THE ATMOSPHERIC INFLUENCE OF TROPICAL DIABATIC HEATING
ASSOCIATED WITH DEVELOPING ENSO ON INDIAN MONSOON

by

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Dedication

This work is dedicated to my family:
Byeonggak, my parents, my parents in law, Yoonji, Youwon, Mingi, Byeongwook.
I would like to deeply thank all people who helped me to complete this dissertation.

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Abstract

THE ATMOSPHERIC INFLUENCE OF TROPICAL DIABATIC HEATING ASSOCIATED WITH DEVELOPING ENSO ON INDIAN MONSOON
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In order to understand the relationship between the El Niño-Southern Oscillation (ENSO) and the Indian summer monsoon, the atmospheric response to the tropical diabatic heating is studied in a general circulation model (GCM). We modify an existing GCM by adding a relatively small diabatic heating with an idealized structure. In this method, the atmospheric circulation forced by the inserted forcing interacts with convection and existing atmospheric circulation to produce a coupled dynamical convective response (GCM effect).

In experiments using climatological sea surface temperature (SST), idealized anomalous cooling over the Maritime continent induces anomalous downward motion (similar to one branch of Walker circulation during El Niño) and asymmetric anti-cyclonic circulation extending toward India, which opposes the climatological monsoon flow. In turn, this anomalous circulation weakens the Indian monsoon precipitation and it seems to explain the relationship between ENSO and the Indian monsoon. Another coupled dynamical convective response is the response of the diabatic heating related
to surface convergence, when the idealized heating is over elevated SST. Depending on the longitudinal and latitudinal location of the forcing, the pattern of atmospheric response influencing the Indian region changes and the sensitivity of the diabatic heating response in terms of magnitude is also different.

In further experiments, a slab ocean model is applied over the western Pacific and Indian Ocean. The air-sea interaction reduces mean precipitation in the region of a large rainfall such as the equatorial Indian Ocean. When the idealized heating is inserted over the warm SST, the response of the full diabatic heating shows a negative feedback. On the other hand, the idealized cooling shows a positive feedback of its response. This is because strengthened atmospheric circulation induced by the idealized forcing (both heating and cooling) releases latent heat flux from the ocean to the atmosphere and cools down the ocean temperature. The anomalous cold SST gives the positive feedback to the heating experiments and the negative feedback to the cooling experiments.

For the more realistic experiments the idealized heating over the central/eastern Pacific is also inserted in the model in addition to the western Pacific cooling. The diabatic heating response of the wide idealized heating over the central/eastern Pacific enhances the western Pacific cooling by anomalous downward motion as compensating vertical motion over the western Pacific. The Kelvin-wave like response also appears to propagate eastward all the way to the Indian region, thus interacting with the monsoon flow. The observed diabatic heating (diagnosed) of El Niño events is used to simulate the anomalous atmospheric circulation related to the Indian monsoon. For the normal monsoon of the 1997/98 event the Indian Ocean warming is added to compensate the effect of the western Pacific cooling on India. For the dry monsoon of 1982 event negative precipitation is simulated but the atmospheric circulation anomaly is too large compared with observation.
Chapter 1: Introduction

1.1 Motivation

The Asian monsoon is one of the most energetic components of the climate system. The Indian summer monsoon (ISM), one of the Asian monsoon components, fluctuates in terms of atmospheric circulation and precipitation on interannual time scales. To understand the interannual variability of ISM, the basic question is what physical processes determine the structure and variability of the coupled ocean-atmosphere-land monsoon system (Webster et al., 1998).

In numerical models, the anomalous atmospheric circulation and monsoon rainfall have been simulated by specifying the interannual variations of tropical sea surface temperature (SST) (Annamali and Hamlitom, 2007; Chen and Yen, 1994; Ju and Slingo, 1995; Lau and Nath, 2000). These experiments were motivated by the hypothesis of Charney and Shukla (1981), who suggested that the interannual variation of the Indian monsoon is forced by slowly varying boundary conditions at the earth’s surface. Both the global atmospheric circulation and the tropical SST have been studied intensively in order to understand the interannual variability of the monsoon.

In terms of the forced seasonal-mean atmospheric circulation, an anomalous anticyclonic circulation near India is both observed and simulated during developing warm El Niño-Southern Oscillation (ENSO) events. This is understood as a Rossby wave response (Gill, 1980; Matsuno, 1966) to a negative heat source due to anomalously cold SSTs in the western Pacific/Indonesian region. The cold SSTs are associated with anomalous descent in one branch of the Walker circulation. The anomalous
anticyclonic circulation can be further related to anomalously negative rainfall in the Indian region. Similarly, developing cold ENSO events are associated with warm SSTs in the western Pacific/Indonesian region, anomalous rising motion locally, cyclonic anomalies over India and hence enhanced monsoon rainfall. Thus there is generally an inverse relationship between a typical ENSO SST index (such as Niño 3\(^1\)) and the monsoon rainfall (Trenberth, 1997).

However, this relationship between ENSO and the Indian monsoon has not been consistent throughout the long-term period (Kirtman and Shukla, 2000; Kripalani and Kulkarni, 1997; Kumar et al., 1999; Makhov et al., 2011). There are still a number of unanswered questions regarding this relationship. For example, why has the strength of the correlation declined in recent decades? Why was the rainfall over India so different during the developing warm event of 1997 (slightly above normal Indian rainfall) and the similar developing warm event of 1982 (deficient rainfall)?

The location of India also brings a difficulty in studying the atmospheric influence of ENSO on India. Many years ago Truop (1965) found that India is located near a node (zero contour) in the pressure pattern of El Niño composite, well away from the regions of anomaly extrema. Small displacements of the node would place India in either a weak positive or weak negative pressure regime compared to the long-term summer average (Truop, 1965; Webster et al., 1998).

In addition, Slingo and Annamalia (2000) suggested that subtle shifts in the domain of influence of ENSO could lead to differing responses of the monsoon. Depending on the strength and detailed configuration on an ENSO event, the response of the “Walker circulation” (i.e. regions of tropical ascent and descent) may be accompanied with anomalies of the local Hadley circulation (longitudinal maxima/minima in meridional convergence). This mixed response constitutes a more complex circulation

\(^1\)The Niño 3 region is bounded by 90°W ~ 150°W and 5°S ~ 5°N
regime than that of a purely anomalous Walker circulation, which is often thought of as the ENSO response.

To understand this complex atmospheric circulation related to ENSO and its influence on the Indian summer mean monsoon (ISM), one traditional approach is to use anomalous tropical SSTs as a boundary forcing in a full atmosphere general circulation model (GCM) (Ju and Slingo, 1995; Lau and Nath, 2000; Wang et al., 2000). However, in the full atmospheric GCM experiments, the effect of prescribed anomalous tropical SST anomalies on anomalous diabatic heating depends very much on the particular convective parameterization used in the model; differences in convective parameterization may lead to very different modeled diabatic heating. Another traditional approach is to use an idealized, dry atmospheric GCM with specified diabatic heating prescribed as forcing (Jin and Hoskins, 1995; Lin, 2009). This method has complete control over the diabatic heating, but it cannot include the feedback of the dynamics on the heating.

In this study, we use anomalous diabatic heating to force a full GCM. This approach allows more control over the diabatic heating than would be the case with forcing solely by SST. In particular, we modify an existing GCM by adding a relatively small diabatic heating with an idealized structure. The idea is to specify idealized changes in the diabatic heating, yet retain the full feedback of the dynamics on the heating in the context of the full model. For example, the idealized rising motion induced by added idealized heating will influence the model’s circulation, which in turn can further change the moist and radiative heating fields produced by the model’s parameterizations. In reality, cumulus convection (diabatic heating in the model) is coupled with the circulation and can interact with the large-scale dynamics (Watanabe and Jin, 2003). So our first general question is: how do the moist and radiative processes in the GCM modify the direct response to additional tropical
diabatic heating in the model? Another basic question is: what is the subsequent feedback of the dynamics on the model heating?

By inserting additional idealized anomalous cooling over the western Pacific in the GCM, anomalous downward motion will be simulated as a branch of the Walker circulation relevant for warm ENSO events. The downward motion over the Maritime continent lies close to the Indian region, but still does not cover India entirely in most warm events. In other experiments, additional diabatic heating is inserted over the central and eastern Pacific. So some specific questions for the forced experiments in this study are as follows:

1) How does the descending motion over the Maritime continent influence India, as appropriate for a warm ENSO event?

2) Is the impact of the additional forcing on India sensitive to the location of the downward motion?

3) Does the ascending branch of the Walker circulation in the central or eastern Pacific (also appropriate for a warm ENSO event) influence the Indian summer monsoon?

The last general question is the role of air-sea interaction in the Indian region. The importance of air-sea interaction for the Indian summer monsoon is suggested by Wang et al. (2003) and Wu and Kirtman (2005). In the regions of large mean rainfall, prescribed SST in a numerical model exaggerates precipitation, whereas atmospheric negative feedback to SST simulates reduced rainfall in an atmosphere-ocean coupled model and hence a more realistic variability of the monsoon. The anomalous circulation induced by descending motion in the Maritime continent is also closely linked to the air-sea interaction in Wang et al. (2003).
In this study, to understand what physical processes influence the Indian monsoon, the coupled dynamical-convective response to additional tropical heating is studied in an atmospheric GCM (AGCM). Data and experiments in this study are described in section 2. The comparisons of the two control runs with different types of the ocean are shown in section 3: (i) climatological SST specified over the global ocean, and (ii) a slab ocean model over the western Pacific and Indian Ocean and climatological SST specified over other basins. By applying a slab ocean model over the western Pacific and Indian Ocean, the precipitation and SST decreases in the convection region, which is the northern Indian Ocean. In section 4, the additional diabatic heating is inserted at different longitudinal and latitudinal locations to see the sensitivity of the atmospheric response and feedback of dynamics on heating in the GCM. The forced experiments are carried out with both types of SST as described above. The slab ocean model experiments are described in Section 4.4.

In the forced experiments with prescribed SST, the anomalous diabatic heating due to the GCM feedback is strong over the Indian region. On the other hand, in the forced experiments with regional coupling with the slab ocean model, the exaggerated response over Indian disappears. The feedbacks now brought into play depend on the sign of forcing: a negative feedback for additional positive diabatic heating and a negative feedback for anomalously negative diabatic heating.

In section 5, we simulate a more realistic El Niño event using composite anomalous diabatic heating: negative anomalies over the western Pacific and positive anomalies over the central/eastern Pacific are inserted to simulate a more realistic El Niño event. In section 6, observed forcing is calculated from diagnosed diabatic heating data by Chan and Nigam (2009). Two El Niño events (1982 and 1997/98) are compared with the response to the observed forcing.
1.2 Literature Review

1.2.1 Relationship between ENSO and ISM rainfall

The relationship between the Indian monsoon rainfall and the Southern Oscillation (first studied by Walker) has received considerable interest from various perspectives in the last three decades (Ju and Slingo, 1995; Lau and Nath, 2000; Lau and Yang, 1996; Liu and Yanai, 2001; Rasmusson and Carpenter, 1983; Shukla and Paolino, 1983; Soman and Slingo, 1997; Webster and Yang, 1992). Previous studies on the interannual variability and decadal variability of the Indian summer monsoon focused on the influence of slowly varying boundary forcing (Charney and Shukla, 1981), while work on the relationship between interannual and intraseasonal variability was motivated by the need to improve the prediction of the ISM.

ENSO has been known to exert the most important external forcing on Indian summer monsoon rainfall (Kumar et al., 1999; Lau and Nath, 2000; Wang et al., 2003). As Walker found out long ago, one phase of the Southern Oscillation is associated with anomalously high pressure over the western Pacific-eastern Indian Ocean and anomalously low pressure over the eastern and central Pacific (the so-called “Walker circulation”). Later it was found that this phase of the Walker circulation is associated with a developing El Niño event and influences the monsoon circulation and precipitation. Kumar et al. (1999) and Palmer et al. (1992) suggest that ENSO shifts the locations of the tropical Walker circulation and brings about a deficit (excess) of rainfall by suppressing (enhancing) the convection over the Indian region. The anomalous Walker circulation leads to a negative relationship between the eastern Pacific SSTs associated with ENSO and the ISM.

ENSO has been used as an input in the seasonal prediction of Indian summer monsoon by many researchers. However, in recent years ENSO has seemingly lost
its impact on the Indian summer monsoon (Kirtman and Shukla, 2000; Kripalani and Kulkarni, 1997; Kumar et al., 1999). In statistical analysis, ENSO-to-monsoon influence is strong in 1950-1980 and not detected in 1920-1950 and after 1980. The weakened relationship between ENSO and the Indian summer monsoon is shown in Wang et al. (2008). This is one of the complex characteristics of relationship between ENSO and ISM. The complex spatial pattern of precipitation near India during El Nino events is also shown in Wang et al. (2003).

1.2.2 Atmospheric circulation related to ENSO

The atmospheric circulation near India has been studied to understand the relationship between ISM and ENSO. A few large-scale seasonal mean circulation patterns that connect the circulation to the interannual variability of the seasonal mean all-India monsoon rainfall were studied by Straus and Krishnamurthy (2007). They found that the atmospheric circulation near India is more tightly connected to SST in the Pacific than is the Indian rainfall. Gadgil et al. (2004) and Ihara et al. (2007) also found that the Indian monsoon rainfall is related to the equatorial wind anomalies during ENSO. Ashok et al. (2004) studied the atmospheric circulation to understand the abnormal monsoon response to the 1997 El Niño event. Similar to the Gadgil et al. (2004) and Ihara et al. (2007) studies, the work of Ashok et al. (2004) focused on the effects of the Indian Ocean dipole (IOD) in SST and the local Hadley circulation on the Indian monsoon, in order to explain the weakening of the relationship between ENSO and ISM. Therefore, the anomalous atmospheric circulation near India is worthwhile to study in order to understand the relationship between ENSO and the ISM.

During the northern summer the atmospheric response to anomalous tropical
heating shows special characteristics near the Indian region compared to the winter response. In Lau and Nath (2000), the ENSO related circulation simulated by an AGCM is composited for winter and summer separately. The response to anomalous descent of the Walker circulation in the western Pacific depends on season: in winter, the anomalous anticyclonic circulation is located very close to the cooling in the western Pacific, while in summer the circulation is located at a distance from the cooling; over the Indian region negative precipitation anomalies are seen. (The precipitation anomalies are very weak in winter). Wu et al. (2006) also show the seasonal dependence of the response to tropical forcing in an AGCM. This seasonally dependent relationship between ENSO and the ISM also appears in an observational study (Ropelewski and Halpert, 1987). One question raised by these findings is why there is a stronger correlation between ENSO and Indian region precipitation during summer, when the peak of ENSO is during winter, especially as the descending motion in winter is as strong as summer over the Maritime continent.

