiments (for the mock-Walker circulation experiments, the increase is 4–6%). Compared to the direct cloud heating effect, the indirect cloud heating effect has the opposite effect on the cloud fraction. The indirect cloud heating effect ($T1C0-T0C0$ or $T1C1-T0C1$) decreases the cloud fraction. The high cloud fraction decreases by 38–41% for the interactive cloud experiments (C1) and by 22–25% for the non-interactive cloud experiments (C0), both for the small-domain experiments (for the mock-Walker circulation experiments, the decrease is 3–6%).

From the above, the total cloud heating effect ($T1C1-T0C0$) is dominated by the indirect cloud heating effect, such that the high cloud fraction is reduced by 11–13% in the limited domain simulations. In the mock-Walker circulation experiments, the total cloud heating effect on cloud fraction is small (1% or less) because the direct and indirect cloud heating effects balance one another. It should be noted that the vertically resolved cloud profiles are not the same between the T1C1 and T0C0 experiments, so while the areal coverage is insensitive to the total cloud heating effect, the cloud properties may not be. We will return to this point later in the paper.

We can ask whether the increase in cloud fraction owing to the direct cloud heating effect is in fact due to increased destabilization within the cloud layer. To investigate this feature, we first look at the turbulent kinetic energy (TKE) profiles for the same experiments

presented in Fig. 4.3. Figure 4.5 shows that the TKE in the upper troposphere shows a strong peak for the experiments with the direct cloud heating effect included (T1C1 and T0C1). Note that convection is stronger in both T0C1 and T0C0 experiments, as can be seen by the stronger TKE throughout the depth of the troposphere. Additionally, the presence of more clouds, as in the weaker stability cases, generates more TKE as well. Still, the difference in TKE between the cloud layer and the mid-troposphere is largest when the direct cloud radiative heating effect is included.

We are interested in asking how much of the turbulent motions are actually driven by the anvil and cirrus clouds. Is there any contribution from the deep convective cores? Convective cores are defined here as cloudy layers where the vertical velocity exceeds 1 m/s (either upward or downward) and the virtual potential temperature anomaly is greater than zero for updrafts (less than zero for downdrafts). Thus, regions of non-convective core cloud are those that are cloudy, but are not part of the strong updraft or downdraft regions of the convection. Figure 4.6 shows the transport of virtual potential temperature in regions of non-convective core cloud. Virtual potential temperature serves as a useful proxy for the buoyancy. Subtracting out the cores allows us to look at the effects of buoyancy transport in the anvil and cirrus clouds separated from the deep convection. Figure 4.6 shows that direct cloud heating effect more than doubles the θ_v transport within the clouds. The direct cloud heating effect does alter the θ_v profile some owing to the enhanced mixing generated, but those changes in θ_v owing to the direct cloud heating effect are less than 1 K. Therefore, the increase in transport must be driven by increases in vertical velocity within the clouds, driven by the direct cloud heating effect. Indeed, the mass flux profiles are similar to the energy transport profiles (not shown). Furthermore, the increase in the buoyancy transport occurs outside of the deep convective cores, and is instead associated with the anvil clouds and extensive cirrus within the model domain. The direct cloud heating effect is larger when the stability effect is not present. It can be seen in Fig. 4.5 and Fig. 4.6 that in the more

stable T1C1 experiment where convection is less frequent or weaker, less cloud is generated compared to the T0C1 experiment, and hence less cloud is present to interact with the radiation and generate turbulent kinetic energy or θ_v transport.

In the mock-Walker circulation experiments, we can see that the lower and middle troposphere show different responses to the direct and indirect cloud radiative heating effects compared to the limited-domain experiments. The mock-Walker circulation experiments have a lot more low cloud, primarily over the cold pool regions of the domain, and radiation is important for their maintenance as well.

4.3.2 How does the top-of-atmosphere cloud radiative effect respond to ACRE?

An increase in high cloud fraction can have a profound effect on the climate system. Harrop and Hartmann (2015) showed that for an increase in high cloud amount of 1%, top-of-atmosphere net CRE decreases by 0.26 W/m^2 (based on satellite retrievals over the West Pacific warm pool). In their study, changes in high cloud amount were determined as changes in the first empirical orthogonal function (EOF), which described an amplification of the mean cloud distribution in cloud optical depth/cloud top pressure space. Combining this value for CRE as a function of high cloud amount with the changes in high cloud fraction above, we calculate the CRE change one might expect to incur from these high cloud changes, all else being equal. The increase of $\sim 12\%$ or $\sim 28\%$ in high cloud fraction owing to the direct cloud heating effect would cause net CRE to decrease by 3.1 or 7.2 W/m^2 . Likewise, the decrease of $\sim 39\%$ or $\sim 24\%$ in high cloud fraction owing to the indirect cloud heating effect would cause net CRE to increase by $10.$ or 6.1 W/m^2 . These numbers are for the limited-domain experiments.

We do not expect, of course, that the change in CRE owing to the cloud fraction spreading by ACRE to simply reflect a change in average high cloud fraction. Therefore, the next question we will address is how the top of atmosphere cloud radiative effect (CRE) responds

to the interactions between clouds and radiation. Not only will we examine the same experiments as the previous subsection, but we also wish to identify which components of the clouds or environment produce the changes in CRE we see between different experiments. In other words, we seek to quantify how much of the change in CRE comes from changes in cloud fraction compared to how much comes from changes in the cloud radiative properties. To accomplish this, we first write the CRE equation following Hartmann et al. (2001b) as:

$$\text{CRE} = \sum_i A_i [S_0 (\alpha_{\text{clr}} - \alpha_i) + F_{\text{clr}} - F_i], \quad (4.1)$$

where A_i is the cloud frequency, S_0 is the top of atmosphere insolation, α_{clr} is the clear-sky albedo, α_i is the cloudy albedo, F_{clr} is the clear-sky OLR, and F_i is the cloudy OLR. The subscript “ i ” refers to the cloud type. We break the cloud into different types based on cloud top pressure and cloud optical depth using the standard histogram bins for the International Satellite Cloud Climatology Project (ISCCP; Schiffer and Rossow, 1983; Rossow and Schiffer, 1991). The equation above gives the CRE for each cloud type, and its sum gives the total CRE. To get the change in CRE between experiments, we can use the above equation to express the total difference in CRE as a sum of the differences in the various terms, plus a covariance term.

$$\begin{aligned} \Delta \text{CRE} = & \sum_i \Delta A_i [S_0 (\alpha_{\text{clr}} - \alpha_i) + F_{\text{clr}} - F_i] + \sum_i A_i [S_0 (\Delta \alpha_{\text{clr}})] + \sum_i A_i [S_0 (-\Delta \alpha_i)] + \\ & \sum_i A_i [\Delta F_{\text{clr}}] + \sum_i A_i [-\Delta F_i] + \sum_i \Delta A_i [S_0 (\Delta \alpha_{\text{clr}} - \Delta \alpha_i) + \Delta F_{\text{clr}} - \Delta F_i] \end{aligned} \quad (4.2)$$

Note that in equation (4.2), the change in clear-sky albedo is typically negligible, but

we leave it in for completeness. Changes in clear-sky OLR are especially important when comparing across different sea surface temperatures. In order to separate the model clouds into the ISCCP histogram bins, we first calculate the cloud optical depth as follows:

$$\tau = \frac{3}{2} \sum_j \left(\frac{\text{LWP}_j}{r_{\text{liq}}} + \frac{\text{IWP}_j}{r_{\text{ice}}(T_j)} \right) \quad (4.3)$$

where τ is the cloud optical depth, LWP is the liquid water path, IWP is the ice water path, r_{liq} is the cloud liquid effective radius, r_{ice} is the cloud ice effective radius, T is temperature, and the j index denotes an individual model layer. The SAM model assumes a fixed cloud liquid effective radius of $14 \mu\text{m}$. The ice effective radius follows the CAM3 parameterization based on Kristjánsson et al. (2000) and is a function of temperature alone, with colder temperatures producing smaller effective radii, and hence, larger cloud optical depths for a constant IWP.

The breakdown of CRE into its components allows us to describe the influence of the direct, indirect, and total cloud heating effect on the CRE as changes in cloud amount, cloud albedo, cloud OLR, clear-sky terms, and a covariance term. The cloud amount term is straightforward. A shift in the abundance of various cloud types will change the CRE based on the CRE values for the cloud types that become more frequent and those that become less frequent. The cloud albedo factor and cloud OLR factor may be interpreted as a change in the mean cloud albedo or cloud OLR in each histogram bin. For example, if the atmosphere warms at a given pressure level, the cloud OLR would change without a shift in the amount histogram. Likewise, changes in albedo can occur within a given histogram bin that are not accounted for by changes in the frequency of that bin. The primary benefit of breaking the CRE changes down into the ISCCP histogram is that it allows us to isolate the changes in CRE owing to the high clouds alone, simply by summing over the histogram bins where $\text{CTP} < 440 \text{ hPa}$. For simplicity, the following discussion will only focus on CRE changes owing to high clouds ($\text{CTP} < 440 \text{ hPa}$). It is worth noting that the CTP and cloud

optical depth are calculated as column amounts looking down so that they are comparable to satellite retrievals. Table 4.1 shows the change in high cloud CRE breakdowns when comparing different experiments.

First, Table 4.1 shows that the direct cloud heating effect increases the CRE from the high clouds, making it less negative. The increase in CRE is greater when stability is weaker and high clouds are more abundant than for the cases where the stability is higher and high clouds are less abundant. For the limited domain experiments, the cloud amount, cloud albedo, and cloud OLR are the dominant factors in determining the total CRE increase. The cloud albedo and cloud OLR changes are largely offsetting, however, such that the total CRE increase is quite similar to that predicted solely by the change in cloud amount. Figure 4.7 shows the breakdown of the CRE change from the direct cloud heating effect. We can see that some cloud types contribute to the increase in CRE, while other cloud types decrease the CRE. Most of the increase in CRE comes from the cirrus ($CTP < 440$ hPa and $\tau < 3.6$). Deep convection ($CTP < 440$ hPa and $\tau > 23$) and cirrostratus ($CTP < 440$ hPa and $3.6 < \tau < 23$) have counterbalancing effects on CRE. The changes in the highest clouds ($CTP < 180$ hPa), increase CRE, while slightly lower clouds ($180 \text{ hPa} < CTP < 440$ hPa) decrease CRE. Figure 4.8 shows the change in cloud amount for each cloud type. As for CRE, most of the increase occurs for the cirrus clouds. The anvil clouds also increase, giving rise to the counterbalancing negative CREs seen in these bins. There is very little change in the cloud amount in the highest cloud top pressure bins.

Figure 4.7 looks similar to an amplification of the basic CRE ISCCP pattern (compare with Fig. 4.9) with the notable exception of clouds whose cloud top pressure is less than 180 hPa while having a cloud optical depth exceeding 3.6. For clouds whose $CTP > 180$ hPa, there is an increase in CRE for cloud types that have a positive CRE and a decrease for those cloud types that have a negative CRE. In other words, the CRE for these clouds is amplified. Ackerman et al. (1988) suggested that thin anvils would be more susceptible to the effects

of radiative destabilization than thicker anvils. Figure 4.7 is in agreement since the largest change in CRE occurs for the thinner clouds. Figure 4.9 shows the contribution to the CRE from each histogram bin for each of the limited domain experiments. As expected from observations (see Hartmann et al., 2001b), there is a balance between the positive CRE from the high, thin clouds and the negative CRE from the high, thick clouds. Figure 4.10 shows the ISCCP histogram for each of the limited-domain experiments. Compared to observations (see Harrop and Hartmann, 2015, figure 7), cirrus are favored over anvil clouds in our model. Our modeled clouds also tend to fall in the 180-310 hPa cloud top pressure bin. Without a Brewer-Dobson circulation, however, our tropopause is likely to be too low in the model, which agrees with the ACRE profile comparison between model and observations in Fig. 4.1).

We can perform the same breakdown of which cloud types contribute to the change in CRE when driven by the indirect cloud heating effect as well. Figure 4.11 shows the same plots as Fig. 4.7, but for these stability differences owing to the indirect cloud heating effect. The increase in stability tends to reduce cloud amounts for all high clouds (Fig. 4.12) and this is reflected in the change in CRE ISCCP histograms in Fig. 4.11. The high, thin clouds that have a positive CRE are reduced, causing a decrease in the total CRE, while the high, thick clouds that have a negative CRE are also reduced, causing an increase in the total CRE. When the direct cloud heating effect is included, more high, thin clouds are lost when stability increases, such that the change in CRE is negative. When the direct cloud heating effect is removed, the opposite occurs. Fewer high, thin clouds are present, so the reduction of clouds is weighted toward the thick clouds and thus, the change in CRE is positive. The covariance terms for the indirect cloud heating effect are the same order of magnitude as the other terms. Therefore, the CRE changes cannot be as simply interpreted for the indirect cloud heating effect as they were for the direct cloud heating effect.

The changes in CRE for increasing sea surface temperature are dominated by the clouds

rising. Figure 4.13 shows that the clouds appear to rise in such a fashion that the cloud amount does not change much. If we refer back to Table 4.1, we see that the cloud amount changes indeed account for very little of the total CRE change for sea surface temperature. Instead, the CRE is dominated by changes in cloud albedo and the clear- and cloudy-sky OLR. The albedo and clear-sky OLR increase the CRE, while the cloudy-sky OLR decreases CRE. The three terms largely cancel one another, such that the total change in CRE with warming is small. We expect the clear-sky OLR to increase CRE since clear-sky OLR will increase in a warmer atmosphere. The albedo response implies the clouds are thinning in a warmer atmosphere, while the cloudy-OLR response implies the clouds are warming some. It has been shown that high clouds do not rise exactly isothermally with increasing SSTs in this CRM owing to the fixed ozone profile impacting stability (see Harrop and Hartmann, 2012). The ozone stability, limits the rise of the clouds, causing them to occur at slightly warmer temperatures than what is predicted by the Fixed Anvil Temperature (FAT) hypothesis (e.g. Hartmann and Larson, 2002; Kuang and Hartmann, 2007; Zelinka and Hartmann, 2010; Harrop and Hartmann, 2012). Because the change in SST is 4 K for each of the differences shown in Fig. 4.13, the feedback would be small for all of the Δ SST experiments (-0.01 W/m^2 , $+0.09 \text{ W/m}^2$, $+0.2 \text{ W/m}^2$, $+0.4 \text{ W/m}^2$).

We have focused on the high cloud changes in our discussion above. As noted in section 4.2, the model resolution is insufficient to accurately simulate low clouds. As such, we do not have confidence in the low cloud changes, or their impact on the top-of-atmosphere CRE. Despite this lack of confidence, our experiments show a consistent reduction in low cloud owing to the direct cloud heating effect and an increase in low cloud owing to the indirect cloud heating effect. There are very few low clouds in the limited domain experiments, which is a result of the microphysics modifications we use. Because all low clouds have a negative CRE over tropical oceans, the reduction (increase) in low cloud decreases (increases) the TOA CRE. The amounts of these changes can be found on the right side of the bottom two

rows in figures 4.7, 4.8, 4.11, 4.12, and 4.13.

Finally, the total cloud heating effect shows an increase in CRE. Again, we see that the clouds thin and warm such that cloud albedo changes invoke an increase in CRE while cloud OLR changes invoke a decrease in CRE. The decrease in total cloud amount leads to an increase in CRE.

Most of our discussion has focused on the limited area domain simulations since they are the more straightforward. It should be noted that, again, like the high cloud fraction changes, many of the changes in CRE owing to the direct, indirect, and total cloud heating effects show similar behavior in the mock-Walker circulation experiments as those in the limited domain experiments. We will elaborate on the differences in sub-section 4.3.3.

4.3.3 Does a large-scale circulation mitigate the stability effect of the cloud radiative heating?

We have focused much of our previous analysis on the limited area domain experiments because they are simpler by construction and easier to understand. The inclusion of a large-scale circulation pattern complicates some of the analysis because there is an additional feedback within the system associated with spatial gradients in ACRE and their effect on circulation strength. Despite the added complexity, many of the responses of the clouds and the top-of-atmosphere cloud radiative effect (CRE) are the same as those seen in the limited area domain experiments. The biggest difference is that of the indirect cloud heating effect.

In a system with both a warm pool and cold pool, the large-scale circulation is capable of transporting some of the energy of ACRE heating from the warm pool to the cold pool. The cold pool can be thought of as a more effective sink of energy, or a “radiator” as described by Pierrehumbert (1995). Thus, the mock-Walker circulation atmosphere does not collect the heat from the clouds and grow quite as stable as it does in the limited area domain experiments. The changes in cloud fraction owing to either the direct or indirect cloud

heating effect are muted compared to the limited area domains, as noted earlier, and the overall cloud fraction in the domain is less.

The changes in CRE in the mock-Walker circulation are less straightforward than the changes in cloud fraction. While the total changes in direct cloud heating effect are more or less the same as the limited area domain experiments, in the mock-Walker circulation experiments, the clear-sky OLR also becomes an important factor. Because we have added in the domain mean ACRE uniformly instead of preserving the time-averaged spatial structure, changes in the circulation occur, and these changes alter the temperature and humidity profiles, and hence the clear-sky OLR.

Despite the changes in cloud fraction from the indirect cloud heating effect being small, the change in CRE owing to albedo, clear-sky OLR, and cloudy-sky OLR changes are still non-trivial. As for the CRE response to direct cloud radiative heating, the clear-sky OLR change results from the circulation response modifying the domain mean thermodynamic profiles. The changes in cloudy albedo and OLR in the mock-Walker circulation experiments behave similarly to those in the limited area domain experiments.

The change in cloud fraction owing to the total cloud heating effect is small, and therefore the increase in CRE owing to the cloud fraction factor is likewise small. The albedo, clear- and cloudy-sky OLR changes are non-negligible, but are largely offsetting, such that the net CRE change in response to the total cloud heating effect is small. Interestingly, the changes in CRE owing to sea surface temperature increase in the mock-Walker circulation experiments are not distinct (both in total and component breakdown) from the limited domain experiments, suggesting that these effects may be robust to the presence of the circulation. More explicit testing in a global-scale circulation is needed to investigate this further.

4.3.4 *What are the implications for cloud feedbacks?*

From the above we have seen that the direct cloud radiative heating effect increases cloud fraction, while the indirect cloud heating effect decreases cloud fraction. The latter of these two is effectively a negative feedback on high cloud fraction. For a decrease in cloud cover, there will be an additional decrease in stability owing to the indirect cloud heating effect. The decrease in stability will favor more convection, and more clouds, thus reducing the initial perturbation. In a warmer climate, the tropical circulation is expected to slow (Vecchi and Soden, 2007). A decrease in cloud amount would be expected to accompany the decrease in mass flux, with a negative feedback on the cloud amount from the indirect cloud heating effect response.

As the climate warms, clouds rise by the Fixed Anvil Temperature (FAT) hypothesis (e.g. Hartmann and Larson, 2002; Kuang and Hartmann, 2007; Zelinka and Hartmann, 2010; Harrop and Hartmann, 2012). All else equal, a rise in the clouds would increase ACRE since the warmer lower troposphere emits more longwave radiation, while the fixed cloud top temperature would mean the clouds emit the same amount of longwave radiation as they did in the cooler climate. Therefore, the heating at anvil cloud base would be expected to increase, while the cooling at the cloud top would remain fixed. This would enhance the destabilization within the cloud layer, as well as increase ACRE within the column, and thus, increase both the direct and indirect cloud heating effects. As we have seen from the above, these two effects counteract one another. As a result, the change in cloud fraction and CRE owing to the increase in ACRE from rising clouds can either increase or decrease depending on the relative magnitudes of the two effects. In the mock-Walker circulation experiments, it seems that the two cloud heating effects more or less cancel one another, depending on how efficiently the atmosphere can remove the excess energy from ACRE. In order for this cancellation to continue in a warmer climate, the energy transport out of the deep convective regions would need to keep pace with the increase in ACRE. This is entirely

possible, as ACRE has been shown to be an important driver in the tropical circulation (see Ramanathan, 1987; Slingo and Slingo, 1988, 1991; Stuhlmann and Smith, 1988a,b; Randall et al., 1989; Sherwood et al., 1994; Zhang and Rossow, 1997; Sohn, 1999; Bergman and Hendon, 2000b,a; Raymond, 2000; Tian et al., 2001; Tian and Ramanathan, 2002, 2003; Lee and Yoo, 2014; Harrop and Hartmann, 2015). In our experiments, ACRE increases by about 1-2% [$\text{W}/\text{m}^2/\text{K}$] when normalized by cloud fraction in the limited domain simulations. The mock-Walker circulation experiments show no consistent response of ACRE to SST.

The total change in ACRE is likely to be small owing to the opposing ways in which the rising of clouds via the FAT mechanism and the decrease in cloud abundance through the circulation slowdown. Comparing integrated domain mean ACRE across the warming experiments in our simulations, we see that ACRE does indeed show little change with sea surface temperature increase. As a result, the indirect cloud heating effect is likely to remain nearly the same as the climate warms. The direct cloud heating effect, the destabilization of the cloud layer, however, is still likely to increase, since it will be impacted more by the clouds rising. Climate models generally predict a reduction in cloud amount over the tropical warm pool (Zelinka et al., 2012), which is in agreement with a reduction in mass flux. It is unlikely that the physics of the anvil spreading by radiative destabilization is captured by these climate models, and as a result, the models may be over-predicting the reduction in cloud amount due to warming in the tropical warm pool. Since the direct cloud heating effect leads to more high, thin clouds, not having the direct cloud heating effect in GCMs may also lead to an underprediction in climate sensitivity.

An increase in ACRE with warming is also likely to thin the clouds and warm the clouds. While these effects tend to cancel one another at the top of the atmosphere, it does so by reducing both the SWCRE and LWCRE. The SWCRE and LWCRE offsetting one another can be thought of as repartitioning energy between the ocean and the atmosphere (Zhang and Rossow, 1997; Tian et al., 2001). Thus, the thinning and warming of the clouds owing

to an increase in ACRE, would effectively be partitioning more energy into the ocean at the expense of the atmosphere.

The FAT response is a feedback, but there is great interest in the so-called fast adjustments that are not dependent on surface temperature change. An instantaneous doubling of CO_2 , for example, can elicit cloud changes in much the same way as ACRE. Making the atmosphere more opaque would limit the surface and lower atmospheric upwelling longwave radiation and thus, reduce the direct cloud heating effect. As we have seen above, a reduction in the direct cloud heating effect will reduce high cloud fraction, it will shift the cloud optical depth toward thicker values, cool the clouds, and decrease the CRE. The fast responses of clouds are typically expected to increase CRE by reducing cloud fraction, so the ACRE response may negate some of that change.