Here, we can relate this question to our first question given earlier (namely, how the descending branch of the Walker circulation remote from India influences the ISM): during summer the response of diabatic heating may be elongated northwestward towards the Indian region. Webster (1972) explained that the distinct background circulation in each season induces a distinct Rossby wave propagation. Xie and Wang (1996) also showed that the easterly vertical shear of the Indian region during summer induces a stronger atmospheric response in the northern Hemisphere than the southern Hemisphere (asymmetry about the equator). In this study, the response to tropical heating in summer can help to explain the relationship between the ISM and ENSO.

Besides AGCM studies, simple numerical models also have been used to study the effects of mountains and diabatic heating on the mean state of the monsoon system
(Rodwell and Hoskins, 2001). Lin (2009) studied the anomalous circulation forced by diabatic anomalies in the India region and western Pacific in a simple model to study diabatic heating variability of the Asian summer monsoon. There has been considerable work on tropical wave dynamics in producing remote response to a heat source (Gill, 1980; Matsuno, 1966; Webster, 1972).

However, these studies in a dry model treated moist convection as an external heating force to the total dynamical system and did not consider the complex interaction between convection and large-scale motions. In reality, cumulus convection generating the anomalous heat source is coupled with the circulation and interactions between with the large-scale dynamics and convection can further change the circulation (Watanabe and Jin, 2003). When a convection anomaly excites an anomalous atmospheric circulation, an additional convection anomaly may be generated in a remote region from the forcing and in turn this additional heating anomaly further forces the atmosphere. Recent studies (Annamali, 2010; Watanabe and Jin, 2003) have examined moist processes in numerical models to explain the complex response of diabatic forcing in the monsoon system. Su and Neelin (2002) and Neelin and Su (2005) also pointed out that moist processes impact descent anomalies during El Niño in the Pacific. In this study, we also use a numerical model with moist physics, and the interaction between equatorial stationary waves and moist processes will be carefully examined to study their impact on precipitation anomalies.

1.2.3 Air-sea interaction

The anomalous anti-cyclonic circulation in the Indian Ocean is generated by cold SST anomalies over the Maritime Continent during a developing El Niño. Air-sea interaction also plays an important role for the maintaining anomalous monsoon circulation during a developing El Niño (summer). Wang et al. (2003) found that
the positive feedback of the air-sea interaction in response to the cold SST anomalies operates as follow. In the eastern part of the anomalous anti-cyclonic circulation, the winds over the eastern Indian Ocean induce active evaporation (more latent heat flux) and hence cold (negative) SST anomalies, which subsequently further enhance the anti-cylonic circulation.

Wang et al. (2003)’s study provides observational evidence to support this idea and in this study we will test this idea numerically by inserting anomalous cooling with air-sea coupling by applying the slab ocean model regionally. Since the slab ocean model allows energy exchange between ocean and atmosphere via surface heat fluxes, the hypothesis related to latent heat flux and SST can be tested. Oceanic process such as coastal and equatorial upwelling may further strengthen the anticyclone-induced SST dipole anomalies, so constituting a positive feedback.

Wu and Kirtman (2005), Fu and Wang (2002), and Kemball-Cook and Wang (2001) studied the role of air-sea coupling in the Asian monsoon. The simulation of Indian monsoon rainfall was improved by air-sea coupling in Fu and Wang (2002). The air-sea coupling induced cold SST in the Indian Ocean and western Pacific by interaction between surface heat flux and wind. Wu and Kirtman (2005) simulated a reasonably realistic variability of the Asian monsoon rainfall with air-sea coupling used in place of prescribed SST in the Indian region. In the regions of large mean rainfall, the atmospheric negative feedback is strong and air-sea interaction is necessary to simulate the proper monsoon rainfall. In this study the question of how air-sea interaction modifies the response of direct response of tropical heating in the Indian region will be addressed, with consideration given to both positive and negative feedback.
Table 1.1: The list of Acronyms

<table>
<thead>
<tr>
<th>Acronym</th>
<th>Description</th>
</tr>
</thead>
<tbody>
<tr>
<td>AGCM</td>
<td>Atmospheric General Circulation Model</td>
</tr>
<tr>
<td>CAM</td>
<td>Community Atmosphere Model</td>
</tr>
<tr>
<td>CCM</td>
<td>Community Climate Model</td>
</tr>
<tr>
<td>CLM</td>
<td>Community Land Model</td>
</tr>
<tr>
<td>CMAP</td>
<td>CPC Merged Analysis of precipitation</td>
</tr>
<tr>
<td>CPC</td>
<td>National Oceanic and Atmospheric Administration Climate Prediction Center</td>
</tr>
<tr>
<td>ECMWF</td>
<td>European Center for Medium-Range Weather Forecast</td>
</tr>
<tr>
<td>ENSO</td>
<td>El Niño and Southern Oscillation</td>
</tr>
<tr>
<td>ERA</td>
<td>ECMWF 40 Year Re-Analysis</td>
</tr>
<tr>
<td>GCM</td>
<td>General Circulation Model</td>
</tr>
<tr>
<td>IOD</td>
<td>Indian Ocean Dipole</td>
</tr>
<tr>
<td>ISM</td>
<td>Indian Summer Monsoon</td>
</tr>
<tr>
<td>ISMR</td>
<td>Indian Summer Monsoon Rainfall</td>
</tr>
<tr>
<td>NCAR</td>
<td>National Center for Atmospheric Research</td>
</tr>
<tr>
<td>NCEP</td>
<td>National Center for Environment Prediction</td>
</tr>
<tr>
<td>SOM</td>
<td>Slab Ocean Model</td>
</tr>
<tr>
<td>SST</td>
<td>Sea Surface Temperature</td>
</tr>
</tbody>
</table>
Chapter 2: Data and methods

2.1 Data

We use several observational and diagnosed quantities including winds, vertical motion, streamfunction at 850hPa, precipitation and diabatic heating from the surface to 100hPa (3D). The precipitation product is from National Oceanic and Atmospheric Administration Climate Prediction Center (CPC) Merged Analysis of precipitation (CMAP) (Xie and Arkin, 1997). The wind and vertical motion ($\omega$) data come from the European Center for Medium-Range Weather Forecasts (ECMWF) 40 Year Re-Analysis (ERA-40) data, and the streamfunction is calculated from the wind data. The diabatic heating data is diagnosed from the ERA-40 data (Chan and Nigam, 2009). All data sets are monthly and for analysis we use a summer mean from May to August. This summer mean is to align with the summer simulation (May to August) of this study. The ERA-40 data extends from 1957 to mid-2002 and the CMAP data extends from 1979 to 2007. The diabatic heating data (Chan and Nigam data) uses only eight months of the last year (2002) of ERA-40 data. The period that we use here is from 1979 to 2001. The summer mean (MJJA) of climatology is shown for SST (figure 2.1a), precipitation (figure 2.1b), diagnosed vertically integrated diabatic heating (figure 2.1c), streamfunction at 850hPa (figure 2.2a), wind at 850hPa (figure 2.2b), and vertical motion vertically integrated from the surface to 100hPa (figure 2.2c). The SST shown here is provided by the National Center for Atmospheric Research (NCAR) as a boundary condition of the model used in this study. The SST is interpolated from the Hadley Center SST (Rayner et al., 2003).
Figure 2.1: Summer climatology (from May to August). a) SST (°C), b) precipitation (mm/day), c) diagnosed diabatic heating (Wm$^{-2}$)
Figure 2.2: Summer climatology (from May to August). a) streamfunction at 850hPa ($\times 10^6 \text{s}^{-1}$), b) wind at 850hPa (m/s), and c) vertical motion ($\omega$) vertically integrated from the surface to 100hPa (Pa · kg/(m²s))
The three dimensional diabatic heating (Chan and Nigam, 2009; Nigam et al., 2000) is diagnosed as a residual in the thermodynamic equation using the analyzed vertical velocity ($\omega$):

$$
\overline{\dot{Q}}(x, y, p, t) = \frac{\Delta T}{\Delta t} + \nabla \cdot \nabla T + \left( \frac{p}{p_0} \right)^{\frac{R}{C_p}} \frac{\omega}{\bar{T}T} \frac{\partial \bar{T}}{\partial p} + \left( \frac{p}{p_0} \right)^{\frac{R}{C_p}} \left[ \nabla \cdot \nabla \bar{\theta} + \frac{\partial (\omega \bar{\theta})}{\partial p} \right]
$$

Here $\overline{\dot{Q}}(x, y, p, t)$ is the diagnosed monthly diabatic heating rate (in K/day), $\mathbf{v}$ the horizontal wind vector, $\omega$ the pressure vertical velocity, $\theta$ the potential temperature $[= T(p/p_0)^{R/C_p}]$, $T$ the temperature, $R$ the gas constant, and $C_p$ the specific heat at constant pressure. The over-bar denotes the monthly-mean and the prime denotes the deviation of the 6-hourly analysis from this mean (i.e., transient component). The transient component thus represents both synoptic and low-frequency (but sub-monthly) fluctuations.

The diabatic heating diagnosed from ERA-40 (Chan and Nigam, 2009) is compared with the heating structure diagnosed from National Center for Environmental Prediction (NCEP) reanalyses and ERA-15. In Nigam et al. (2000), heating diagnosis from the ERA-15 and NCEP reanalyses is used to evaluate the ENSO heating distribution produced by the NCAR Community Climate Model, version 3 (CCM3) atmospheric GCM. In this study, this diagnosed diabatic heating (Chan and Nigam data), particularly for ENSO, is used to compare the simulated diabatic heating with the inserted heating in our experiments.
2.2 Model

2.2.1 AGCM

The NCAR Community Atmosphere Model Version 3 (CAM3) will be used for the Atmosphere General Circulation Model (AGCM). The standard version, used here, has 26 vertical levels and a 42-wave triangular spectral truncation. The formulation of the physics and dynamics of CAM3 is detailed by Collins et al. (2006). A spectral Eulerian formulation is used to resolve dynamical motions along with a semi-Lagrangian transport scheme for dealing with large-scale transport of water and chemical species. The dynamical core of CAM3 is identical to that of NCAR CCM3 (Kiehl et al., 1998). A number of physics enhancements are described in Collins et al. (2006), including the physics of cloud and precipitation (Boville et al., 2006) and the parameterization of radiation (Collins et al., 2006). CAM3 also includes the Community Land Model (CLM) version 3.0 for the treatment of land surface energy exchanges. CLM3 is integrated on the same horizontal grid as CAM3, although each grid box is further divided into a hierarchy of land units, ground cover, and plant types (Oleson et al., 2004).

2.2.2 SOM

A slab-ocean model (SOM) will be used for air-sea coupling over the Indian Ocean and western Pacific. The ocean model coupled to the AGCM will be a simple thermodynamic slab mixed layer model, which is a part of the NCAR CAM3 modeling system. The depth of the mixed-layer (figure 2.3) is fixed to a climatological annual cycle with a 200 meter cap. At each grid point, the AGCM supplies heat flux to the ocean model, and the ocean model returns SST to the AGCM. The SST evolves according to the heat flux given by the AGCM. The coupling will be allowed in only
the Indian Ocean and western Pacific (a blue line box in figure 2.3). In the other ocean basins, prescribed SST will be used as a boundary condition of the AGCM. In the tropical central and eastern Pacific, the SST forcing is very strong and we expect that SST mainly influences the atmosphere (Lau and Nath, 2000, 2003), so the prescribed SST seems reasonable in these regions. In the extratropics where SSTs are relatively cold, the atmosphere forces the ocean (Lau and Nath, 1994) so the air-sea interaction is important in these regions. However, in this study, the main interest is the impact of tropical heating on the Indian region, so in order to avoid a complexity in the extratropics the prescribed SST is used.

![Figure 2.3: The annual mean depth (m) of the mixed layer. The blue line box is the domain of the slab ocean model applied in the model.](image)
2.3 Control runs

Two control runs of 21 years length are run. The first control run uses globally prescribed SST and is called the non-SOM control run. The second control run uses prescribed SSTs outside the Indian and western Pacific oceans, but is coupled to the SOM in the Indian and western Pacific oceans; this run is called as the SOM control run. The prescribed SST has only a climatological annual cycle, so that the effect of interannual variability of the Pacific SST field is not introduced. These two control runs will be compared with forcing experiments in which ENSO-related effects will be introduced through the structure of the added idealized heating. These forced experiments were run both without the SOM and with the SOM.

2.4 Forced runs

2.4.1 Method

To study the relationship of the seasonal mean Indian monsoon to the eastern Pacific SST (related to ENSO), anomalous tropical SST has been used as a boundary forcing in a full AGCM. This has been traditional approach, but it gives no control over the diabatic heating; differences in convective parameterization may lead to very different modeled diabatic heating anomalies for the same SST. Another traditional approach is to build an idealized GCM and force it with specified diabatic heating. This method has complete control over the diabatic heating, but at the expense of completing losing the feedback of the dynamics on the heating.

In this study, we choose a different method to force the GCM. We modify the existing GCM by adding a relatively small diabatic heating with idealized structure. The idea is to specify idealized changes in the diabatic heating, yet retain the full
feedback of the dynamics on the heating in the context of the full model. For example, the idealized rising motion induced by the added idealized heating (or idealized cooling) will influence the model’s circulation, which in turn can further change the moist and radiative heating fields produced by the model’s parameterizations. These second-order, or induced changes are referred to as the GCM effect.

In this method, small heating profiles are added to temperature tendencies in the AGCM. In the horizontal, three types of forcing are shown as Exp A (Idealized forcing), Exp B (Realistic Forcing), and Exp C (Observed forcing) in Table 2.1 and will be discussed in detail in the following section. The vertical structure of each heating profile will be also discussed later.

Twenty seasonal integrations (May to August) are performed using the twenty initial conditions from 1 May from each year of the control run. The simulations are run through August, and the twenty-run ensemble average of the May-August seasonal means is compared to the corresponding ensemble seasonal mean from the control simulation. In the analysis of the following chapters, the response to each forcing is calculated as the ensemble time mean change (forced experiments minus control). This ensemble procedure is expected to reduce the effect of mid-latitude disturbances in particular, and internal variability in general, and we can expect the ensemble mean model response to be dominated by forced waves from tropical heating or cooling. There is one caveat to this interpretation of the model results however: while the idealized additional heating is the same in each of the 20 integrations, the changes in the AGCM’s own diabatic heating induced by the additional forcing (called the GCM effect, as explained below), will not necessarily be the same in each of the 20 integrations. The role of internal variability is taken into account in a simple way through the use of the t-statistic to assess the difference in all ensemble seasonal means shown.
Table 2.1: The list of experiments according to forcing types

<table>
<thead>
<tr>
<th>Experiments</th>
<th>Forcing shape</th>
<th>Forcing location</th>
<th>Forcing magnitude</th>
<th>Ocean type</th>
</tr>
</thead>
<tbody>
<tr>
<td>Non-SOM control run</td>
<td>None</td>
<td>None</td>
<td>none</td>
<td>Prescribed SST (non-SOM)</td>
</tr>
<tr>
<td>SOM control run</td>
<td>None</td>
<td>None</td>
<td>None</td>
<td>SOM over the Indian Ocean and western Pacific</td>
</tr>
<tr>
<td></td>
<td></td>
<td></td>
<td></td>
<td>prescribed SST in other basins (SOM)</td>
</tr>
<tr>
<td>Exp A</td>
<td>Circular</td>
<td>• Longitude:</td>
<td>±0.5K/day $^2$</td>
<td>non-SOM</td>
</tr>
<tr>
<td>Idealized Forcing</td>
<td>shape</td>
<td>11 locations</td>
<td>±1K/day</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>60°E ~ 100°W</td>
<td>±2K/day</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>with 20° interval</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Latitude: equator</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Exp A-1</td>
<td>Circular</td>
<td>same 11 forcings</td>
<td>±1K/day</td>
<td>SOM</td>
</tr>
<tr>
<td>with SOM</td>
<td>shape</td>
<td></td>
<td>±2K/day $^3$</td>
<td></td>
</tr>
<tr>
<td>Exp A-2</td>
<td>Circular</td>
<td>• Longitude:</td>
<td>±1K/day</td>
<td>non-SOM</td>
</tr>
<tr>
<td>different latitudes</td>
<td>shape</td>
<td>same 11 locations</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Latitude: 7°N and 7°S</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Exp B</td>
<td>Elliptical</td>
<td>• Western Pacific(WP)</td>
<td>−0.5K/day</td>
<td>non-SOM</td>
</tr>
<tr>
<td>Realistic Forcing</td>
<td>shape</td>
<td>• Central/ eastern(CP)</td>
<td>+1K/day</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>Pacific</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>• Combined forcing</td>
<td></td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>of WP and CP(both)</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Exp B-1</td>
<td>Elliptical</td>
<td>• WP</td>
<td>−0.5K/day</td>
<td>SOM</td>
</tr>
<tr>
<td>with SOM</td>
<td>shape</td>
<td>• CP</td>
<td>+1K/day</td>
<td></td>
</tr>
<tr>
<td></td>
<td></td>
<td>• both</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Exp C</td>
<td>Observed</td>
<td>1982, 1997 summer</td>
<td>Observed</td>
<td>SOM</td>
</tr>
</tbody>
</table>

$^*$The spatial patterns of each forcing will be described in the following sections. The pattern of the idealized forcing (Exp A) is in section 4, the pattern of the realistic forcing (Exp B) is in section 5, the observed forcing (Exp C) is in section 6.

$^1$The domain of SOM is 60°E ~ 150°E, 30°S ~ 20°N

$^2$±0.5K/day represents heating of +0.5K/day and cooling −0.5K/day, each inserted separately into the model for heating and cooling experiments

$^3$only for 120°W, 100°W
The procedure used here is analogous to that used by Barsugli and Sardeshmukh (2002), who perturbed SST directly. The initial conditions from non-SOM control and SOM control will be used for the experiments with non-SOM and SOM, respectively.

### 2.4.2 Vertical structure of forcing

The spatial patterns of forcing will be different according to the types of the experiments in Table 2.1. But the vertical structure of the forcing is the same for all experiments. We, thus, consider the forced responses due only to different horizontal patterns. First, two types of idealized vertical structure are approximated, and then they are compared with the observed vertical structure to decide which one gives a better approximation to the observed structure. The vertical structure of the perturbation is generated as follows. The hybrid vertical coordinates are specified as input, and the corresponding mid-layer pressures are calculated from these assuming a fixed surface pressure of $p_0 = 1000\text{hPa}$. We use a standard set of $K=26$ levels, with mid-layer pressure values $h_k$ as given in the Table 2.2. Then a simple vertical profile of heating is

$$ F_k = \sin\left( \frac{\pi h_k}{p_0} \right) $$

This profile becomes very small at the bottom and top of the model. We then reflect the profile about its value at the top level, and define an extended heating over $2K$ levels as:

$$ F_k = \sin\left( \frac{\pi h_k}{p_0} \right) \quad -K \leq k \leq K - 1 $$

$$ F_k = -F_{1-k} \quad 0 \leq k \leq K - 1 $$
Table 2.2: Standard values \((h_k)\) of mid-level pressure (in Pa) for hybrid coordinate system for CAM3

<table>
<thead>
<tr>
<th>Level K</th>
<th>Pressure</th>
<th>Level K</th>
<th>Pressure</th>
</tr>
</thead>
<tbody>
<tr>
<td>1</td>
<td>354.5</td>
<td>14</td>
<td>22651.3</td>
</tr>
<tr>
<td>2</td>
<td>739.9</td>
<td>15</td>
<td>26648.1</td>
</tr>
<tr>
<td>3</td>
<td>1396.7</td>
<td>16</td>
<td>31350.1</td>
</tr>
<tr>
<td>4</td>
<td>2394.5</td>
<td>17</td>
<td>36881.8</td>
</tr>
<tr>
<td>5</td>
<td>3723.0</td>
<td>18</td>
<td>43389.5</td>
</tr>
<tr>
<td>6</td>
<td>5311.5</td>
<td>19</td>
<td>51045.5</td>
</tr>
<tr>
<td>7</td>
<td>7005.9</td>
<td>20</td>
<td>60052.4</td>
</tr>
<tr>
<td>8</td>
<td>8543.9</td>
<td>21</td>
<td>69679.6</td>
</tr>
<tr>
<td>9</td>
<td>10051.5</td>
<td>22</td>
<td>78770.2</td>
</tr>
<tr>
<td>10</td>
<td>11825.0</td>
<td>23</td>
<td>86716.1</td>
</tr>
<tr>
<td>11</td>
<td>13911.5</td>
<td>24</td>
<td>92964.9</td>
</tr>
<tr>
<td>12</td>
<td>16366.2</td>
<td>25</td>
<td>97055.5</td>
</tr>
<tr>
<td>13</td>
<td>19254.0</td>
<td>26</td>
<td>99255.6</td>
</tr>
</tbody>
</table>

This new (extended) function nearly vanishes at the end points, and is defined for the domain \(-K \leq k \leq K - 1\). It has zero vertical mean, and is well approximated by its first few harmonics. The approximation is:

\[
\hat{F}_k \approx \sum_{m=1}^{M} \left( A_m \cos\left(2\pi \frac{mk}{K}\right) + B_m \sin\left(2\pi \frac{mk}{K}\right) \right)
\]  

(2.2)

where \(M\) is a small number. We only need the values of \(\hat{F}_k\) for the first \(K\) levels, which however must be numerically given as \(-K \leq k \leq K - 1\). Equation 2.2 is implemented in the subroutine (dynamics and physics coupling module), in which the full temperature tendency is available in grid point configuration.

The formulae for the coefficients \(A_m\) and \(B_m\) are standard:

\[
A_m = \frac{2}{2K} \sum_{j=-K}^{K} F_j \cos\left(2\pi \frac{mj}{2K}\right)
\]  

(2.3)

22
\[ B_m = \frac{2}{2K} \sum_{j=-K}^{K} F_j \sin\left(2\pi \frac{m_j}{2K}\right) \]  

and the coefficients are given in the Table 2.3 for two cases \((m = 2, m = 3)\). For the heating given in equation 2.1 only \(m = 2\) terms are a good approximation, and the resulting profile is given by the blue dots in figure 2.4.

![Figure 1: Heating profiles derived from equations 1 and 5 by harmonic approximation. The ordinate is given by 1 \(-\frac{h_k}{p_0}\). Values are given only at the 26 model levels.](image)

In order to achieve a somewhat shallower profile, we modify the formula in equation 2.1:

\[ F_k = \sin\left(\frac{h_k}{p_0}\right) - 0.3\sin\left(2\frac{h_k}{p_0}\right) - 0.1\sin\left(4\frac{h_k}{p_0}\right) \]  

(2.5)
Table 2.3: Coefficients of $A_m$ and $B_m$

<table>
<thead>
<tr>
<th></th>
<th>m=1</th>
<th>m=2</th>
<th>m=3</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A_m$</td>
<td>-0.0418</td>
<td>-0.0392</td>
<td></td>
</tr>
<tr>
<td>$B_m$</td>
<td>-0.6917</td>
<td>-0.3230</td>
<td></td>
</tr>
</tbody>
</table>

$M = 2$, harmonic coefficients corresponds to heating of equation 2.2

$M = 3$, harmonic coefficients corresponds to heating of equation 2.5

<table>
<thead>
<tr>
<th></th>
<th>m=1</th>
<th>m=2</th>
<th>m=3</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A_m$</td>
<td>-0.0307</td>
<td>-0.0597</td>
<td>-0.0346</td>
</tr>
<tr>
<td>$B_m$</td>
<td>-0.5068</td>
<td>-0.4920</td>
<td>-0.1998</td>
</tr>
</tbody>
</table>

An extended heating is defined as above, and an approximation in terms of harmonic coefficients works well with only $m = 3$ terms. Figure 2.4 shows this heating profile by the red dots.

To compare the two profiles (figure 2.4) with observed diabatic heating, anomalous diabatic heating diagnosed from ERA-40 data by Nigam and Chan (2009) is used for El Niño and La Niña event in figure 2.5. Two types of the vertical structures are shown: one certain point (solid lines with closed circles) and spatially averaged values (dashed lines). In figure 2.4, the shallow vertical profile has the maximum at 600mb and the deep profile has the maximum at about 450mb. The diagnosed ERA-40 diabatic heating for both El Niño and La Niña shows the maximum below 500mb, although each event has a different maximum level. For this reason, in all the following experiments, the shallower profile of figure 2.4 is used for the vertical structure of each idealized forcing.

Note that since we have assumed the surface pressure equal to $p_0$. The ordinate in figure 2.5 should really be the model coordinate; the correspondence with pressure is only valid if the true surface pressure is close to $p_0$, typically over the ocean.
Figure 2.5: The vertical structure of ERA40 Diabatic Heating anomalies rate (K/s) over the western Pacific. a) for El Niño events (1982, 1987, 1991, 1994, and 1997). The solid lines with closed circles are at 132°E and 2.5°S. The dashed lines are averaged from 120°E to 140°E and 5°S to 2°N. b) for La Niña events (1984, 1988, 1998, and 1999). The solid lines with closed circles are at 137°E and 2.5°S for 1984 and 1988 and for 1998 and 1999 at 100°E and 2.5°S. For 1984 and 1988 the dashed lines are averaged from 130°E to 140°E and 5°S to 2°N and for 1998 and 1999 they are averaged from 95°E to 105°E and 5°S to 2°N.
2.5 Forcing types

2.5.1 Idealized experiments

The added idealized heating (Exp A in table 2.1) with the vertical profile in figure 2.4 is added to the temperature tendency equation. In the horizontal, the heating assumes an isotropic Gaussian form with e-folding distances corresponding to 20 degrees and the center of Gaussian distribution located near the equator (1.7°S) and at a particular longitude. For the various experiments, the longitudes of the center of the additional idealized heating vary from 60°E to 100°W with a 20-degree interval. Idealized additional cooling experiments will be run with the same manner except for the sign reversal. For the non-SOM experiments, magnitudes corresponding to ±0.5, ±1, ±2K per day (at the level of peak forcing) is used with the prescribed SST (Exp A in table 2.1). For the SOM experiments ±1K per day peak forcing is used (Exp A-1 in table 2.1) and ±2K per day peak forcing is used at 120°W and 140°W. In addition, the heating and cooling at different latitudinal locations such as 7°S, the equator, and 7°N (Exp A-2 in table 2.1) are used. Each experiment has the same 11 longitudinal locations.

The experiments with the idealized cooling near the western Pacific will have only a downward branch of anomalous Walker circulation: this models one piece of the anomalous Walker circulation during El Niño. (Similarly, by using the idealized heating near the Western Pacific, we introduce only the anomalous upward branch of the Walker circulation that occurs during La Nina.) The purpose of this idealized cooling is to understand the direct effect of the western Pacific Walker circulation anomalies on the Indian summer monsoon, without completely describing the influence of those in the eastern Pacific. In addition, the forcing of the different longitudinal locations is used to test the sensitivity of the additional tropical heating/cooling in the GCM.
Since the model used in this study is a full AGCM, the changes of temperature
tendencies will induce strengthening or weakening of convection by moist processes.
In a linear moist model, an inserted heating and cooling induces convectively coupled
atmospheric response (Watanabe and Jin, 2003). By inserting the idealized heating
into the full AGCM, there is a feedback of dynamics on heating and this additional
heating by GCM can modify the forced atmospheric response by the idealized heating.
This additional heating/cooling is considered as the GCM effect. The different longitudi-
dinal locations of the heating/cooling are expected to lead to different GCM effects.
To test the sensitivity of the GCM effect, different magnitudes of the heating/cooling
are varied so that the maximum of each forcing is ±0.5, ±1, ±2K/day.

2.5.2 Realistic experiments (combined effect of heating and cooling)

In a real ENSO event, anomalous rising motion over the equatorial central/eastern
Pacific is accompanied by sinking motion over the western Pacific. The realistic exper-
iments in table 2.1 (Exp B) will have simultaneous heating over the central/eastern
Pacific and cooling over the western Pacific to understand the combined effect of
heating and cooling. Moreover, each heating and cooling is also inserted separately
in the model and the linear combination of these two experiments is studied. Cir-
cular shapes of forcing are used for simplicity in Exp A. On the other hand, since
warming over the central/eastern Pacific is longitudinally wide, elliptical shapes of
the forcing will be used for Exp B. The realistic experiments are also performed with
non-SOM and SOM, separately. The two types of ocean (non-SOM and SOM) have
climatological SST where the ocean is prescribed. If observed SST were to be used
here along with the added heating, such a procedure would “double count” the effect
of the anomalous heating/SST.

2.5.3 Observed forcing experiments

After the general response of the idealized forcing is studied, observed anomalous heating is selected from particular El Niño years and inserted in the SOM ocean in Exp C of table 2.1. ERA-40 diagnosed diabatic heating (Chan and Nigam, 2009) is used to calculate the observed anomalous forcing; the summers of 1982 and 1997 are selected for El Niño developing cases. The spatial differences of the two observed forcing are expected to simulate different responses in AGCM. The previous idealized experiments will be a guide to understand the response to the observed forcing.
Chapter 3: Two control runs

3.1 Change of state

Two control simulations are carried out before the idealized forcing experiments. First, climatological SST is prescribed to force the AGCM; this is called the non-SOM control run in this study. The second control run is coupled with a slab ocean model over the Indian Ocean and western Pacific (60°E to 150°E and 30°S to 20°N) and the other basins are prescribed with climatological SST. This control run will be called the SOM control run. Since the forced experiments will be implemented during northern summer (May to Aug), we will compare the mean state from the two control runs during summer and the change of state induced by air-sea coupling.

In the slab ocean model, the net atmosphere-to-ocean heat flux at the surface mainly determines the ocean mixed layer temperature. The net downward heat flux at the surface in the absence of sea ice consists of the net solar flux absorbed by the ocean mixed layer, the net downward long-wave energy and sensible heat fluxes at the ocean surface, and minus the latent heat flux from the ocean to the atmosphere. In this section, the interaction between SST and the heat fluxes changes will be studied to understand the change of the state induced by the regional air-sea coupling.