4.4 Conclusions

We show that the atmospheric cloud radiative effect (ACRE) affects the high cloud fraction in two ways. First, the turbulence generated by warming at cloud bottom and cooling at cloud top enhances the areal extent of the cloud, consistent with earlier findings (e.g. Fu et al., 1995). Second, the net ACRE is a heating term that stabilizes the atmosphere and reduces cloud cover. In a limited area domain, the stability effect is the dominant control. In a mock-Walker circulation the two effects largely cancel one another, owing to the ability of the large-scale circulation to transport heat from the warm pool to the cold pool, where it is removed by emission to space.

The cloud fraction is not the only aspect of the clouds that ACRE changes. In consequence of the direct cloud radiative heating effect, the average cloud optical depth decreases and cloud top temperature increases, and these are reflected in an increase in the top of atmosphere cloud radiative effect (CRE). By breaking down the change in CRE into its component cloud fraction, clear-sky albedo, clear-sky OLR, cloudy-sky albedo, cloudy-sky

OLR, cloud fraction, and covariance terms, we have shown that the cloud fraction, cloudy albedo, and cloudy OLR contribute most to the total change in top-of-atmosphere fluxes. The cloudy albedo and cloudy OLR, however, largely offset one another, making the total increase in CRE similar to that driven by cloud fraction changes alone. We have also done these breakdowns for the total cloud heating effect, the indirect cloud heating effect, and for changes in sea surface temperature. We focus on high clouds (those whose cloud top pressure is less than 440 hPa). In short, we find ACRE increases CRE in all experiments (though the effect is larger in the limited area domain simulations than in the mock-Walker experiments). The increase in CRE largely comes from the direct interactions of clouds and radiation, while the cloud-radiative heating induced stability changes tend to decrease CRE. The cloud changes (albedo, OLR, and fraction) tend to be the dominant factors in determining CRE while the clear-sky and covariance terms tend to be secondary. Warming also leads to a small increase in CRE of about $0\text{--}1\text{ W/m}^2$, owing to changes in clear-sky OLR and cloudy albedo offsetting changes in cloudy-sky OLR.

The total CRE changes are often the result of cancellation between different factors. For example, cloud fraction and albedo changes tend to increase CRE while OLR changes tend to decrease CRE. In other words, not only does ACRE increase the occurrence of high, optically thin clouds, but it also thins and warms those clouds, reducing their albedo while enhancing their OLR. It is worth pointing out that even CRE changes of a few W/m^2 are large percentage changes from the mean because the net cloud radiative effect is close to zero owing to the near cancellation of the shortwave and longwave components of CRE.

The stabilization of the vertical temperature profile tendencies by ACRE, which we show to be the dominant factor in the limited area domains, is mitigated in the mock-Walker experiments. In our warming experiments, the gradient in SST is held fixed, which may exert an unrealistic control on the energy transport from the warm pool to the cold pool. It is unclear whether the balance in cloud fraction enhancement and reduction owing to the

cloud-radiation interactions will be maintained if the SST gradients were to be enhanced or reduced.

It is important to note that by using fixed sea surface temperatures, we are limiting our analysis of the cloud-radiation interaction to their within-atmosphere component only. In the real world, the cloud-radiation interactions could also have a profound effect on the surface budget, which is likely to further influence the structure of convection in the tropics. Much of the changes in cloud amount for the direct cloud heating effect occur for high, thin clouds, which do not have much impact on the surface energy budget, so the atmospheric influence is still likely to be the dominant component. The changes in clouds for the indirect cloud heating effect, however, show larger changes for cloud types with high optical depths, which do have a strong surface-forcing component. Future work should be done to quantify how the surface and atmospheric cloud radiative effects work in tandem to alter the tropical circulation patterns.

We have also shown that the direct cloud heating effect increases the ice water path. The increase in cloud fraction cannot simply be thought of as a spreading out of the existing clouds. Given that the mass flux in the convective cores does not change as a result of the direct cloud heating effect, the increase in cloud mass cannot be explained by an increase in convection. Instead, it results from enhanced upward transport of moisture in the anvil and cirrus clouds.

We have also discussed the potential impacts our results have for understanding cloud feedbacks. Since ACRE is expected to increase in a warmer climate owing to the rising of clouds, we can expect the destabilization of the clouds, the direct cloud heating effect, to increase as well. In our experiments, we find an increase of ACRE per unit cloud fraction of about 1-2% $\text{W}/\text{m}^2/\text{K}$. Models that are unable to resolve the within-cloud circulations may miss out on this cloud spreading as well as the thinning and warming that accompany it.

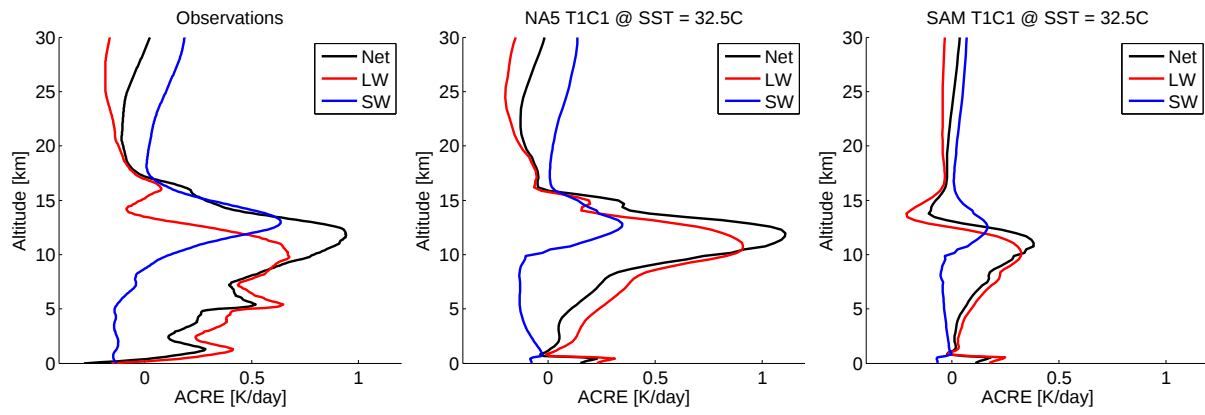


Figure 4.1: Atmospheric Cloud Radiative Effect (ACRE) profiles for (left) observations, (middle) NA5 microphysics, and (right) SAM base microphysics. The longwave contribution is in red, the shortwave in blue, and the net in black.

Table 4.1: Changes in top-of-atmosphere cloud radiative effect (CRE) for high clouds only (CTP < 440 hPa). All values are expressed in W/m². A “w” indicates a mock-Walker simulation. The factors use the same notation as equation (4.2).

C1–C0	ΔA_i	$\Delta \alpha_{\text{clr}}$	$\Delta \alpha_i$	ΔF_{clr}	ΔF_i	covariance	total
T1 28	1.81	-0.00	0.78	0.02	-0.88	-0.19	1.55
T0 28	5.35	0.01	1.51	0.34	-1.49	-0.07	5.65
T1 32	0.99	0.00	0.52	0.00	-0.39	0.02	1.14
T0 32	4.89	0.00	1.17	0.18	-1.02	-0.15	5.07
T1 28w	0.41	-0.00	0.39	0.28	-0.38	0.09	0.78
T0 28w	1.06	-0.00	-0.07	0.09	0.18	0.06	1.32
T1 32w	0.09	0.00	0.30	0.64	-0.27	0.45	1.21
T0 32w	0.29	0.00	0.13	0.59	-0.00	0.25	1.26
T1–T0	ΔA_i	$\Delta \alpha_{\text{clr}}$	$\Delta \alpha_i$	ΔF_{clr}	ΔF_i	covariance	total
C1 28	1.59	-0.08	4.67	-0.45	-11.92	4.75	-1.44
C0 28	4.06	-0.04	1.97	0.02	-5.46	2.11	2.66
C1 32	0.24	-0.07	2.95	-0.31	-8.68	3.53	-2.33
C0 32	3.12	-0.04	1.38	-0.01	-4.69	1.84	1.60
C1 28w	0.01	-0.02	0.73	-0.56	-1.25	0.23	-0.87
C0 28w	0.44	-0.01	-0.05	-0.66	-0.27	0.22	-0.33
C1 32w	-1.33	-0.00	2.30	2.22	-4.13	-0.04	-0.98
C0 32w	-0.93	-0.01	1.28	1.19	-2.46	-0.01	-0.93
32–28	ΔA_i	$\Delta \alpha_{\text{clr}}$	$\Delta \alpha_i$	ΔF_{clr}	ΔF_i	covariance	total
T1 C1	0.23	-0.11	1.07	1.41	-2.82	0.16	-0.05
T1 C0	0.75	-0.06	0.76	0.76	-1.90	0.04	0.36
T0 C1	0.89	-0.27	4.23	3.47	-8.30	0.82	0.85
T0 C0	1.21	-0.15	3.18	2.13	-5.29	0.34	1.42
T1 C1w	-0.54	-0.05	2.61	2.96	-5.16	0.64	0.46
T1 C0w	-0.18	-0.04	2.08	1.68	-3.63	0.11	0.02
T0 C1w	0.99	-0.08	1.99	1.22	-3.73	0.18	0.57
T0 C0w	1.42	-0.06	1.24	0.44	-2.35	-0.06	0.63
T1C1– T0C0	ΔA_i	$\Delta \alpha_{\text{clr}}$	$\Delta \alpha_i$	ΔF_{clr}	ΔF_i	covariance	total
28	7.15	-0.04	3.91	0.08	-7.83	0.93	4.21
32	5.32	-0.04	2.35	0.00	-5.38	0.48	2.75
28w	1.14	-0.02	0.45	-0.31	-0.71	-0.10	0.45
32w	-0.79	0.00	1.77	2.22	-2.90	-0.02	0.28

Table 4.2: Averages for precipitation, ACRE, high cloud CRE, and high cloud fraction (standard deviation in parentheses).