The mean precipitation during the northern summer is shown for the two control runs in figure 3.1. The Indian region has large precipitation and both control runs simulate large precipitation as the Indian monsoon rainband. However, the non-SOM control run (figure 3.1b) has excessively large precipitation over the equatorial Indian Ocean compared to the observed precipitation in figure 3.1d. In Fu and Wang (2002),
excessive rainfall over the equatorial Indian Ocean is also mentioned as a common problem of AGCMs. On the other hand, in the SOM control run (figure 3.1a) the magnitude of the precipitation is similar to the observed precipitation. In Fu and Wang (2002) and Wu and Kirtman (2005), the exaggerated precipitation over the equatorial Indian Ocean is reduced by applying the regional air-sea coupling.

Wu and Kirtman (2005) suggested a negative feedback between the atmosphere and ocean in the regions of large mean rainfall. Positive SST anomalies can enhance rainfall through surface evaporation. The enhanced rainfall can intensify the oceanic mixing and entrainment and reduce the incoming short wave radiation. These effects could in turn lower the SST, thus providing the negative feedback (Wu and Kirtman, 2005). While in the air-sea interaction the negative feedback will bring cool SST and will weaken precipitation, in the prescribed SST experiment, the atmosphere continuously receives heat flux from the ocean, thus resulting in excessive precipitation. In the SOM control run (figure 3.1c) and previous studies (Fu and Wang, 2002; Wu and Kirtman, 2005), a reduced and more realistic precipitation is simulated in the regions of convection by the negative feedback of the air-sea interaction.

However, in this SOM control run there is still a problem of exaggerated precipitation over the western Indian Ocean. The SOM control run shows precipitation over the western Indian Ocean near India. In the observations, the precipitation in this area does not expand westward as much as the SOM control runs. One reason for this error might be the choice of the coupling region. The coupling region does not cover the western Indian Ocean entirely.

In figure 3.2, the summer mean of SST is shown for the SOM, non-SOM control runs and differences between them. Negative anomalies (differences) of SST appear over the northern Indian Ocean where there is large precipitation. As mentioned in Wu and Kirtman (2005)’s negative feedback, the reduced incoming solar radiation
Figure 3.1: The mean precipitation during summer (May to August). a) is for the control run with a slab ocean model over the Indian Ocean and western Pacific. b) is for the control run with climatological SST. c) is differences between a) and b). Shading is at 5% significance. d) is observed precipitation.
by convection cools down the ocean, and further the cooled SST reduces precipitation. On the other hand, in the southern Indian Ocean positive anomalies of SST appear. The southern Indian Ocean does not have strong convection, so here different feedbacks between the ocean and atmosphere might be operation.

In addition, the state of wind at 850hPa is shown in figure 3.3. In the experiment chapters, the forced atmosphere response and its relationship with heat flux is considered extensively, and information about the climatological wind will be helpful. The mean summer flow in the Indian region is easterly over the southern Indian Ocean towards South Africa, where the flow turns northward and becomes westerly (Somalia jet) towards India in figure 3.3d. These main features also appear in the two control runs (figure 3.3a and b). However, in the model the westerlies to west of India and the easterlies over the southern Indian Ocean are too strong, while the westerly flow over the Bay of Bengal is too weak. By applying SOM regionally, the excessively strong easterly flow over the southern Indian Ocean becomes weaker (compare figure 3.3a to figure 3.3b). This is shown as positive westerly anomalies in figure 3.3c, indicating weakened easterlies.

### 3.2 Interaction between SST and heat flux

In this slab ocean model, the surface heat flux determines SST, with the surface total heat flux consisting of radiative heat flux, latent heat and sensible heat flux. In order to understand the changes of SST and precipitation in terms of the heat flux, the lead-lag correlation of an SST index with the fields of SST, precipitation, latent heat flux, radiative heat flux, and surface total heat flux are examined for the SOM and non-SOM control runs. Since anomalous SST in the SOM reverses sign between the northern and southern Indian Ocean, the variables are averaged
Figure 3.2: The mean SST during MJJA a) for the second control run with a slab ocean model over the Indian Ocean and western Pacific, b) for the control run with climatological SST, c) differences between a) and b) and shading is at 5% significance
Figure 3.3: The mean wind during summer (May to August). Shading indicates the zonal wind. a) is for the control run with a slab ocean model over the Indian Ocean and western Pacific. b) is for the control run with climatological SST. c) is differences between a) and b) and the zonal wind at 5% significance are selected and shaded in c). d) is observed wind (ERA-40). Mean zonal wind is shaded in a), b), and d). The box in c) is the SOM domain.
for two regions: 1) northern Indian Ocean (60°E ∼ 100°E, the equator ∼ 20°N), and 2) southern Indian Ocean (60°E ∼ 100°E, 30°S ∼ 10°S). Correlations are computed using these area averages.

### 3.2.1 Northern Indian Ocean

The lead-lag correlations of SST with the heat fluxes, precipitation, and SST over the northern Indian Ocean are shown for the SOM and non-SOM control runs in figure 3.4 (*total surface and radiation heat fluxes are atmosphere-to-ocean, that is positive if downward, and latent and sensible heat fluxes are ocean-to-atmosphere, that is positive if upward*). To calculate the correlation coefficients, the monthly climatological cycle of the time-series are removed. The correlations of one or two-month lead and lag show significantly large values for most variables in the SOM control run (figure 3.4a). This implies that SST both influences the heat fluxes and precipitation and is also influenced by them. But in the non-SOM control there is no correlation between SST and other variables (heat fluxes and precipitation) in figure 3.4b. It is surprising that without climatological variation there is no correlation between precipitation and SST. In the non-SOM control run, SST does not respond to the change of the environment.

The significant lead and lag correlation values imply an interaction between SST and surface heat flux in figure 3.4a. The purple line in figure 3.4a shows significant lead and lag correlation between SST and the surface total heat flux. When the surface heat flux leads SST by one month, the correlation is the largest and its coefficient is about 0.6. This lagging SST correlation means that the increased (decreased) surface latent heat flux raises (lowers) SST. This relation is consistent with the formulation of the temperature of the mixed layer in the slab ocean model of CAM. This relation cannot be seen in the non-SOM control runs when the climatology is removed.
Figure 3.4: Correlation of northern Indian Ocean SST index with surface heat flux (purple), SST (black), precipitation (blue), latent heat flux (red), sensible heat flux (orange) and radiative heat flux (green) (see text for details). a) is for the control run with the regional SOM. b) is for the control run with climatological SST. Correlation coefficient 0.126 is at 5% significance.
In figure 3.5a, to see the evolution of SST and surface heat flux, the variables are averaged over the northern Indian Ocean and monthly means calculated using 20 years for the SOM (open circle) and the non-SOM control run (closed square). The time-series have mean quantities varying monthly (climatological annual cycle is not removed), while the correlations in figure 3.4 are based on deviations about the climatological monthly cycle. The SST in the SOM and non-SOM control runs increase from January to May and decrease from May to August. During May, the SST in the SOM control run is larger than in the non-SOM. During July and August, SST in the SOM control run is smaller than in the non-SOM. In the evolution of SST, the differences between the two control runs show anomalously cold SST in the SOM control run and this is consistent with the negative anomalies of SST over the northern Indian Ocean in figure 3.2c.

The time-series of the surface total heat flux for both control runs in figure 3.5a have a maximum during April and decrease during summer, and they lead SST by one month. The minimum of the heat flux occurs in June (SOM control run) and July (non-SOM control run). Although the minimum of the surface heat flux is similar (about 450Wm$^{-2}$) for both control runs, the SST in the non-SOM control run does not cool down as much as the SOM control run. In the non-SOM control run, SST is prescribed so it does not respond to the change of the surface heat flux.

The surface heat flux consists of radiative heat flux, and the negative of latent heat flux and sensible heat flux. Each heat flux is considered separately to see the interaction between SST and the heat fluxes. In figure 3.5b, the time series of SST and the radiative heat flux are averaged over the northern Indian Ocean for the two control runs. The radiative heat fluxes have positive correlation with SST in one month-lagging SST in figure 3.4 and no significant correlation with leading SST. In the climatological time-series (figure 3.5b), this radiative heat flux for both control
runs leads SST. In terms of the magnitudes, the radiative heat flux contributes most
to the surface heat flux compared to the latent and sensible heat flux. However, the
radiative heat flux contributes only about 20 ∼ 30% to the differences of the total
surface heat flux between the SOM and non-SOM control runs.

In figure 3.4a, the latent heat flux anomalies are correlated with one-month-leading
and lagging SST implying that both SST and the latent heat flux influence each
other. The positive correlation with one-month leading SST means that increased
(decreased) SST is related to the increment (decrement) of the latent heat flux from
the ocean to the atmosphere in one month. The negative correlation with one-month
lagging SST implies that the increased (decreased) latent heat flux from the ocean
is related to cooling (warming) SST. The SST leading relation can be also seen in
climatological time-series of SST and the latent heat flux in figure 3.5c. In the SOM
control run, the maximum of SST occurs in May and then the maximum of the latent
heat flux occurs in the next month (June). The pattern of two time-series for both
runs shows that SST leads the latent heat flux from the ocean to the atmosphere.
On the other hand, when we consider climatological latent heat flux leading SST,
we can see that SST of the non-SOM control run does not respond to the change of
the latent heat flux. For example, during July the latent heat flux of the non-SOM
control run is larger than the SOM control run yet the next month (August) SST
cools down only a small amount compared to the previous month and less than the
SOM SST decreases

When SST increases, the sensible heat flux also increases from the ocean to the
air. On the other hand, after removing the climatology the sensible heat flux does
not have significant correlation with SST in figure 3.4a (orange). The contribution of
the sensible heat flux in figure 3.5d is also negligible compared to the radiative heat
flux and latent heat flux in figures 3.5b and c. So in the northern Indian Ocean the
Figure 3.5: Climatological time-series of the monthly mean for 20 years averaged from 60°E to 100°E and from the equator to 20°N (the northern Indian Ocean). The open circle marker is for the control run with the regional SOM and the closed square is for the control run with climatological SST. a) SST(black) and total surface heat flux(purple). b) SST(black) and radiative heat flux(green). c) SST(black) and latent heat flux(red). d) SST(black), sensible heat flux(orange). (SST scales always on left; other quantity plotted in each box has scale on the right and units of SST and each heat flux are K and Wm$^{-2}$, respectively)
Figure 3.6: Correlation of precipitation with surface heat flux (purple), SST (black), precipitation (blue), latent heat flux (red), sensible heat flux (orange) and radiative heat flux (green) in the region averaged from 60°E to 100°E and from the equator to 20°N (northern Indian Ocean). a) is for the control run with the regional SOM. b) is for the control run with climatological SST. Correlation coefficient 0.126 is at 5% significance.
radiative and latent heat fluxes play an important role in interacting with SST. The radiative heat flux leads and mainly influences SST and SST leads latent heat flux.

The radiative heating is negatively correlated with precipitation in simultaneous correlation (figure 3.6a). In the strong convection region, clouds reflect solar radiation so that incoming heat flux to the surface decreases. In the northern Indian Ocean, there is a rain-band in figure 3.1a and b and in turn reduced radiative heat flux by the clouds cools SST in the SOM control run. But SST in the non-SOM does not respond to this and there is no significant correlation in figure 3.6b except simultaneous correlations. Due to the negative feedback of the air-sea interaction in the strong convection region, the differences in SST between the SOM and non-SOM control runs are negative, as seen in figure 3.2c.

### 3.2.2 Southern Indian Ocean

In figure 3.7, SST and the heat fluxes are averaged over the southern Indian Ocean (60°E ∼ 100°E, 30°S ∼ 10°S). The time series in figure 3.7 show the climatological monthly cycle in the same manner as in figure 3.5. The time-series of the total surface heat flux leads the monthly evolution of SST with a lag of several months ahead in figure 3.7a. For both control runs the radiative heat flux varies in almost the same manner throughout the year due to small change of precipitation in this non-convection region. Yet, SST in the SOM control run is higher than the non-SOM control run. This is consistent with the positive anomalies of SST over the southern Indian Ocean in figure 3.2c. The SST is prescribed in the non-SOM control run so SST does not respond to the radiative heating. In figure 3.7c and d, SST leads the latent and sensible heat fluxes. So in the southern Indian Ocean the lead-lag relationship between SST and heat flux is similar to the northern Indian Ocean. On the other hand, in the southern Indian Ocean, SST and heat flux vary more slowly.
than in the northern Indian Ocean. SST and the latent (sensible) heat flux respond slowly to the radiative heat flux and SST, respectively.

Although the relationship between SST and heat flux is similar between the northern and southern Indian Ocean, the anomalies of SST between the two control runs are different in the two regions. While in the northern Indian Ocean (convection region) SST in the non-SOM control run appears not to cool down the decreasing radiative heat flux, in the southern Indian Ocean (non-convection region) SST in the SOM warms more than the non-SOM control run. So there are negative anomalies of SST in the northern Indian Ocean and positive anomalies of SST in the southern Indian Ocean.

When the slab ocean model is applied in the Indian region and western Pacific, a change of the state in terms of SST and precipitation is seen. In the SOM control run, SST responds to the surface heat flux and the precipitation also responses to the changed SST. When the atmospheric response by tropical heating is studied, Wang et al. (2003) suggested the importance of air-sea interaction to understand a sustained response over the Indian Ocean during the development of El Niño.

In the following experiments, the idealized forcing is inserted in the model for the two control runs to study how the prescribed SST (the non-SOM control run) or the air-sea interaction (the SOM control run) interplays with the forced responses.
Figure 3.7: Climatological time-series of the monthly mean for 20 years averaged from 60°E to 100°E and from 30°S to 10°S (southern Indian Ocean). The open circle is for the control run with the regional SOM and the closed square is for the control run with climatological SST. a) SST (black) and surface heat flux (purple). b) SST (black) and radiative heat flux (green). c) SST (black) and latent heat flux (red). d) SST (black) and sensible heat flux (orange). (SST scales always on left; other quantity plotted in each box has scale on the right and units of SST and each heat flux are K and Wm$^{-2}$, respectively)
Chapter 4: Idealized experiments

4.1 GCM effect

The composites of diabatic heating for El Niño (1982, 1987, 1991, 1994, and 1997) and La Niña (1984, 1988, 1998, and 1999) are shown in figure 4.1. The diabatic heating is calculated from the ERA-40 data set by Chan and Nigam (2009). The diabatic heating is vertically integrated in figure 4.1a (El Niño) and figure 4.1c (La Niña) and the vertical distributional diabatic heating at 3°S is shown in figure 4.1b (El Niño) and figure 4.1d (La Niña).