Name	Precipitation (mm/day)	ACRE (W/m^2)	High CRE (W/m^2)	High CF (%)
T1C1 28	2.5 (0.35)	27 (5.8)	-0.58 (2.0)	27 (6.8)
T1C0 28	2.5 (0.38)	16 (4.5)	-2.1 (2.1)	14 (4.7)
T0C1 28	3.1 (0.59)	78 (11)	0.86 (5.6)	68 (11)
T0C0 28	3.1 (0.48)	55 (7.5)	-4.8 (4.2)	39 (7.3)
T1C1 32	3.0 (0.46)	28 (6.4)	-0.63 (2.1)	25 (7.4)
T1C0 32	3.0 (0.47)	18 (4.8)	-1.8 (2.0)	14 (4.6)
T0C1 32	3.7 (0.69)	76 (12)	1.7 (5.7)	63 (12)
T0C0 32	3.7 (0.69)	52 (10)	-3.4 (4.2)	36 (9.3)
Name	Precipitation (mm/day)	ACRE (W/m^2)	High CRE (W/m^2)	High CF (%)
T1C1 28w	3.4 (1.1)	8.9 (3.2)	-0.87 (1.1)	15 (3.7)
T1C0 28w	3.4 (1.2)	6.9 (3.6)	-1.7 (1.2)	10 (3.4)
T0C1 28w	3.8 (2.2)	15 (5.2)	-0.0 (2.0)	19 (4.8)
T0C0 28w	3.7 (2.0)	11 (5.1)	-1.3 (2.0)	13 (4.4)
T1C1 32w	4.6 (2.9)	7.3 (4.3)	-0.41 (1.6)	15 (4.2)
T1C0 32w	4.4 (2.2)	2.6 (3.1)	-1.6 (1.2)	9.4 (3.0)
T0C1 32w	4.3 (2.4)	17 (4.2)	0.58 (1.9)	21 (4.1)
T0C0 32w	4.1 (2.2)	13 (4.9)	-0.69 (2.0)	15 (4.4)

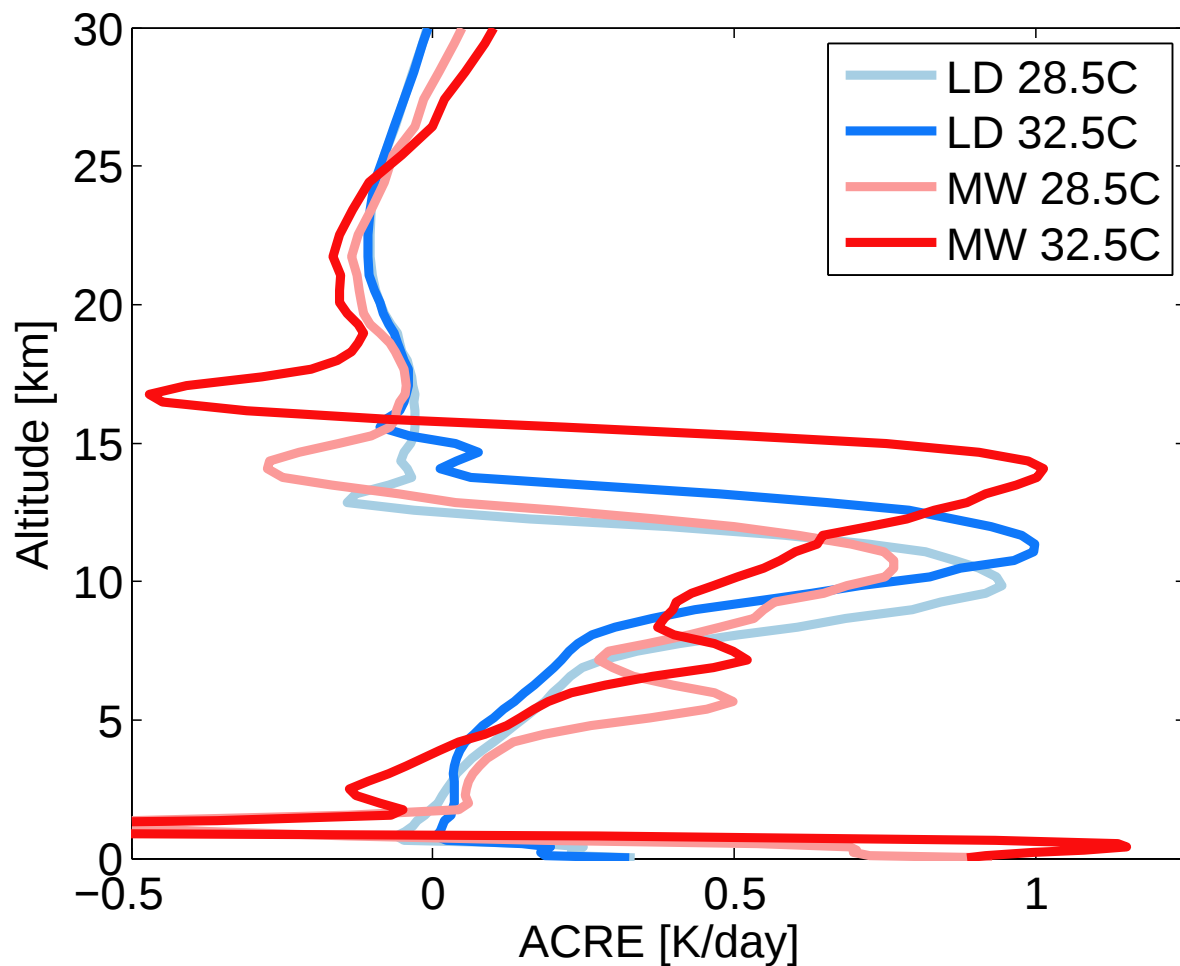


Figure 4.2: Atmospheric Cloud Radiative Effect (ACRE) profiles for the limited domain experiments (blue) and the mock-Walker circulation experiments (red). The lighter lines are for the 28.5°C experiments, and the darker lines are for the 32.5°C experiments. For the limited domain experiments, the ACRE profile is the domain average. For the mock-Walker circulation experiments, ACRE is averaged over only the warm pool (defined as the half of the domain where the sea surface temperatures exceed the average).

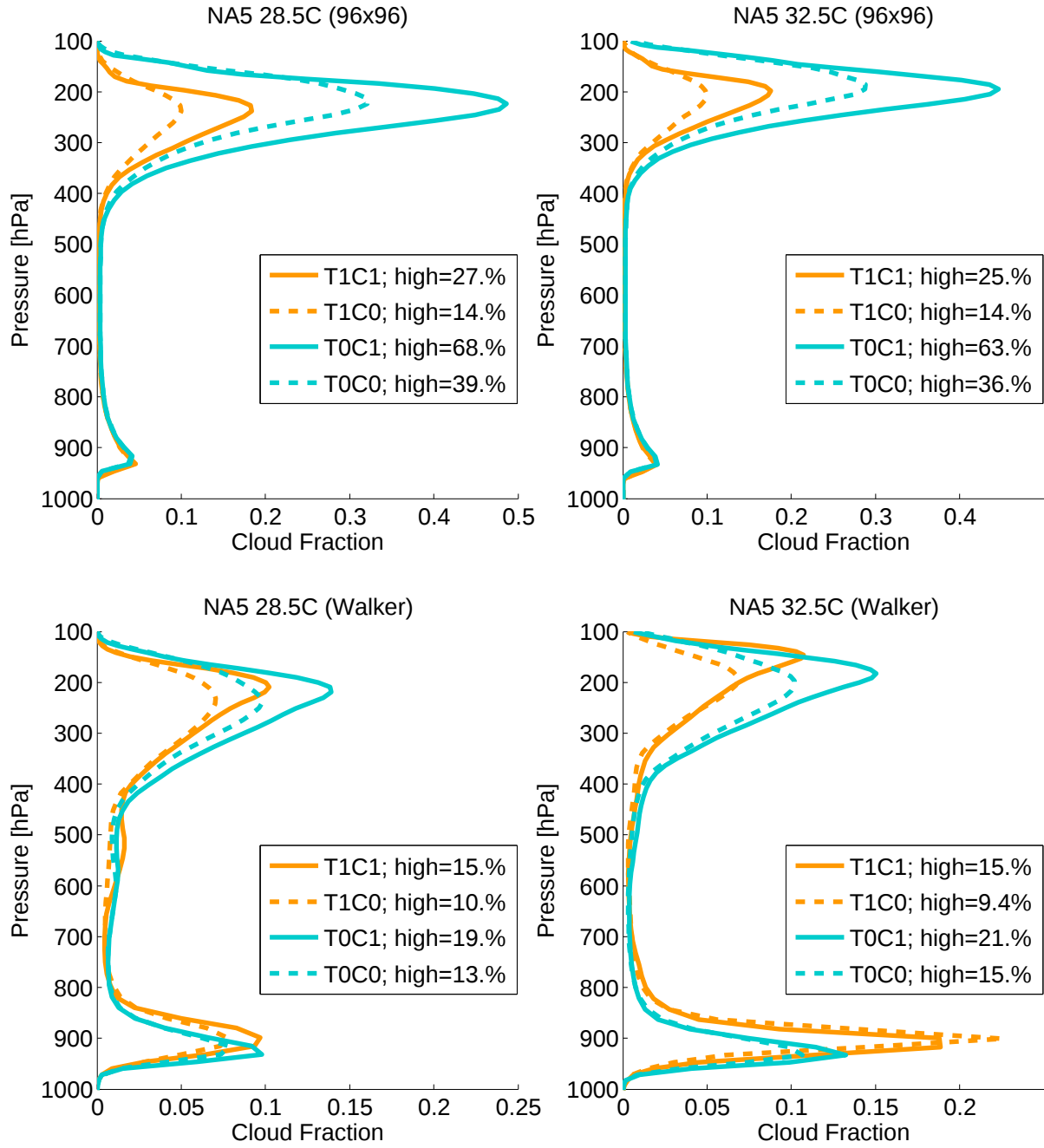


Figure 4.3: Cloud fraction profiles for (top) the limited domain experiments and (bottom) the mock-Walker circulation experiments. (Left) are the profiles for SST = 28.5°C and (right) are the profiles for SST = 32.5°C. The domain mean high cloud fraction (CTP < 440 hPa) is given in the legend of each figure.

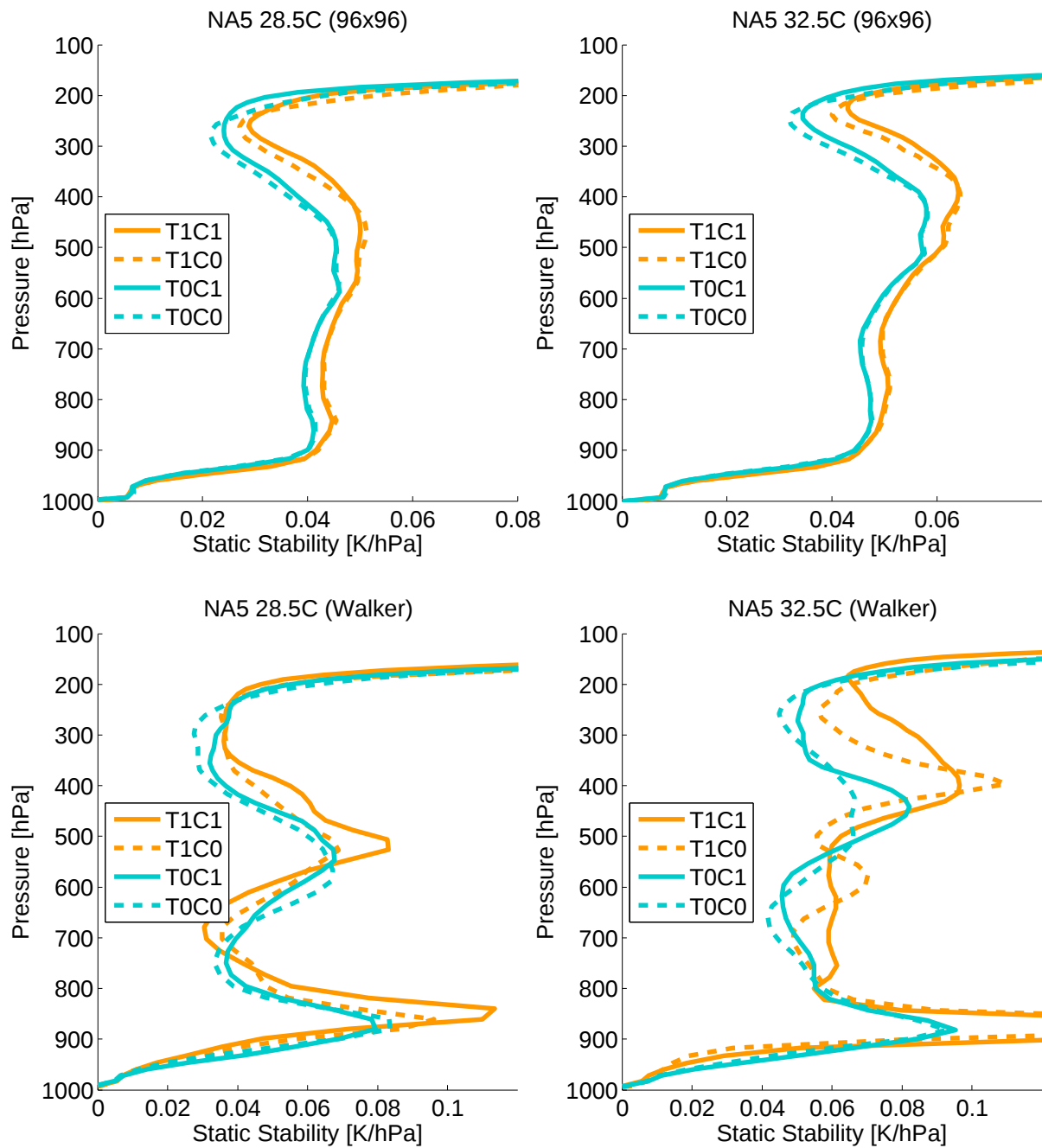


Figure 4.4: Same as Fig. 4.3 only for the stability profiles of the various experiments.

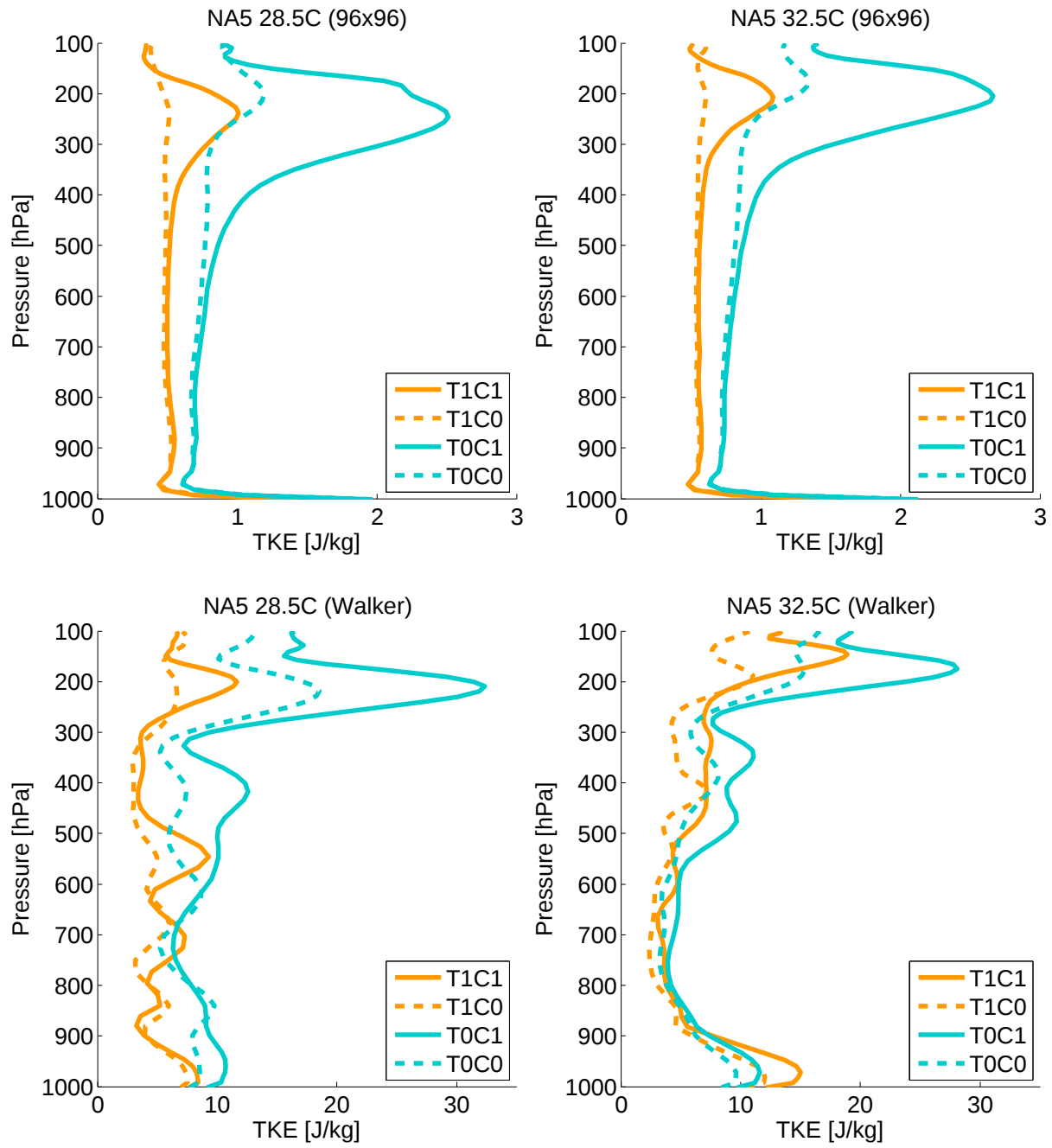


Figure 4.5: Same as Fig. 4.3 only for turbulent kinetic energy (TKE) profiles.

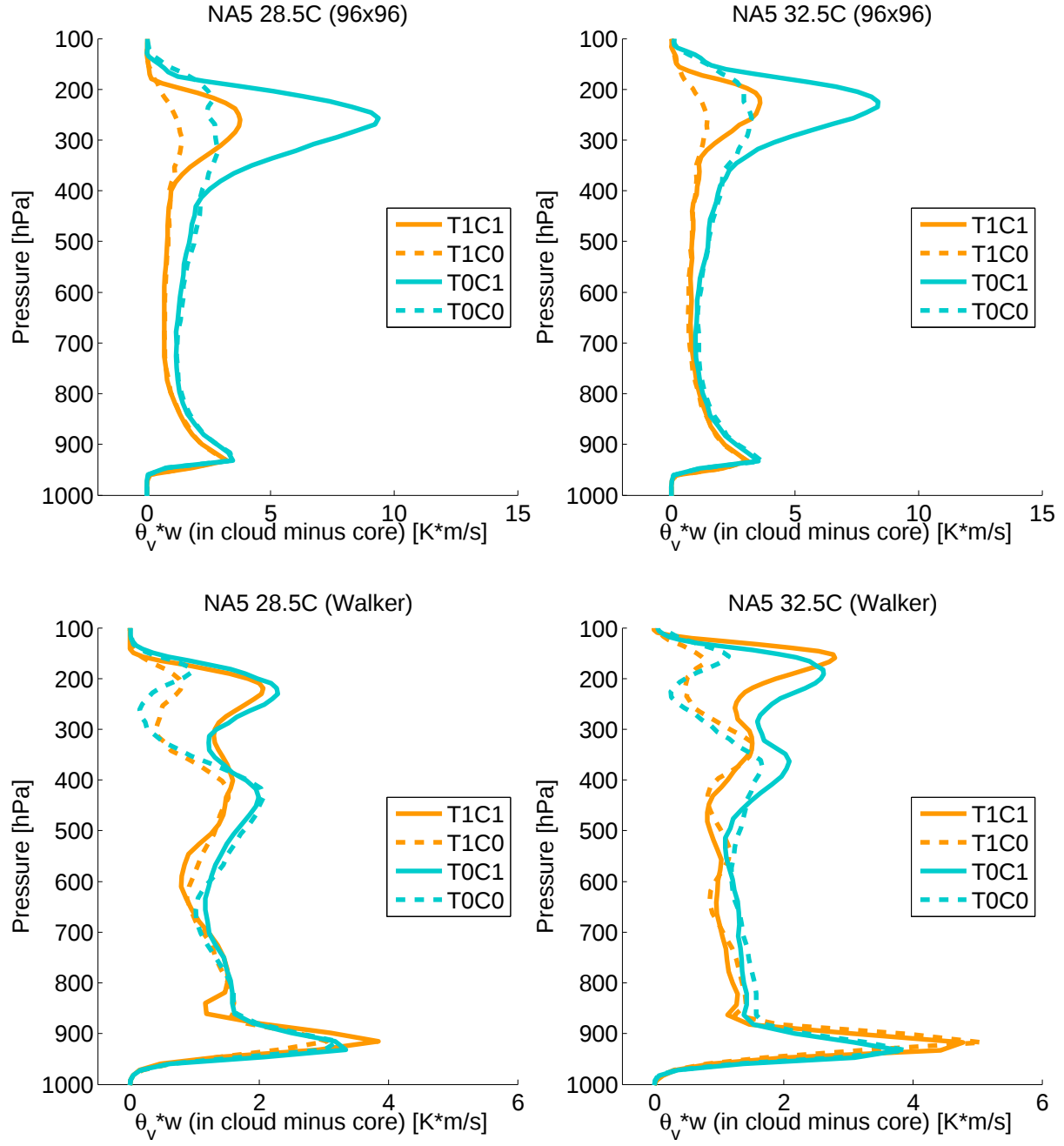


Figure 4.6: Same as Fig. 4.3, but for liquid/ice moist static energy (h_L) transport in cloudy areas outside the convective cores. Positive values indicate upward energy transport.