The horizontal distribution of idealized additional heating at different longitudinal locations shown in figure 4.2 can be compared with the composite maps of the diagnosed diabatic heating from the ERA-40 data in figure 4.1a and c. The same idealized forcing with an opposite sign is also used for the cooling experiments. The idealized heating centered at 180°, 160°W, 140°W, 120°W, and 100°W can be compared with the anomalous diabatic heating over the central/eastern Pacific in the composite of El Niño (figure 4.1a). The negative anomalies of the diabatic heating over the western Pacific during El Niño in figure 4.1a can be compared with the idealized cooling centered at 100°E and 120°E in the model. During La Niña events the anomalous diabatic heating in figure 4.1c also can be compared with the idealized cooling (heating) over the central/eastern Pacific (western Pacific).

In this way, each component (western, central, and eastern Pacific) of the diabatic heating is studied for its own influence on the Indian region; this will help to understand the impact of the total diabatic heating. Although the general characteristic
Figure 4.1: Composite of diagnosed diabatic heating anomalies from ERA-40 data during developing (MJJA) El Niño (1982, 1987, 1991, 1994, and 1997) and La Niña (1988, 1998, and 1999) from 1979 to 2001. a) diabatic heating integrated vertically from the surface to 100 hPa during El Niño. b) vertical diabatic heating is shown at 3°S. c) diabatic heating integrated vertically from the surface to 100 hPa during La Niña. d) vertical diabatic heating is at 3°S. Shading is at 5% significance.
Figure 4.2: Positive idealized diabatic forcing is integrated vertically (Wm$^{-2}$). The maximum of the forcing is +2K/day. Forcing is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. Black dots show the center of the forcing.
of ENSO is an anomalously zonal circulation (Walker circulation), each ENSO event has a different spatial pattern in details. This method lets us test the atmospheric sensitivity to different longitudinal diabatic heating location, and may help to understand how the atmospheric circulation responds differently to each discrete ENSO event.

\section*{4.1.1 Diabatic heating responses}

Idealized diabatic heating is inserted in the AGCM to understand how tropical heating changes atmospheric circulation. The idealized changes are specified in the diabatic heating and the feedbacks between the dynamics and the heating by GCM are retained in the context of the full model. For example, the rising motion induced by the added heating can further change the moist and radiative heating fields produced by the model’s parameterizations. Watanabe and Jin (2003) and Annamali (2010) explained that equatorial forced atmospheric circulation interacts with tropical convection and induces both additional heating and changes in atmospheric circulation related to the additional heating. In this study, this additional diabatic heating in the coupled dynamical convective response is referred to as the GCM effect.

The vertical structure of the idealized added heating has been discussed in Section 2. The maximum magnitude of this added heating within the vertical column, referred to as simply the magnitude of the heating $M_Q$, is taken to be in the range $+0.5\text{K/day}$ to $+2\text{K/day}$ in separate experiments to examine the GCM effect. The idealized heating inserted in the AGCM is shown at eleven geographical locations in figure 4.2 for the case of the heating magnitude of $+2\text{K/day}$. In figure 4.2, the heating is shown vertically integrated from the surface to 100hPa, in Wm$^{-2}$.

To see the feedback of the moist and radiative processes in the AGCM that are induced by the added idealized heating, we show the sum of the changes (experiment
minus control) for the moist, long-wave and short-wave radiative heating rates in the experiments in figure 4.3 through 4.5 (corresponding to values of $M_Q$ of 0.5, 1.0 and 1.5 K/day, respectively). These heating changes do not include the idealized inserted heating, but represent the GCM response (GCM effect). Here the heating response is also vertically integrated from the surface pressure to 100hPa.

For +1K/day experiments (figure 4.3) there are positive anomalies of diabatic heating in the forcing region (figure 4.2) and these anomalies are considered as a local response. When the heating is inserted over warm SST to the west of 180° (figure 4.3a~f), a relatively strong response is seen in the Indian region, the equatorial Indian Ocean, and the western Pacific. The heating over the cold SST (the east of 180°: figure 4.3h~k) shows weak responses in the forcing regions and regions remote from the forcing.

The local response has a range of spatial patterns and magnitudes depending on the location of the additional forcing. In the case of 60°E (figure 4.3a), the positive anomalies are centered to the north and south of the forcing center (stronger to the south). The forcing over the warm SST (60°E, 80°E, 100°E, and 120°E) has the same pattern of a meridionally extended response (figure 4.3a to d), while the forcing over the cold SST (figure 4.3i~k; forcing at 140°W, 120°W, and 100°W) leads to anomalies only to the north of the forcing. In the cases of 140°E (e), 160°E (f), and 180° (g), regions to the north of the forcing show a wider and stronger diabatic heating response than to the south of the forcing. In the cases of idealized forcing added at 80°E (b), 100°E (c), and 120°E (d), the response is stronger and wider to the north of the forcing and this is connected to the response over the Indian region. None of the responses have the maximum anomalies in the center of the forcing location. Thus, the diabatic heating does not respond with the same horizontal structure as the inserting forcing (figure 4.2). This will be discussed further.
Figure 4.3: Diabatic heating response (GCM effect; anomalies from the non-SOM control run) to inserted positive diabatic forcing. The response of the diabatic heating is integrated vertically (Wm\(^{-2}\)). The maximum of the forcing is +0.5K/day. Forcing is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. Statistically significant values at 5 % are shaded. The closed black square is the center of the forcing.
Figure 4.4: The same as figure 4.3 with maximum of the forcing of is +1K/day
Figure 4.5: The same as figure 4.3 with maximum of the forcing of +2K/day
For the forcing at 60°E (figure 4.3a), negative heating anomalies appear to the north of the heating. For the range of forcings from 80°E (b) to 180°E (g), relatively strong positive anomalies of the diabatic heating are shown to the east and west side of India, consistently regardless of the location of the forcing. This is considered as one remote effect in this AGCM. Another remote effect consists of negative anomalies of diabatic heating over the equatorial Indian Ocean in the case of forcing at 80°E to 180° (figure 4.3b to g). There are other remote responses over the northeastern Pacific. However, the remote response near India will be the focus in this study.

The magnitude of the heating is increased (to +1K/day and +2K/day) to examine GCM effect further. The local and remote responses described for the +0.5K/day case also appear in the experiments for +1K/day (figure 4.3) and +2K/day (figure 4.4). The northward enhanced responses, the negative response over the equatorial Indian Ocean, and the positive anomalies in the vicinity of India show increased magnitudes with the strengthened forcing. Over the eastern Pacific, remote negative anomalies to the forcing at the eleven locations also become large in the +1K/day and +2K/day cases. The larger magnitudes of the forcing are useful to see clearly the features of the remote and local response in the GCM effect.

In the experiments using +2K/day, the GCM effect over the Bay of Bengal and western North Pacific (figure 4.5d, e, and f) becomes stronger than the inserted idealized heating (figure 4.2), while in the case of +0.5K/day, the local response is very weak when the heating is inserted over the eastern Pacific. In the remaining analysis, we will focus on the +1K/day results to study further the atmospheric circulation associated with the remote and local responses of the diabatic heating. The response of the diabatic heating to the inserted heating is a part of the total forcing associated with an atmospheric response in circulation. On the other hand, for all three different magnitude experiments the idealized heating over the warm SST
brings a stronger diabatic heating response than the heating over the cold SST. This is useful to know when we use observed diabatic heating as forcing in the model in a later chapter.

### 4.1.2 Precipitation responses

The diabatic heating feedback shown earlier consists of heating rates by moist processes (convection and resolved condensation) and longwave and shortwave radiation. Here each heating rate is considered separately. The heating rate by moist processes in figure 4.6 has all the features of total diabatic heating response, and the magnitude is also comparable to the total response in figure 4.4. The response of longwave heating rate in figure 4.7 is in general about 20Wm$^{-2}$ in the heating region as the local response, while the remote response over the Indian region and equatorial Indian Ocean is in general about 20Wm$^{-2}$. This is a small portion of the total diabatic heating response. The positive response of the longwave heating at each forcing region implies less longwave cooling due to anomalously increased convection. The strengthened convection traps longwave energy under the clouds. Shortwave heating rate anomalies are too small to compare with other processes (not shown).

Since the moist processes are dominant in the diabatic heating response, the response of precipitation is also shown in figure 4.8. Note that the precipitation anomalies are related to the sum of the vertical velocity induced directly from the idealized inserted heating and the vertical velocity caused by the diabatic heating response. For the forcing in the range from 60$^\circ$E to 180$^\circ$, positive precipitation anomalies appear to the north and south of the forcing center. The remote response in the vicinity of India and negative anomalies in the equatorial Indian Ocean appear. For the forcing in the range from 160$^\circ$W to 100$^\circ$W, positive precipitation anomalies appear directly in the heating region and negative precipitation anomalies in the vicinity of the heating
Figure 4.6: Vertically integrated heating rate response (Wm$^{-2}$) by moisture processes in +1K/day forcing at the eleven different locations. Shading indicates 5% significance. Black dots show the center of the forcing.
Figure 4.7: Long wave radiation heating response (Wm$^{-2}$) in +1K/day at the eleven different locations
region. These characteristics of the anomalous precipitation at each forcing location are similar to those of the total diabatic heating (figure 4.4). As shown in figure 4.2, none of the locations of the inserted heating is directly over India. However, when the forcing is relatively near to India (forcing at 60°E to 180°), there are significant precipitation anomalies over the Indian region, both as local and remote responses.

The feedback of dynamics on heating in GCM (the GCM effect) might be important to explain the precipitation anomalies over India because the inserted heating is confined near the equator. The next section addresses how the diabatic heating response is related to the atmospheric circulation and if the responses are physically reasonable in the absence of air-sea interaction.

4.2 Local vs. remote response

4.2.1 Local response

The atmospheric circulation responses to the diabatic heating are compared for two cases of inserted heating magnitude of +1K/day at 140°E and at 120°E in figure 4.9 and 4.10. The center of the local response in the diabatic heating appears somewhat to the north of the center of the inserted forcing (which is 1.4°S); the anomalies are also spread to the north and south of the center of the forcing (figure 4.9a and figure 4.10a). In the precipitation (figure 4.9b and figure 4.10b), these features also appear in both cases. Lower level divergence fields in figure 4.9d and figure 4.10d (integrated from the surface to 800hPa) are consistent with the heating anomalies. On the other hand, the upper level divergence fields (integrated from 275hPa to 80hPa) show positive anomalies centered near the center of the forcing in figure 4.9c and figure 4.10c. The inserted heating in the tropics induces ascending motion and divergence at the upper level to compensate this ascending motion. The location of
Figure 4.8: Precipitation response (mm/day) to +1K/day forcing at eleven locations
divergence at the upper level is identical with the forcing region but the responses of the divergence fields, diabatic heating, and precipitation near the surface are slightly displaced from the forcing region. Thus, the northward shift in local response might be related to the boundary layer and/or lower tropospheric physical process.

The evolution of streamfunction at 850hPa (figure 4.11) and surface pressure with surface divergent wind (figure 4.12) for the forcing at 140°E is shown before baroclinic instability grows strongly. At day 1 (i.e. 1 day after the heating is turned on), the surface pressure (figure 4.12a) is centered at the forcing with surface convergence. At day 3, the four cells of the Gill type response in the stream-function response appear and strengthen with time (figure 4.11). The two cyclonic circulations to the west of the forcing induce lows in the center of the each circulation cell appearing in the surface pressure (figure 4.12c-f). With the development of the Gill type circulation the low in surface pressure spreads to the north and south of the forcing region, The surface convergence consistent with the surface pressure can induce moisture convergence and leads to the additional heating response. The interaction between the forced atmospheric response near the surface and moist processes causes the local feedback response (GCM effect).

As noted above, the response of the diabatic heating in the cases of forcing at 140°E, and 120°E is stronger to the north of the equator than to the south (figure 4.9a and figure 4.10a). This asymmetry also appears in the atmospheric response. In Wang et al. (2003), the Rossby response to tropical heating is asymmetric about the equator during northern summer. Easterly vertical wind shear is strong over the Indian Ocean during the northern summer and this induces the asymmetry of the response. In this study, the streamfunction (figure 4.9e and figure 4.10e) and vorticity (figure 4.9f and figure 4.10f) also show the strengthened response to the north of the equator. In turn, this asymmetry of the atmospheric circulation response might
Figure 4.9: The response to idealized heating inserted at 140°E. a) vertically integrated diabatic heating (W m\(^{-2}\)), b) precipitation (mm/day) c) divergence integrated from 275hPa to 80hPa (\(\times 10^{-3}\)kg/(m\(^2\)s)) d) divergence integrated from surface to 800hPa (\(\times 10^{-3}\)kg/(m\(^2\)s)). e) stream-function at 850hPa (\(\times 10^6\)s\(^{-1}\)) f) vorticity integrated from the surface to 800 hPa (\(\times 10^{-3}\)kg/(m\(^2\)s)). Statistically significant values at 5% are shaded. Black dots show the center of the forcing.
Figure 4.10: The same as figure 4.9 except the response to 120°E heating.
Figure 4.11: The response of +1K/day heating at 140°E. Stream-function anomalies ($\times 10^6$ s$^{-1}$) at 850hPa a) at 1 day, b) at 3 day, c) at 6 day, d) at 9 day, e) at 12 day, and f) at 15 day. Black dots show the center of the forcing.
Figure 4.12: The anomalies of surface pressure (hPa) and divergent wind (m/s) at 1000hPa for +1K/day heating at 140°E a) at 1 day, b) at 3 day, c) at 6 day, d) at 9 day, e) at 12 day, and f) at 15 day. Black dots show the center of the forcing.
generate the asymmetry of the diabatic heating and precipitation response.

This heating over the western Pacific can be considered as upward motion during La Niña. By the asymmetry of the atmospheric response the upward branch of anomalous Walker circulation can preferentially influence Indian monsoon during La Niña (i.e. greater response to heating in the western than in the eastern Pacific). Wang et al. (2003) also suggested that this asymmetry of the circulation might be able to explain the complexity of precipitation related to anomalous tropical heating during ENSO.

4.2.2 Remote response

There are two manifestly remote responses in these experiments. Strong positive diabatic heating anomalies appear in the vicinity of India and negative anomalies over the equatorial Indian Ocean in figure 4.4b to g (the heating at 80°E~180°). The positive anomalies of the diabatic heating near the Indian region will be related to anomalous zonal wind near the surface and latent heat flux. The negative anomalies of the diabatic heating over the equatorial Indian Ocean will be related to vertical motion and meridional divergent wind.

In figure 4.13, zonal wind anomalies at 850hPa from each run (a~k) and the summer mean wind of climatological run (l) are shown. Westerly wind impacting the Indian monsoon region is seen in the Arabian Sea in the climatological run. In the case of the forcing in the range of 80°E (b) to 180° (g), positive anomalies appear and strengthen the total westerly wind near India. In the forcing at 60°E, negative wind anomalies weaken the westerlies. The forcing over the cold SST (160°W to 100°W: figure 4.13h~k) leads to very weak wind anomalies over the Indian region. The latent heat flux anomalies shown in figure 4.14 are generally positive near the Indian region where the low-level wind is strengthened due to increased wind-induced evaporation,
consistent with positive anomalies in the diabatic heating.

Since the SST is prescribed in these experiments, the ocean temperature cannot respond to the loss of the energy due to positive latent heat flux anomalies. Thus the stronger wind can keep supplying latent heat flux from the ocean to the atmosphere. Without air-sea interaction this one-way process can continue to increase the latent heat from the ocean, leading to an exaggeration of the diabatic heating. In the following chapter, we will study how the air-sea interaction modifies the anomalous latent heat flux.

One of the main purposes of this study is to simulate anomalous vertical motion in the Maritime Continent (anomalous Walker Circulation) during El Niño and La Niña. The response of vertical motion (figure 4.15) and divergent meridional wind (figure 4.16) for each heating location illustrates the simulated anomalous Walker circulation in the AGCM. The vertical motion in a simple dry model consists of a single upward branch of the Walker circulation; in the AGCM the interaction between convection and large-scale dynamics creates additional forcing in the remote regions of the forcing, leading to additional complexity.

Vertical sections of omega averaged between 5°S and 5°N are shown in figure 4.15. Over the cold SST the omega response to +2K/day idealized heating at 120°W and 100°W is plotted (figure 4.15j and k) and the response to +1K/day heating at 60°E ∼ 140°W is plotted in figure 4.15a∼i. The omega response to the +2K/day heating at 120°W and 100°W has a similar magnitude as the response to the +1K/day heating at 140°E and 120°E. Thus, stronger forcing should be inserted over the cold SST to induce the same strength of vertical motion than the forcing over the warm SST. It was also shown previously that the diabatic heating response is relatively weak over the cold SST compared to the response over the warm SST. This is useful to know when we try to mimic observed diabatic heating in the model. Since there is
Figure 4.13: The response of zonal wind (m/sec) at 850hPa for +1K/day heating case. The additional idealized forcing is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W l) climatological mean of zonal wind. The values at 5% significance are shaded for a ~ k. Black dots show the center of the forcing.
Figure 4.14: The same as figure 4.13 except for latent heat flux (Wm$^{-2}$).
Figure 4.15: The response of vertical structure of omega (Pa/s) for +1K/day heating case (a~i) and +2K/day case (j, k). The additional idealized forcing is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. Black dots show the center of the forcing. (See text for details.)
Figure 4.16: The same as figure 4.15 except divergent meridional wind (m/s) at 850hPa
a weak response over the cold SST, here the observed diabatic heating can be inserted directly as an idealized heating. On the other hand, over the warm SST the response is strong, so the reduced observed diabatic heating would be better to induce realistic total diabatic heating in the model.

For the idealized heating over the cold SST (160°W to 140°W: figure 4.15h and j), the anomalous omega ranges from 1Pa/s to 3Pa/s. Over the warm SST the anomalous omega is as large as 9Pa/s. To the west and east of the upward motion induced by the heating, there are anomalous downward compensating motions. When the heating is inserted over the cold SST (at 160°W, 140°W, and 120°W), the descending motion is confined to regions immediately to the east and west of the forcing. On the other hand, when the heating is inserted over the warm SST (at 80°E, 100°E, 120°E, 140°E, and 160°E), remote compensating descending motion over the Indian Ocean and western Pacific is strong. Another downward motion branch appears over the eastern Pacific.

The anomalous upward motion by the idealized heating implies one branch of anomalous Walker circulation. When the additional heating is over the western Pacific considered as a La Niña event, there is compensating downward motion over the Indian Ocean. When the idealized heating is at 180°, it can be considered as a part of El Niño and there is downward motion over the western Pacific and Indian Ocean in figure 4.15g.

The negative diabatic heating occurring in the equatorial Indian Ocean as the remote response is co-located with the compensating downward motion; the descending motion weakens convection over the equatorial Indian Ocean. The compensating downward motion has been also shown in a moist baroclinic model (Watanabe and Jin, 2003).

The response of the divergent meridional wind (the so-called “local Hadley circulation”) at 850hPa is shown in figure 4.16. In all heating regions, there are negative
wind anomalies to the north of the heating and positive anomalies to the south, corresponding to low-level convergence, which in turn is related to ascending motion by the tropical heating (figure 4.15). When the heating is to the west of 180° (over the warm SST), strong positive wind anomalies are seen over the southern tip of India while in the southern Indian Ocean negative wind anomalies appear, leading to divergence. This in turn is related to the downward motion weakening the convection over the equatorial Indian Ocean, as seen in figure 4.15.

4.2.3 Atmospheric circulation

The responses of streamfunction at 850hPa are shown to understand atmospheric response to the heating at the eleven locations in figure 4.17. A common response for all locations is the four-celled pattern of streamfunction anomaly. There is a Rossby response (a pair of cyclonic circulations) to the west of the forcing and a Kelvin response to the east of the forcing; this is the Gill type response. Depending on the location of the forcing, however, there are some variation in the magnitudes and spatial patterns. In the case of the forcing over the warm SST (60°E ∼ 180°), the streamfunction anomalies are stronger, and for the forcing over the Indian Ocean (100°E, 120°E) the response is the strongest. When the forcing is located to the west of the date line (140°E, 160°E, and 180°), an anomalous streamfunction response is anchored over the Indian region. The anchored response looks like an extension or propagation of the Rossby wave. In addition, when the heating is over the eastern Pacific, the Kelvin wave type response seems to propagate eastward to the Indian region (figure 4.17j and k). In the diabatic heating response there are significant positive anomalies over the Indian region in figure 4.4. The circulation response is likely to be a combined response of the direct response to the forcing and the response by the additional diabatic heating over the Indian region.
Figure 4.17: The response of streamfunction ($\times 10^6$ s$^{-1}$) at 850hPa in +1K/day forcing case. The additional idealized forcing is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. 5% significant values are shaded. Black dots show the center of the forcing.
As a summary of the idealized heating experiments, some results of the experiments are compared with the El Niño composite (1982, 1987, 1991, 1994, and 1997) from ERA-40 data in figure 4.18. Since the vertically integrated diabatic heating anomalies of El Niño extend from 150°E to 90°W in figure 4.18a, the diabatic heating responses to the inserted forcing at 180° ∼ 100°W are shown. For the case of 180° (figure 4.18b) the model induces a remote response over the Indian region as the GCM effect. The positive anomalies in the vicinity of India and negative anomalies in the Indian Ocean do not appear in the El Niño composite. On the other hand, over the Maritime continent negative heating anomalies appear in the case of the forcing at 180° and 160°W, corresponding to strong negative values in the El Niño composite. Without the inserted cooling over the Maritime continent in this model the idealized heating over the central Pacific may induce a downward branch of anomalous Walker circulation despite its weak magnitude.

Streamfunction and wind at 850hPa are also shown for the El Niño composite and the idealized forcing located at 180° ∼ 100°W. In figure 4.19a, the El Niño composite has westerly anomalies and two cyclonic circulations over the central Pacific and these are also shown to the west of the forcing location in each heating experiment (figure 4.19b to f). The idealized heating at 180° is close to the Indian region and has an extended response in figure 4.19b. But the elongated response is in the opposite sense compared to the observed anomalies. All other cases have weak but significant easterlies over the Indian Ocean. These might be Kelvin wave-like responses that have propagated eastward all the way to the Indian Ocean. This becomes clear with significant streamfunction anomalies when the heating is over the eastern Pacific such as at 120°W and 100°W. These easterlies over the equatorial Indian Ocean are in the same sense as the observed wind anomalies. The easterlies over the equatorial Indian Ocean is oppose the monsoon flow and cause a weakening of monsoon. Therefore,
Figure 4.18: Diabatic heating anomalies integrated from the surface to 100 hPa (Wm$^{-2}$) a) El Niño composite of diagnosed heating from ERA-40. Shading is 5% significance (purple is for negative and orange for positive). b) sum of the inserted heating and the response to the idealized heating when the heating is centered at 180°, c) 160°W, d) 140°W, e) 120°W, and f) 100°W. From b) to f) values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.19: Streamfunction ($\times 10^6$ s$^{-1}$) and wind (m/s) anomalies at 850hPa a) El Niño composite from ERA-40. b) response when the heating is centered at 180°, c) 160°W, d) 140°W, e) 120°W, and f) 100°W. Values at 5% significance are shown (shading and vectors). Black dots show the center of the forcing.
this implies that if the anomalous warming over the eastern Pacific is strong, it might influence the Indian monsoon and interact with the downward motion over the western Pacific.

For developing La Niña events (1988, 1998, and 1999 northern summer) we expected that the idealized heating near the western Pacific could be compared to our idealized heatings. However, in figure 4.20b, the anomalous streamfunction at 850hPa does not show a response related to anomalous heating or upward motion over the western Pacific. Vertically integrated diabatic heating anomalies are still significantly positive over the western Pacific and Indian Ocean in figure 4.20a. During northern winter of the same La Niña years there are two cyclonic circulations over the western Pacific, related to the anomalous heating over the western Pacific (figure 4.20d). For the corresponding summers, individual events also rarely show the cyclonic circulation over the western Pacific (not shown) so we do not compare La Niña summer with the heating experiments.

4.3 Heating vs. cooling response

4.3.1 GCM effect

The cooling experiments carried out by inserting the same forcing as in figure 4.2 except for a negative sign. To compare the GCM effect of the cooling experiments with the heating experiments, the diabatic heating responses of $\pm 0.5\text{K/day}$, $\pm 1\text{K/day}$, and $\pm 2\text{K/day}$ are shown for two cases; the forcing centered at 120°E (figure 4.21) and 140°E (figure 4.22). While the heating experiments show a local response in figure 4.21b that has two maximum anomalies towards the north and south of the equator (the center of the forcing), in the cooling experiments the local response has maximum anomalies near the center of the forcing (the equator) in figure 4.21e. The
Figure 4.20: La Niña composite from ERA-40. a) diagnosed diabatic heating anomalies (Wm$^{-2}$) vertically integrated from the surface to 100hPa during summer (MJJA). b) streamfunction ($\times 10^6$s$^{-1}$) at 850hPa during summer (MJJA). c) diagnosed diabatic heating anomalies vertically integrated from the surface to 100hPa during winter (NDJ). d) streamfunction ($\times 10^6$s$^{-1}$) at 850hPa during winter (NDJ). Shading is at 5% significance.
different local response between the heating and cooling experiments is also clear with all magnitudes of the forcing.

The remote responses show differences in the heating and cooling experiments. As mentioned in the previous section, positive diabatic heating by strengthened surface wind in the vicinity of India is shown as one of the remote responses in the heating experiments when the forcing is over the warm SST in figure 4.21a, b, and c and figure 4.22a, b, and c. In the cooling experiments, negative anomalies of the diabatic heating appear over India but there are no strong anomalies in the Bay of Bengal comparable in magnitude to those in the heating experiments in figure 4.21d, e, and f. Instead, positive anomalies of the diabatic heating are shown in the north of the Bay of Bengal and the Philippine seas. The positive anomalies in the two regions also appear in the cooling centered at 60°E, 80°E, and 100°E in figure 4.23. The responses of the diabatic heating to added idealized heating and cooling are compared for all 11 locations in figure 4.23.

Over the cold SST two responses are shown for the stronger idealized heating/cooling (±2K/day) at 120°W and 100°W (figure 4.23j, j-1, k, and k-1). The magnitude of the response in the forcing region is similar to the response over the warm SST such as the forcing at 120°E (figure 4.23d and d-1). Although the location of the forcing is far from the India, there are significant diabatic heating anomalies for both heating and cooling experiments in figure 4.23j, j-1, k, and k-1.

Another remote response is negative anomalies over the equatorial Indian Ocean in the heating experiments in figure 4.21a, b, and c. This remote response is consistent with the compensating downward motion of the upward motion by the inserted idealized heating. In the cooling experiments, significant anomalies over the equatorial Indian Ocean are also shown but positive in sign and weaker in magnitudes than in the heating experiments (figure 4.21d, e, and f). The remote responses over the
Figure 4.21: The responses of diabatic heating (Wm$^{-2}$) in the heating (left panel) and cooling (right panel) experiments. The forcing is centered at 120°E. The maximum of each forcing is a) +0.5 K/day, b) +1K/day, c) +2K/day, d) -0.5K/day, e) -1K/day, and f) -2K/day. Statistically significant values at 5% are shaded. Black dots show the center of the forcing.
Figure 4.22: the same as figure 4.21 except the forcing is centered at 140°E. Black dots are the centers of each forcing.
Figure 4.23: The response of diabatic heating (Wm$^{-2}$). The maximum of the inserted idealized forcing is $\pm 1$K/day for a~i (a-1~i-1) and $\pm 2$K/day for j~k (j-1~k-1). The left panel is the response from the heating experiments and the right panel is the response from the cooling experiments. Heating or cooling is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) are continued on the next page. Statistically significant values at 5% are shaded. Black dots show the center of the forcing.
Figure 4.23: continued
equatorial Indian Ocean disappear when the cooling is centered at 160°E, while the heating experiments still have strong and significant remote responses in figure 4.23.

When the forcing is over the cold SST (160°W ~ 100°W), the local and remote responses are similar in magnitude and patterns between the heating and cooling experiments. However, when the forcing is over the warm SST, the local and remote responses of the diabatic heating as the GCM effect have significant differences in the heating and cooling experiments. These differences might be generated by different interaction between forced atmospheric circulation and moist dynamics.

Figure 4.24 shows the response of precipitation and heating rates by moist processes and longwave radiation for the heating and cooling experiments, when the forcing is centered at 120°E. The diabatic heating rates are the sum of contributions due to moist processes, longwave, and shortwave radiation. Compared to the response of the total diabatic heating in figure 4.21b and e, the heating rate by moist processes in figure 4.24a and b is dominant in both heating and cooling experiments.

In the cooling experiments, the local response of longwave radiation is negative in figure 4.24d, while it is positive for heating experiments in figure 4.24c. By adding the additional cooling, anomalous convection weakens and clouds trap less longwave energy. So more longwave flux goes out to the space than incoming longwave flux to the surface. There is no response in the remote region of the forcing, while the heating experiments have significant remote positive anomalies in the Bay of Bengal and negative anomalies over the equatorial Indian Ocean in figure 4.24c. The response of precipitation in figure 4.24e and f in the heating and cooling experiments has similar features to the corresponding response of the heating rates by moist processes.
Figure 4.24: The response to the forcing at 120°E. The left panel is for the heating experiments. The right panel is for the cooling experiments. Heating rates by moist process (Wm$^{-2}$) for the heating (a) and cooling experiments (b). Long wave radiation (Wm$^{-2}$) for the heating (c) and cooling experiments (d). Precipitation (mm/day) for the heating (e) and cooling experiments (f). a)~d) are integrated vertically from the surface to 100hPa. The values at 5% significance are shaded. Black dots show the center of the forcing.
4.3.2 Atmospheric circulation: local response

The local response of the diabatic heating to forcing at 120°E (shown earlier) is compared to the surface divergence fields in figure 4.25a. The anomalous diabatic heating is centered to the north and south of the heating center in figure 4.25a and two cells of convergence near the surface in figure 4.25c also appear to the north and south of the heating center for the heating experiment. However, the divergence field integrated from 275hPa to 80hPa shows only one divergent center located over the forcing region. The local diabatic heating response can be most closely related to the surface convergence in the forcing region. On the other hand, in the cooling experiments, the response of divergence fields near the surface has maximum anomalies near the center of the forcing in figure 4.25d. One anomalous divergence center is seen near the surface in figure 4.25d and one anomalous convergence center at the upper level (figure 4.25f). North-south dipole local separated responses do not appear in the cooling experiments.