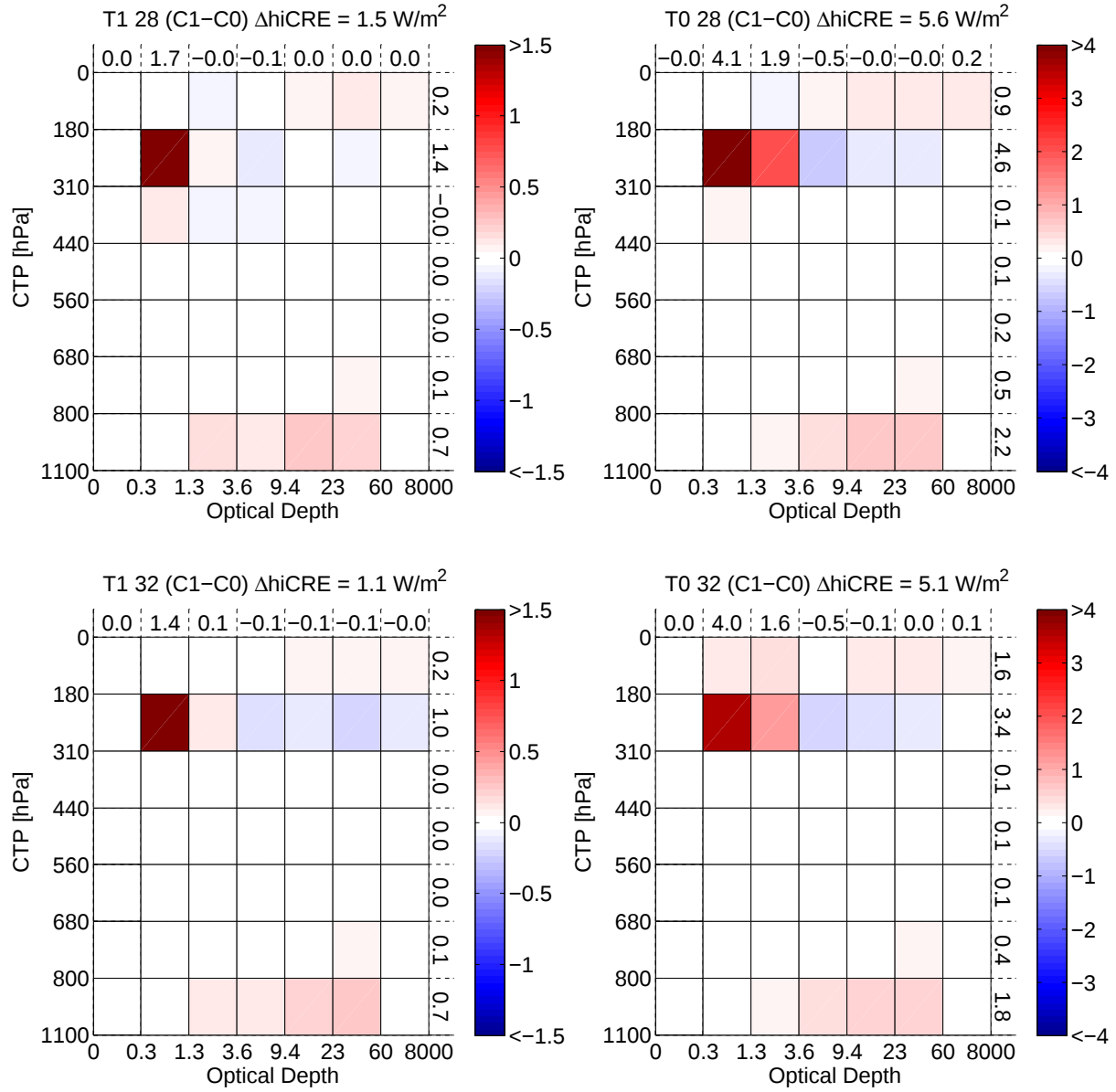


Figure 4.7: Change in CRE for each bin of the ISCCP histogram. The total high cloud ΔCRE is given in the title of each panel. The top row is for SST = 28.5°C and the bottom row is for SST = 32.5°C. The left column is for the strong stability experiments (T1) and the right column is for the weak stability experiments (T0). Note that the color scale changes between the left column and the right column. The numbers along the top are the change in high cloud CRE for each optical depth bin summed over all cloud top pressures. The numbers along the right side are the change in all-cloud CRE for each CTP bin summed over all optical depths.

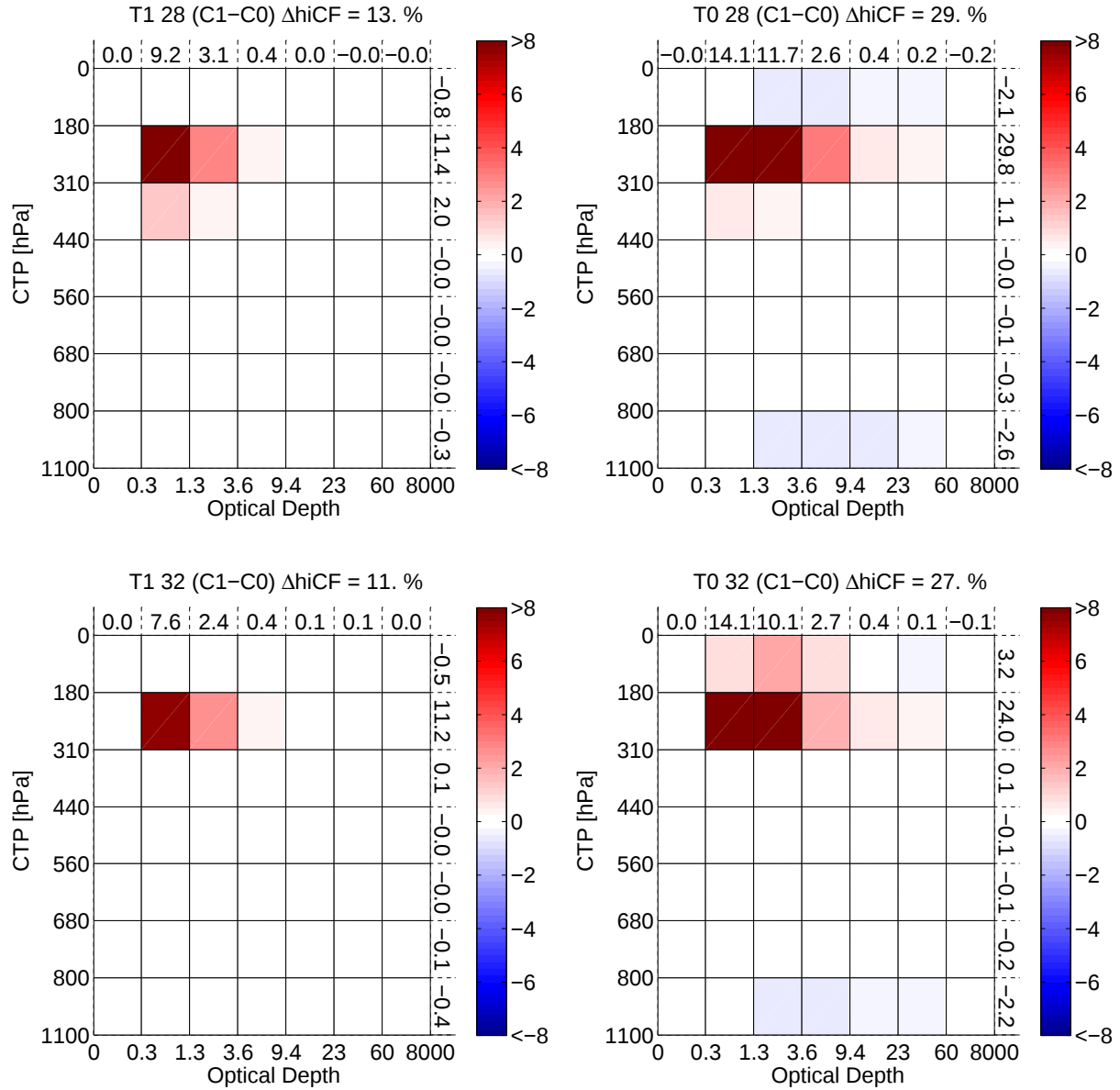


Figure 4.8: Change in cloud fraction for each bin of the ISCCP histogram. The total high cloud fraction change is given in the title of each panel. The top row is for SST = 28.5°C and the bottom row is for SST = 32.5°C. The left column is for the strong stability experiments (T1) and the right column is for the weak stability experiments (T0). The numbers along the top are the change in high cloud fraction for each optical depth bin summed over all cloud top pressures. The numbers along the right side are the change in total cloud fraction for each CTP bin summed over all optical depths.

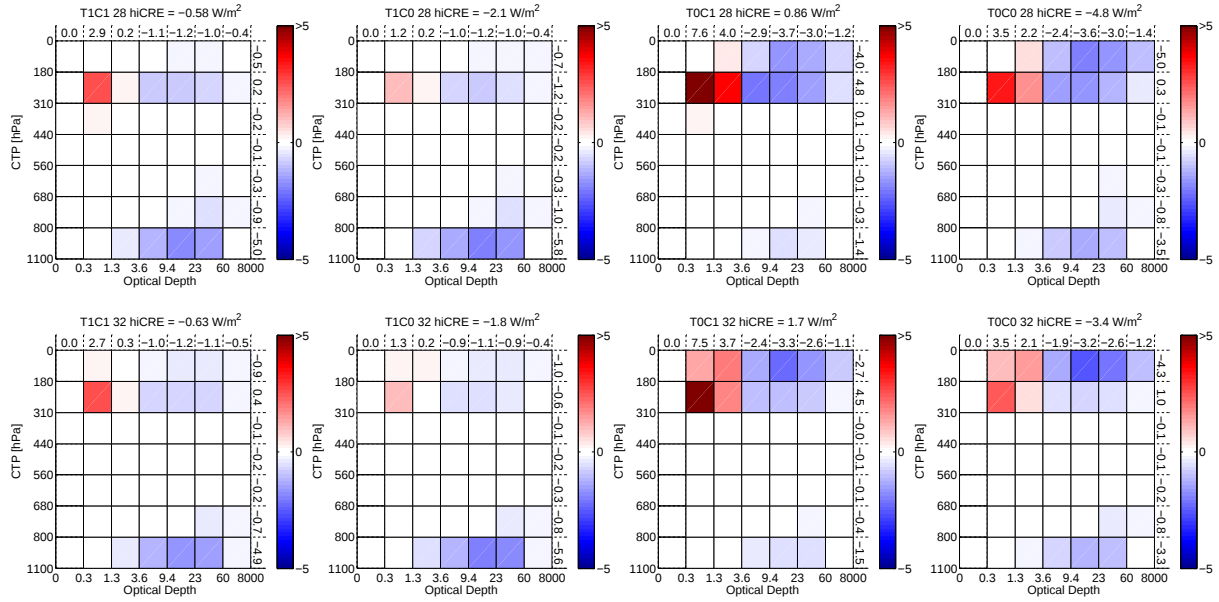


Figure 4.9: CRE ISCCP histograms for the limited domain experiments. For each histogram bin, the CRE is calculated as the summand of equation (4.1). The top row is for SST = 28.5°C and the bottom row is for SST = 32.5°C. Like Fig. 4.7, the numbers along the top are the change in high cloud CRE for each optical depth bin summed over all cloud top pressures. The numbers along the right side are the change in all-cloud CRE for each CTP bin summed over all optical depths.

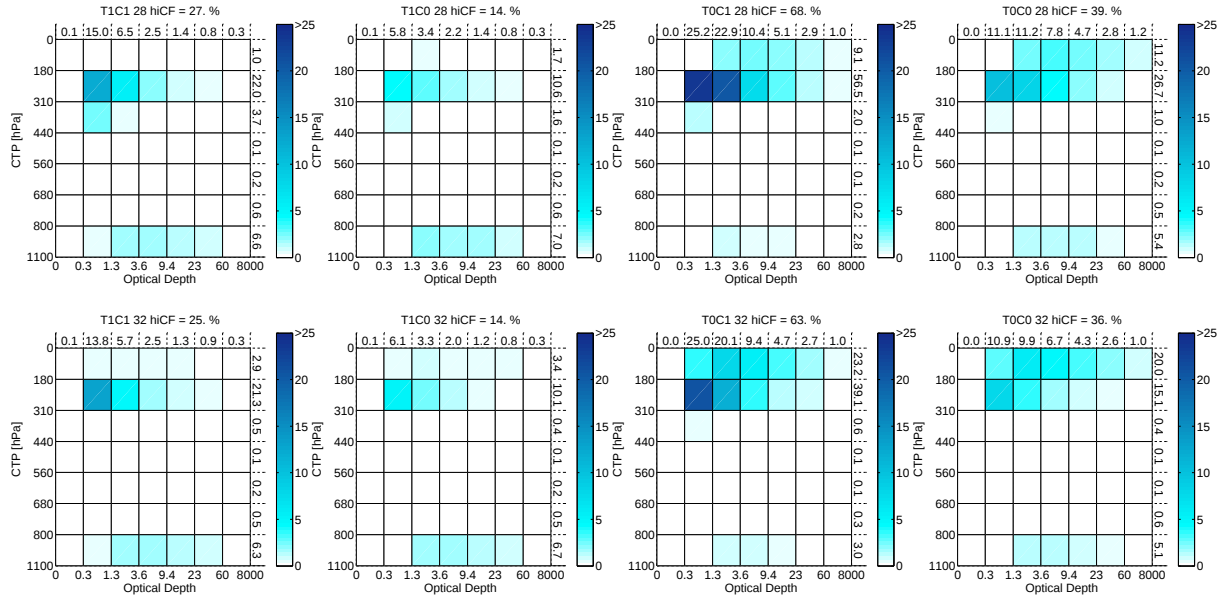


Figure 4.10: Cloud fraction ISCCP histograms for the limited domain experiments. The value of each histogram is A_i from equation (4.1). The top row is for SST = 28.5°C and the bottom row is for SST = 32.5°C. Like Fig. 4.8, the numbers along the top are the change in high cloud for each optical depth bin summed over all cloud top pressures. The numbers along the right side are the change in all-cloud fraction for each CTP bin summed over all optical depths.

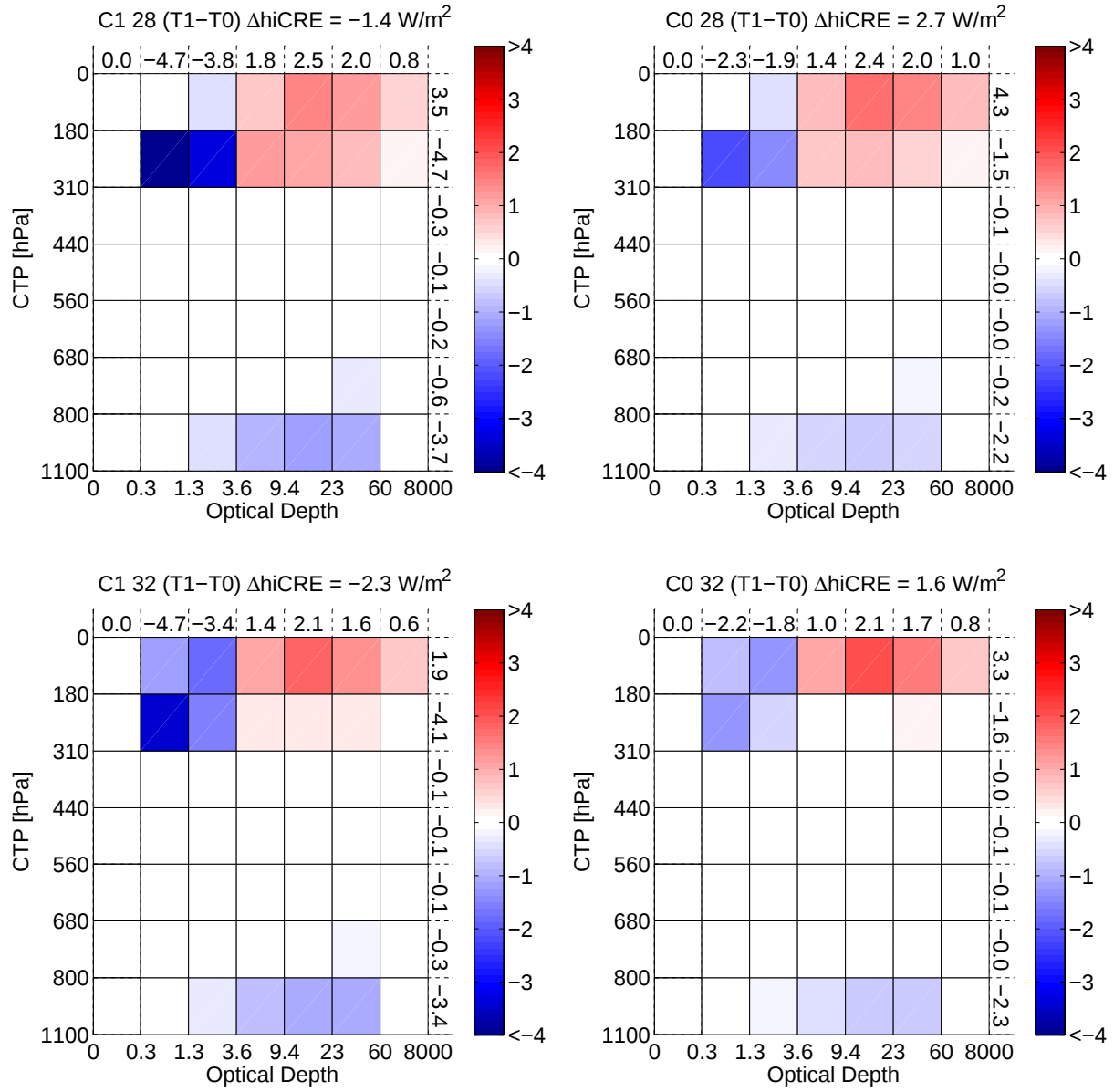


Figure 4.11: Same as Fig. 4.7, but for differences in the indirect cloud heating effect ($T1-T0$).

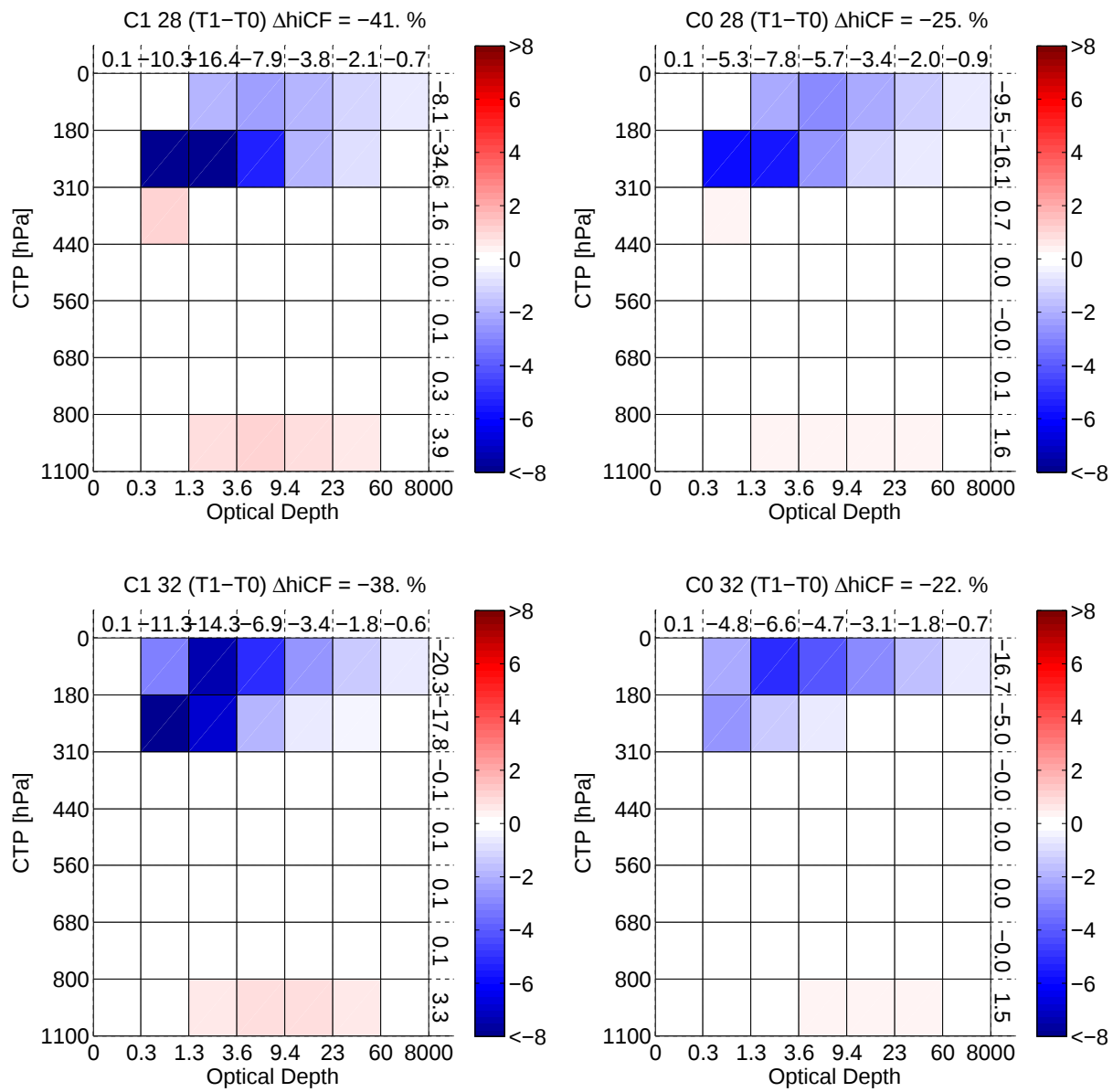


Figure 4.12: Same as Fig. 4.8, but for differences in the indirect cloud heating effect (T1-T0).

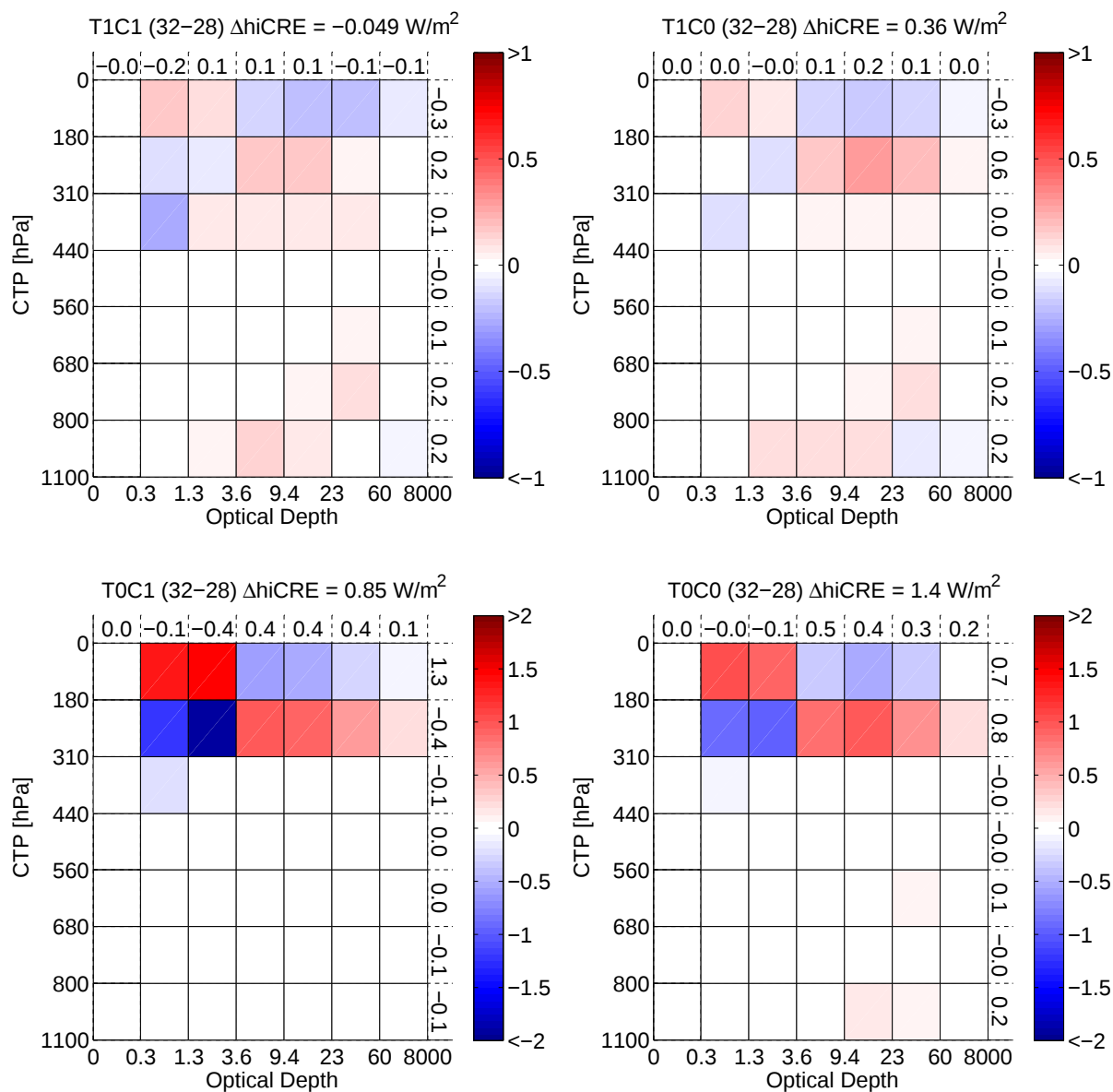


Figure 4.13: Same as Fig. 4.7, but for differences in sea surface temperature (32.5°C–28.5°C).