In contrast, when the additional cooling is over the cold SST (160°W, 140°W, 120°W, and 100°W), the local response of diabatic heating is similar in pattern to the response in the heating experiments (figure 4.23h~k and h-1~k-1). Figure 4.26 shows the response of diabatic heating and divergence fields for the 140°W forcing experiments. In both heating and cooling experiments, divergence fields integrated from the surface to 800hPa show only one cell at the forcing region, convergence and divergence for the heating and cooling experiments, respectively (figure 4.26c and d). In 275-80 hPa divergence fields, there are two cells of divergence and convergence for each experiment (figure 4.26e and f).

Here, two anomalous upper level divergence centers seen in figure 4.26e are partly aligned with the two cyclonic centers in the classic Gill responses. The upper level divergence induces anomalous upward motions reinforcing the diabatic heating. The
Figure 4.25: The response to the forcing centered at 120°E. The left panel is the heating experiments. The right panel is the cooling experiments. Diabatic heating response (Wm$^{-2}$) for the heating (a) and cooling (b) experiments. Divergence fields integrated from the surface to 800hPa for the heating (10$^{-3}$kg/(m$^2$s)) (c) and cooling (d) experiments. Divergence fields integrated from the 275hPa to 80hPa for the heating (e) and cooling (f) experiments (10$^{-3}$kg/(m$^2$s)). The values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.26: the same as figure 4.25 except the forcing centered at 140°W.
same reinforcement is seen in the cooling experiment. For both heating and cooling experiments the anomalous convergence (divergence) at the upper level is accompanied with the local response of the meridionally extended diabatic heating in figure 4.26a and b. On the other hand, when the heating is over the warm SST, the anomalous convergence near the surface occurs with the local response of the diabatic heating.

4.3.3 Different latitudinal forcings

The local response is sensitive to the latitudinal location of the forcing. In this set of experiments, the forcing is inserted at $7^\circ$S and $7^\circ$N at the same 11 longitudinal locations. To consider the local response near only the Indian region, two cases are shown for the forcing at $100^\circ$E (figure 4.27) and at $120^\circ$E (figure 4.28) with three latitudinal locations ($7^\circ$N, the equator, and $7^\circ$S). When the additional heating is inserted at $100^\circ$E (figure 4.27a, b, and c) and $120^\circ$E (figure 4.28a, b, and c), the Bay of Bengal and northern India have significant positive diabatic heating anomalies. Although the magnitudes of the positive anomalies become weak as the forcing goes southward, there are still large anomalies.

On the other hand, when the cooling is at $7^\circ$S and $100^\circ$E ($7^\circ$S and $120^\circ$E), anomalously negative diabatic heating anomalies in the southern Bay of Bengal weaken or disappear, although positive heating anomalies still remain in the northern Bay of Bengal. For the cooling experiments the response of surface wind consists of easterlies over southern India and the equatorial western Indian Ocean in figure 4.27d, e, and f and figure 4.28d, e, and f. When the cooling is at $7^\circ$S and $120^\circ$E (figure 4.28f), the easterlies over southern India weaken. The response of streamfunction at 850hPa also shows the weakening of the anti-cyclonic circulation for the forcing at $7^\circ$S compared to the response of other latitudinal locations of the cooling in figure 4.29. The weakening and disappearance of the local response in India is due to a
Figure 4.27: The response of wind (m/s) at the surface and diabatic heating (shading) vertically integrated from the surface to 100hPa (Wm$^{-2}$) for the forcing at 100°E. a) is for the forcing of +1K/day at 7°N. b) is for +1K/day at the equator. c) is for +1K/day at 7°S. d) is for -1K/day at 7°N. e) is for -1K/day at the equator. f) is for -1K/day at 7°S. The values at 5% significance are plotted as shading and vectors. Black dots show the center of the forcing.
Figure 4.28: The same with figure 4.27 except the forcing is at 120°E.
Figure 4.29: The response of stream-function \( \times 10^6 \text{s}^{-1} \) at 850hPa for the forcing at 120E. a) is for the forcing of +1K/day at 7°N. b) is for +1K/day at the equator. c) is for +1K/day at 7°S. d) is for -1K/day at 7°N. e) is for -1K/day at the equator. f) is for -1K/day at 7°S. The values at 5% significance are shaded. Black dots show the center of the forcing.
meridionally longer distance of the forcing to the Indian region. The response of the cooling becomes weaker than the heating when the forcing is at 7°S.

As a summary of the local response, the local response to the idealized heating (cooling) includes additional heating (cooling) which extends towards the Indian region (the GCM effect). The idealized heating and cooling induces anomalous upward and downward motion as one branch of Walker circulation over the western Pacific, which does not cover the Indian region. The upward and downward motion can be considered as La Niña and El Niño, respectively. The Indian region is under the influence of one branch of Walker circulation over the western Pacific, which is opposite to the other branch over the central/eastern Pacific. Therefore, simple forcings can generate the various relationships between ENSO and the Indian monsoon by the GCM effect in these experiments. In addition, this local response is sensitive to the latitudinal location of the anomalous cooling. So the details of spatial pattern in terms of anomalous diabatic heating are important to decide the influence of ENSO on the Indian monsoon.

4.3.4 Atmospheric circulation: remote response

In the previous section, the remote response to idealized heating in the vicinity of the Indian region is related to increased latent heat flux by wind anomalies near the surface. The response of latent heat flux is shown for both heating and cooling experiments in figure 4.30. The latent heat in the heating experiments has significantly positive anomalies near India in the cases of the forcing ranging from 80°E to 180°, while in the cooling experiments significantly positive anomalies of the latent heat flux near India appear only in the cases of the forcing ranging from 60°E to 120°E. For the cooling experiments a remote response in the vicinity of the Indian region appears only when the cooling is close to the Indian region.
When the response of the zonal wind is significant near India (forcing at 60°E, 80°E, 100°E, and 120°E) for both experiments, the spatial pattern of the anomalies in figure 4.31 is symmetric (that is with the opposite sign). However, the response of the latent heat flux near India is not symmetric in figure 4.30. In the cooling experiments at 80°E, 100°E, and 120°E, positive anomalies of the heat flux are shown; these anomalies are of the same sign as those in the heating experiments. Positive anomalies of the heat flux occur in locations of positive zonal wind anomalies in the vicinity of India, while negative wind anomalies occurred with very weak negative anomalies of latent heat flux. The forcing induces cyclonic circulation or anticyclonic circulation (depending of the sign of the forcing) but only positive wind anomalies induce latent heat flux anomalies for both cases by strengthening the monsoon flow. In the vicinity of the Indian region, weakened winds may not play an efficient role in the transfer of the latent heat flux from the atmosphere to the ocean. This leads to the asymmetry of the remote response in the total diabatic heating near India between the heating and cooling experiments in figure 4.23.

The remote response over the equatorial Indian Ocean is related to the vertical motion and divergence (anomalous Walker circulation) as discussed previously. The responses of omega are shown in figure 4.32 to compare the anomalous Walker circulation between the heating and cooling experiments. When the idealized negative temperature tendency (cooling) is added in the model, anomalous downward motion is seen as a direct response of the cooling (right panel of figure 4.32). For the experiments of the cooling located at 100°E, 120°E, and 140°E (figure 4.32c-1, d-1, and e-1), the anomalies of omega are relatively large. For the forcing centered at 160°E (figure 4.32f and f-1), the descending motion (direct response) becomes weak in the cooling experiment, while the ascending motion (direct response) is strongest in the heating experiment. This means a positive feedback between rising motion and moist
Figure 4.30: The response of latent heat flux (Wm$^{-2}$) in the heating experiments (left panel) and cooling experiments (right panel). Heating or cooling is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) are continued on the next page. The values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.30: continued
Figure 4.31: The response of zonal wind (m/s) at the surface in the heating (left panel) and cooling (right panel) experiments. Heating or cooling is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) are continued on the next page. l) summer mean of zonal wind at the surface in climatology run. The values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.31: continued
heating. This is consistent with the control run summer mean omega showing strong upward motion with strong diabatic heating (mainly moist heating) 60°E to 160°E, as shown in figure 4.32l and l-1. Thus, the direct response of vertical motion in the heating experiments is more sensitive than in the cooling experiments. On the other hand, when the forcing is over the cold SST (160°W, 140°W, 120°W, and 100°W), the response of the omega is similar (about 3Pa/s anomalies) for both experiments in figure 4.32h~k and h-1~k-1. The direct response of the vertical motion is sensitive to the location of the forcing and it is different depending if the forcing is heating or cooling.

In addition, the compensating vertical motion differs between the heating and cooling experiments. When the forcing is centered at 160°E, in the heating experiments strong upward motion (maximum is -9Pa/s) appears with downward motion as compensating vertical motion over the equatorial Indian Ocean in figure 4.32f. However, the cooling experiment at 160°E has about 4Pa/s downward motion near the forcing region and very weak downward motion over the equatorial Indian Ocean in figure 4.32f-1. The forcing at 180° also shows weak direct and remote responses of vertical motion in figure 4.32g-1. The existence of the compensating motion is dependent on the strength of the direct response from the forcing. The weak remote response of the omega in the cooling experiments also agrees with the weak remote responses of the diabatic heating (figure 4.23). Thus, the different sensitivity of the vertical motion for the heating and cooling experiments might directly influence the remote response of the diabatic heating over the equatorial Indian Ocean.

To understand the response of large-scale atmospheric circulation in three dimensions, consistency with the vertical motion and surface divergence fields will be useful. In this section the response of divergent meridional wind at 850hPa is used to understand near surface divergence (figure 4.33). This is a measure of the local Hadley
Figure 4.32: The response of omega (Pa/s) in the heating experiments (left panel) and cooling experiments (right panel). The response is averaged between 5°S and 5°N. Heating or cooling is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) are continued on the next page. The values at 5% significance are shaded. l) mean (May ∼ Aug) of omega from the control run. l-1) mean of diabatic heating (K/day) from the control run. Black dots show the center of the forcing.
Figure 4.32: continued
circulation. In the cooling region (see right panel), there are positive anomalies to the north and negative anomalies to the south of the forcing, giving divergence near the surface. The downward motion by the cooling (figure 4.32) is consistent with surface divergence in figure 4.33. When the upward motion appears as the remote response over the equatorial Indian Ocean in figure 4.32, convergence also occurs to compensate this ascending motion. The anomalous divergence of the meridional wind only seems to agree with the vertical motion in the cooling experiments, and a similar manner in the heating experiments. When the cooling is centered at 160°E and 180°, the remote response of the convergence in the Indian Ocean becomes weak.

We look at the response of the vertical motion and divergence in an adiabatic framework. Gill (1980) studied the atmospheric response by tropical heating in terms of the planetary Rossby and Kelvin waves. To see the sensitivity of this type of the forced atmospheric response, the anomalies of streamfunction at 850hPa are shown in figure 4.34. When the forcing is centered at 140°E (figure 4.34e), there are four cells: one pair of cyclonic circulations west of the heating (Rossby response) and a Kelvin response to the east of the forcing (Gill-type response). Depending on the longitude of each forcing, the magnitudes and patterns are different for both heating and cooling experiments.

In the cooling experiments, similar to the heating experiments, there are anchored responses near the Indian region in spite of changing forcing locations. When the forcing is over the warm SST (from 80°E to 180°: figure 4.34a∼g and 4.34a-1∼g-1), the Rossby response is stronger in the heating experiments than the cooling experiments. In the heating experiments, the stronger response of the streamfunction matches well with the stronger response of the diabatic heating (figure 4.23) compared to the cooling experiments. The inserted idealized heating and cooling have the same magnitude so the asymmetry of the atmospheric circulation is due to the different GCM effect in
the heating and cooling experiments.

When the forcing is over the cold SST (from 160°W to 100°W: figure 4.34h-1~k-1), the magnitudes of the Rossby response are similar in both heating and cooling experiments. However, the Kelvin-like responses to the east of the forcing has propagated and traveled until the Indian region in the heating experiments, but this does not appear in the cooling experiments. The cooling (heating) over the eastern Pacific can be considered as a part of La Niña (El Niño) forcing. The atmospheric response of El Niño forcing only over the eastern Pacific propagates toward the Indian region. It may also contribute to the asymmetry of the response between El Niño and La Niña.

In the cooling experiments, the additional diabatic heating is weaker than in the heating experiments. As a result, the vertical motions of the direct and remote responses become weak and in turn, any Gill type response to induced atmospheric anomalies becomes weak as well. Here the local and remote responses of diabatic heating are generated in the context of fixed SST but in the next section air-sea interaction will be considered to verify the additional diabatic heating in this section.

As a summary of the cooling experiments, some results are compared with the El Niño and La Niña composites. First, the El Niño composites of diabatic heating and wind anomalies are shown from the ERA-40 derived heating with the response of cooling at 100°E, 120°E, and 140°E with 7°S in figure 4.35. By inserting the idealized cooling near the Maritime continent (black contours in figure 4.35b, c, and d), the GCM effect induces anomalous negative diabatic heating near the coast of Sumatra and the Maritime continent and this implies a downward branch of anomalous Walker circulation during El Niño. The strong response near the coast of Sumatra (figure 4.35b and c) is similar to the El Niño composite (figure 4.35a). In figure 4.35e, the composite of streamfunction at 850hPa shows two anti-cyclonic circulations near India
Figure 4.33: The response of divergent meridional wind (m/s) at 850hPa in the heating experiments (left panel) and cooling experiments (right panel). Heating or cooling is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) are continued on the next page. The values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.33: continued
Figure 4.34: The response of streamfunction \((\times 10^6 \text{s}^{-1})\) at 850hPa in the heating experiments (left panel) and cooling experiments (right panel). Heating or cooling is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) are continued on the next page. The values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.34: continued
and the surface wind anomalies indicate easterlies to the south of India (the equatorial Indian Ocean). The two anti-cyclonic circulations and easterlies are produced by inserting only cooling near the Maritime continent (figure 4.35f, g, and h) as the El Niño response. When the inserted cooling is relatively close to India (figure 4.35b: 100°E), the anomalous circulation is strong. When the cooling is relatively far from India such as the cooling at 140°E (figure 4.35d), the anomalous circulation becomes very weak (figure 4.35h). Thus, if the strength of the anomalous circulation is tightly related to the Indian precipitation, the location of anomalously negative diabatic heating near the western Pacific associated with El Niño could be an important factor for the monsoon prediction.