Chapter 5

CONCLUSIONS

In this thesis, we present work on the role of the Atmospheric Cloud Radiative Effect (ACRE) on tropical convection. We used a collection of reanalysis data, satellite observations, cloud-resolving model results, and aqua planet general circulation models to test our hypotheses. Our results illuminate several interesting relationships between ACRE and tropical convection.

In chapter 2, we investigated the climatological relationship between the Atmospheric Convective Radiative Effect (ACoRE), the combination of ACRE and the Atmospheric Moisture Radiative Effect (AMRE), and atmospheric heat transport (AHT). We first constructed a novel method to calculate AMRE. We then showed that in regions of tropical deep convection, it is positive and thus, acts to heat the atmosphere in the same way ACRE does. We have updated the mean values presented by Tian et al. (2001) with a new climatology based on a combination of reanalysis and satellite data. We found mean ACoRE values account for roughly two thirds of the total AHT. We further showed that ACoRE and AHT are extremely well-correlated on monthly timescales, increasing at a roughly one-to-one rate with convective activity over the tropical warm pool. This strong correlation results from the insensitivity of the surface turbulent heat fluxes to convection.

Additionally, we found that the top-of-atmosphere (TOA) net radiation is insensitive to average vertical velocity owing to an offset between AMRE and CRE. Performing principal component analysis on MODIS data allowed us to show that increases in cloud amount and average cloud optical depth decrease CRE as convective activity intensifies, but the additional upper tropospheric water vapor absorbs enough longwave radiation to counterbalance that

decrease in CRE. This new result highlights the importance of the covariance of water vapor and clouds on the radiation budget. It is unclear from this research whether the insensitivity of TOA net radiation to vertical motion will be maintained in a warming climate. If it does not, then strong feedbacks may arise for any change in circulation strength or convective activity.

In chapter 3, we explored the role of ACRE in determining the location of the intertropical convergence zone (ITCZ). We used zonally symmetric aquaplanet models to examine the response of the ITCZ position to the climatological ACRE. Despite the varying resolutions, physical parameterizations, and dynamics within these models, they all consistently show that ACRE contracts the ITCZ toward the equator, where sea surface temperatures are warmest.

While previous research had found seemingly conflicting results, we have proposed a mechanism that, for the first time, consistently explains the role of ACRE on the ITCZ location across both our results and the previous research. The physical process by which ACRE contracts the ITCZ equatorward is by increasing the environmental stability profile, which we measure by the Convective Available Potential Energy (CAPE). We make use of the mechanism proposed by Landu et al. (2014) (similar to the “upped-ante” mechanism of Neelin et al., 2003) to illustrate the physical processes causing the equatorward contraction of the ITCZ. The abundant high clouds heat the upper troposphere by absorbing longwave and shortwave radiation (with longwave dominating), requiring greater surface moist static energy (MSE) to initiate convection. The increase in stability requires the ITCZ to contract toward warmer sea surface temperatures (near the equator) where the surface MSE is high enough to initiate convection.

Despite the increase in stability, the circulation strength actually increases. In other words, not only does the upward branch of the Hadley circulation (the ITCZ) contract equatorward, but it also intensifies. We explain this seemingly paradoxical result as a greater

need to export MSE from the tropics when cloud-radiation interactions are present. The enhanced poleward energy transport comes from the meridional gradient in ACRE. While ACRE is positive in the tropics owing to the abundance of high clouds, it is negative in the extra-tropics. In theory, the ocean could transport the energy required to balance the ACRE gradient. While there are some changes in implied ocean heat transport (what the ocean heat transport would need to be to maintain the fixed SSTs used), we find that the atmosphere supplies most of the poleward energy transport. We might expect this from the fact that the tropical cloud heating occurs in the upper troposphere, but it cannot be known for certain from simulations using fixed sea surface temperatures. This increase in the poleward transport of energy is consistent with the results of chapter 2.

It is worth noting that our results of the ITCZ position to ACRE are for the mean climatological ACRE. Results from warming experiments by Voigt and Shaw (2015) show that cloud feedbacks do not necessarily mirror their climatological pattern and thus, the ITCZ may either contract or expand in a warmer climate owing to the cloud-radiative heating changes. The warming results of Voigt and Shaw (2015) agree with the basic mechanism we have presented in chapter 3, suggesting the underlying physics is well-understood by the mechanism described here and is robust within models.

In chapter 4, we showed ACRE affects the high cloud fraction in two distinct ways. The first way is through turbulence generated by the radiative destabilization of the cloud layer. The second way is through the tropospheric stability increase owing to the net warming of the upper tropical troposphere. These two pathways to altering cloud fraction by ACRE have opposing effects. The first (which we term the direct cloud heating effect) enhances the cloud fraction, consistent with earlier findings (e.g. Fu et al., 1995). The second (which we term the indirect cloud heating effect) diminishes the cloud fraction. We have combined these two effects in equilibrium experiments for the first time. We have shown that in limited domain, uniform sea surface temperature experiments, the stability effect is the dominant

control. In mock-Walker circulation type experiments, however, the direct and indirect cloud heating effects largely offset one another, such that the cloud fraction is largely insensitive to ACRE.

In addition to the cloud fraction response, we also examined the response of the top-of-atmosphere cloud radiative effect (CRE) to ACRE. Again, the direct and indirect cloud heating effects differ in their role, but it is slightly more complicated than the cloud fraction response. While Fu et al. (1995) showed that including ACRE reduces OLR owing to increased cloud cover, we have expanded on this idea and have broken down the top-of-atmosphere CRE owing to high clouds into its component cloud macrophysical properties. By breaking down the CRE change into its component cloud fraction, clear-sky albedo, cloudy albedo, clear-sky OLR, cloudy OLR, and a covariance term, we identified which features of convection account for the total CRE changes. It is not surprising that the cloud radiative properties exhibit larger contribution to CRE change than the clear-sky atmosphere. These results offer a new understanding of how cloud radiative heating within the atmosphere alters the top-of-atmosphere radiation.

For the direct cloud heating effect, the clouds thin (lower albedo) and warm (greater OLR) and these effects largely offset one another such that the net CRE increase reflects the cloud amount increases. We showed that most of the cloud amount increase is for those cloud types whose CRE is positive, thus increasing the cloud radiative effect. We focused on high cloud changes (cloud top pressure < 440 hPa) and the CRE response to the high clouds since the resolution of the cloud-resolving model is insufficient to accurately resolve low clouds.

ACRE increases CRE in all experiments with the effect being the largest in the limited domain experiments. The increase in CRE primarily results from the direct cloud heating effect. The indirect cloud heating effect, on the other hand, tends to decrease CRE. It is worth noting that even small CRE changes of a few W/m^2 are a large percentage change

given that the mean CRE is near zero.

We showed that the indirect cloud heating effect depends strongly on the experimental setup. It is the dominant control in the limited area domain simulations, while it is equal in magnitude to the direct cloud heating effect in the mock-Walker circulation experiments. It is unclear from our experiments whether the relative magnitudes of these effects will continue to be equal if the gradient in sea surface temperature were to change.

ACRE also plays an important role in cloud feedbacks and rapid adjustments. We have shown that ACRE increases by about 1–2% $\text{W/m}^2/\text{K}$. Models that are unable to resolve the within cloud circulations may miss out on this cloud spreading as well as the thinning and warming that accompany it and may be missing important cloud and top-of-atmosphere radiation changes. As a result, models that cannot resolve the direct cloud heating effect may be underpredicting climate sensitivity since the direct cloud heating effect increases the top-of-atmosphere cloud radiative effect.

There are several interesting future directions that arise from this work. The first concerns gross moist stability (GMS) — an important marker for how effective convection is at diverging moist static energy. Su and Neelin (2002) and Bretherton and Sobel (2002) have already shown the importance of accounting for cloud radiative heating in determining gross moist stability. Utilizing experiments like those in chapter 4 would allow us to examine how the direct and indirect cloud heating effects alter GMS for convection. Recent work by Masunaga and L’Ecuyer (2014) and Inoue and Back (2015) highlight the importance of the second baroclinic mode in determining the divergence of MSE in tropical convection, and this mode is likely influenced by the direct cloud heating effect.

Similarly, it is important to consider the effect of ACRE on the occurrence of negative gross moist stability (NGMS; e.g., Back and Bretherton, 2006; Sobel and Neelin, 2006). Raymond et al. (2009) suggest NGMS is important for the onset of convection as well as the Madden Julian Oscillation (MJO). They further note that under the weak temperature gra-

dient approximation, NGMS corresponds to moisture convergence. The total cloud radiative effect (atmospheric plus surface) involves both heating of the atmospheric column as well as cooling of the surface. Sobel (2003) showed that this shading is critical to a coincident minimum in evaporation and maximum in precipitation. These results suggest that precipitation is driven more by moisture convergence than local surface latent heat fluxes with cloud radiative effects present. Whether the cloud radiative effects drive this mechanism or are simply controlled by the same processes, however, is unclear.

Additionally, it is important to consider coupling the atmospheric cloud radiative effects to the surface cloud radiative effects. There has already been some work in an idealized framework (see, for example, Peters and Bretherton, 2005) that suggests that the radiative effect of high clouds on convective area or circulation strength is negligible owing to cancellations between ACRE and the surface cloud radiative effect. While a reduction in surface turbulent heat fluxes may offset ACRE and maintain a fixed column MSE between experiments with and without cloud radiative heating, changes in the vertical profile of heating owing to different physical mechanisms (latent heat release versus cloud radiative heating) can still alter the nature of convection.

Finally, it is worth revisiting the Fixed Anvil Temperature (FAT) and Proportionately Higher Anvil Temperature (PHAT) hypotheses (see Hartmann and Larson, 2002; Kuang and Hartmann, 2007; Zelinka and Hartmann, 2010; Harrop and Hartmann, 2012). In the FAT hypothesis, water vapor controls the cloud top temperature for detraining convective clouds and causes the cloud top temperature to be insensitive to changes in surface temperature for a fixed relative humidity atmosphere. The idea is that the clouds detrain where the radiative cooling drops off, measured by the “radiatively-driven mass convergence,” and the temperature where this convergence occurs is set by water vapor. Zelinka and Hartmann (2010) showed that while this radiatively-driven mass convergence aligns with the cloud top temperature in climate models, as the climate warms, both the convergence and clouds

warm. Harrop and Hartmann (2012) suggested that this warming comes from fixed ozone profiles, but the results presented in chapter 4 suggests the clouds may also contribute to the warming of the clouds as the climate warms.

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VITA

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