In the previous section describing the idealized heating experiment during the summer of La Niña, anomalous cyclonic circulation related to anomalously positive diabatic heating over the western Pacific (figure 4.36a) is not strong near the Indian region in figure 4.37a. Thus, we could not compare the results of the idealized heating experiments with the observed anomalies. Yet during summer of La Niña anomalous anticyclonic circulation (figure 4.37a) is strong to the west of the anomalously negative diabatic heating over the central/eastern Pacific (figure 4.36a). The observed atmospheric circulation can be compared with the results of the idealized cooling experiments.

The composites of observed La Niña (figure 4.36a) are compared with the idealized cooling experiments. Since the locations of the observed anomalously negative diabatic heating in figure 4.36a extends from 150°E to 120°W, the response (added to the inserted cooling) of diabatic heating is shown for the cooling experiments (additional cooling over the central and eastern Pacific in figure 4.36b~f). Although the magnitude of the inserted cooling is constant (1K/day), the diabatic heating response is strongest at 160°E. When the cooling is at 140°W and 120°W, the response
Figure 4.35: El Niño composite of diabatic heating (Wm$^{-2}$; vertically integrated from the surface to 100hPa) (a) and 850hPa stream-function with the surface wind (m/s) (b). The diabatic heating response at 7°S, 100°E (b), 7°S, 120°E (c), and 7°S and 140°E (d). Contours in b)–d) are inserted idealized cooling (contours are -20 and 50 Wm$^{-2}$). The responses of streamfunction at 850hPa and the surface wind at 7°S, 100°E (f), 7°S, 120°E (g), and 7°S and 140°E (h). For shading in a) is 5% significance (purple/orange is negative/positive t-value) and from b) to h) values are selected at 5% significance and plotted (shading and vectors). Black dots show the center of the forcing.
Figure 4.36: Diabatic heating anomalies integrated from the surface to 100 hPa (Wm$^{-2}$) a) La Niña composite of diagnosed diabatic heating from ERA-40. Shading is 5% significance (purple is for negative and orange for positive). b) response with the inserted idealized cooling when the heating is centered at 160°E. c) at 180°, d) at 160°W, e) at 140°W, and f) 120°W. From b) to f) values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.37: Streamfunction ($\times 10^6s^{-1}$) and wind (m/s) anomalies at 850hPa. a) La Niña composite from ERA-40. b) response when the cooling is centered at 160°E. c) at 180°, d) at 160°W, e) at 140°W, and f) 120°W. Values at 5% significance are plotted (shading and vectors). Black dots show the center of the forcing.
is relatively small but there are significantly positive diabatic heating centers near
the Indian region. In the atmospheric circulation response, the cooling experiments
simulate some characteristics of the La Niña composites. In figure 4.37, the observed
anomalies show easterlies over the central and western Pacific, and two anti-cyclonic
circulations (albeit at higher latitudes). In the experiments, there are also easter-
lies to the west of each forcing in figure 4.37b~f. When the cooling is relatively
close to the western Pacific, significant easterlies extend into the Indian region, which
does not appear in the observed wind anomalies. Moreover, when the cooling is at
140°W and 120°W, there are westerlies over the equatorial Indian Ocean, which is the
same direction with the monsoon flow. Although this does not occur in the observed
wind of the composites, the strong cooling over the eastern Pacific might cause the
circulation anomalies to propagate toward India, thus strengthening the monsoon.
Therefore, the Indian monsoon can be sensitive to the spatial pattern of anomalous
cooling during La Niña.

So far the idealized heating and cooling have been inserted in the model with the
same width and strength. As in the summary figures, each forcing can be a part of
the mean diabatic heating during El Niño or La Niña. For example the cooling over
the western Pacific can be a downward branch of Walker circulation associated with
El Niño, while the cooling over the eastern Pacific can be associated with La Niña.
The partial forcing can produce some features of observed atmospheric circulation
and show the possibility of sensitivity of the real atmosphere to the locations of
anomalous diabatic heating. This might be helpful to understand individually the
monsoon’s response to distinct ENSO events.
4.4 SOM effect

4.4.1 Heating experiments

The slab ocean model (SOM) is used in the Indian Ocean and western Pacific in the new experiments with the same idealized forcing as in the previous section. The responses to the additional heating centered at 120°E with the SOM experiments are compared ones with the non-SOM experiments in figure 4.38. In the non-SOM experiments (previous section), a remote appeared anchored over India for a variety of forcing location. For the case of 120°E, the positive anomalies of the diabatic heating over the ocean near India in the non-SOM experiments in figure 4.38a disappears by applying SOM in figure 4.38b. The significant positive latent heat flux in the vicinity of India in figure 4.38c also does not appear in the SOM experiments in figure 4.38d. Instead of the latent heat flux, negative SST anomalies appear in the vicinity ocean of India in figure 4.38g.

In the previous section, we found that the inserted heating induces anomalous circulation over the Indian region and this circulation strengthens the existing monsoon flow. The strengthened wind near the surface induces the transfer of latent heat flux from the ocean to the atmosphere. In the non-SOM experiment, SST does not respond to the change of heat flux at the surface so the increased latent heat flux continues to supply heat to the atmosphere. On the other hand, in the SOM experiment the increased latent heat flux constitutes a loss of the heat from the ocean so the SST cools down. The reduced SST cannot supply continuous latent heat flux to the atmosphere. This relationship is shown in figure 4.38 as the disappearance of the latent heat flux and weakened SST in the vicinity of India. Therefore, the remote response of GCM effect in the vicinity of India is due to lack of air-sea interaction.
In order to see the change of the local response by the air-sea interaction, divergence fields are integrated from the surface to 800hPa for both SOM and non-SOM experiments in figure 4.39. The local divergence response in the heating experiments with the non-SOM mirrors the north and southward centered local response of the diabatic heating in figure 4.38a. The same features also appear in the SOM experiments in figure 4.38b and figure 4.39b although the magnitudes of the local response in the diabatic heating and convergence are smaller than in the non-SOM experiments.

Since the north and southward centered responses of the diabatic heating are related to the surface moisture convergence, the decreased magnitudes of the responses in terms of the diabatic heating and convergence fields might be related to the surface condition such as anomalous SST. In figure 4.38g, the northern Indian Ocean and western Pacific show negative anomalous SST. It is possible that the ocean does not provide as much moisture convergence with the anomalously cold SST as in the non-SOM with the prescribed SST. The responses of the diabatic heating in the SOM become smaller than in the non-SOM due to the negative anomalous SST. So this relationship appears as a negative feedback of the air-sea interaction in the heating experiments.

In figure 4.39e and f, the responses of streamfunction at 850hPa are shown for both non-SOM and SOM. In the SOM experiments, the two cyclonic circulations to the west of the forcing weaken and the spatial pattern is not as extended towards the Indian region as in the non-SOM experiments. The maximum of the circulation is close to the forcing region in the SOM, while the maximum is in the Bay of Bengal in the non-SOM. This might be due to the strong response of the diabatic heating in the Bay of Bengal in the non-SOM (figure 4.38a). The anomalous divergent meridional wind also becomes weak with the weakened response of the diabatic heating in the Indian Ocean and the Arabian Sea in the SOM in figure 4.39d. The air-sea interaction
Figure 4.38: The response to the heating centered at 120°E in non-SOM (left panels) and SOM (right panels). a), b) diabatic heating integrated from the surface to 100hPa (Wm$^{-2}$). c), d) wind at 850hPa (m/s) and latent heat flux (shading) (Wm$^{-2}$). Mean wind and LH (shading) for the non-SOM(e) and the SOM control run(f). The purple contours in f) are zonal wind anomalies of SOM from non-SOM (-0.8,-0.4, 0, 0.4, and 0.8m/s). Values (a~d) of shading and contour are selected at 5% significant level. f) difference (contours) in SST (K) between two forced experiments: one with SOM and the other with climatological SST; shading indicates 5% significant. Black dots show the center of the forcing.
Figure 4.39: The response to the heating centered at 120°E without SOM (left panels) and with SOM (right panels). a), b) divergence fields integrated from the surface to 800hPa. \((10^{-3}\text{kg/(m}^2\text{s)})\) c), d) divergent meridional wind (m/s) at 850hPa. e), f) stream-function at 850hPa \((\times10^6\text{s}^{-1})\)). All values of shading and contour are shaded at 5% significant level. Black dots show the center of the forcing.
induces the negative feedback between the additional diabatic heating and SST as the GCM effect and also the changes atmospheric circulation in the case of the forcing at 120°E.

The responses of the diabatic heating for all eleven locations of added idealized heatings for both non-SOM and SOM are shown in figure 4.40. Similar to the case of 120°E centered heating in figure 4.38, when the heating is centered at 60°E ∼ 180°, the remote response in the vicinity of the Indian region and the equatorial Indian Ocean disappears or weakens and the local response at each forcing region also weakens by applying the SOM. With SOM the local response is limited to the forcing region and does not extend toward to the Indian region. These experiments imply that an upward branch of anomalous Walker circulation should be near the Indian region for La Niña to have an impact on the Indian monsoon in the presence of a SOM. When the heating is over the cold SST (160°W ∼ 140°W), the response is almost the same for both non-SOM and SOM. When the heating is stronger (+2K/day) at 120°W and 100°W than +1K/day, there are significant responses near the Indian region for non-SOM but in SOM the response appears over only central and eastern Pacific.

The responses of streamfunction at 850hPa are shown for all locations of the heating for both non-SOM and SOM in figure 4.41. Also similar to the heating centered at 120°E in figure 4.39, the maximum of the cyclonic circulation is near the western Philippine Sea in the SOM (right panels of figure 4.41 corresponding to heating at 100°E ∼ 160°E); in the non-SOM the corresponding anchored response over the Bay of Bengal remains strong as the idealized heating moves eastward (left set of panels). The anchored cyclonic responses over India weaken in the SOM experiments, although the cyclonic circulation is still extended westward towards India. For the heating ranging from 60°E to 120°E the cyclonic circulations are stronger in the non-SOM experiments. This agrees with the weakened local and remote response by the
Figure 4.40: The response of diabatic heating (Wm$^{-2}$; vertically integrated from the surface to 100hPa) in the heating experiments. The left panel is for non-SOM and the right panel is for SOM. The maximum of the idealized heating is +1K/day for a~i (a-1~i-1 for SOM) and +2K/day for j~k (j-1~k-1 for SOM). Heating is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) (g-1 to k-1) are continued on the next page. Statistically significant values at 5% are shaded. Black dots show the center of the forcing.
Figure 4.40: continued
Figure 4.41: The same with figure 4.40 except streamfunction at 850hPa ($\times 10^6$ s$^{-1}$).
Figure 4.41: continued
SOM in the vicinity of the Indian region in terms of the diabatic heating.

When the heating is +2K/day at 120°W and 100°W, with SOM the diabatic heating response near the Indian region disappears in figure 4.40k-l compared to SOM in figure 4.40k. Yet the atmospheric circulation still shows significant anomalies over the Indian region. The Kelvin wave type anomalies (positive streamfunction anomalies) even become stronger in SOM than in non-SOM.

4.4.2 Cooling experiments

We have seen that the responses in the idealized cooling experiments are smaller than those in the corresponding heating experiments with the non-SOM. In this section, the responses of the cooling experiments with the SOM are shown to study how the air-sea interaction modifies the response of the cooling experiments. First, one example is shown for the case of the forcing centered at 120°E in figure 4.42. The positive anomalies of the diabatic heating in the north western Pacific and the northern Bay of Bengal in the non-SOM (figure 4.42a) disappear in the SOM experiments (figure 4.42b). In the prescribed SST experiments, positive latent heat flux in the northwestern Pacific and the northern Bay of Bengal appears with wind anomalies in figure 4.42c. On the other hand, when the air-sea interaction is applied to this region, the anomalies of latent heat flux (figure 4.42d) disappear and instead negative anomalies of SST (figure 4.42g) appear in the Bay of Bengal and western North Pacific. Similar to the heating experiments with the SOM, in the Bay of Bengal and western North Pacific the remote response of the diabatic heating related to latent heat flux also disappears in the cooling experiments.

While the local response in the idealized heating experiments generally has two centers to the north and south of the forcing, the cooling experiments have the maximum response near the forcing center for both SOM and non-SOM experiments. The
Figure 4.42: The response to the cooling centered at 120°E in non-SOM (a left panel) and SOM (a right panel). a), b) diabatic heating integrated from the surface to 100hPa (Wm$^{-2}$). c), d) wind at 850hPa (m/s) and latent heat flux (shading) (Wm$^{-2}$). Mean wind and LH (shading) for the non-SOM (e) and SOM control run (f). The purple contours in f) are zonal wind anomalies of SOM from non-SOM (-0.8, -0.4, 0, 0.4, and 0.8 m/s). Values (a~d) of shading and contour are selected at 5% significant level. g) difference (contours) in SST (K) between two forced experiments: one with the SOM and the other with the climatological SST and shading is 5% significant. Black dots show the center of the forcing.
difference of the local response in the SOM is a stronger negative response near Sumatra (southwest of the forcing center), and a wider negative response over the western north Pacific. Why does this strengthened negative response occur? This difference may be due to the air-sea interaction. The responses of the latent heat flux in the non-SOM and SOM experiments are positive and similar in terms of the magnitudes in the forcing region and near the coast of Sumatra. The positive anomalies only weaken the negative responses of the diabatic heating so this is not a direct impact to the strengthened negative response near the coast of Sumatra. Nonetheless, in the SOM experiments, the positive latent heat flux cools SST, here negative anomalies of SST appear over the Maritime continent (figure 4.42e). The negative anomalous SST might strengthen the negative diabatic heating in figure 4.42b. In the non-SOM experiments, SST is prescribed and it cannot have this positive feedback from the ocean. In addition, the wind response is stronger in the SOM results. Off the coast of Sumatra easterly anomalies strengthening the total wind in figure 4.42d are the direct response of the idealized cooling, thus adding to the climatological easterlies (figure 4.42f) in this region to increase the magnitude of the wind.

In addition, by applying SOM the area of significance has also changed in terms of the diabatic heating response in figure 4.42a and b. Particularly, positive diabatic heating anomalies over the equatorial Indian Ocean in figure 4.42a almost disappear in the SOM experiment (figure 4.42b). This positive diabatic heating implies enhanced convection by anomalous upward motion, which is a compensating vertical motion for anomalous downward motion over the western Pacific by the western Pacific cooling (Walker circulation type). Forced dynamics such as the anomalous Walker circulation weakens in the presence of air-sea interaction. Thus, internal variability becomes also important to understand.

The anti-cyclonic responses to the west of the forcing region are stronger in the
SOM (figure 4.43f) than in the non-SOM (figure 4.43e). In addition, the anomalous divergence in the forcing region and divergent meridional wind also are stronger in the SOM experiments. This may be related to the strengthened anomalies of the diabatic heating in the SOM experiments. The strengthened response by air-sea interaction also agrees with the results of Wang et al. (2003): the anticyclonic circulations during developing El Niño are stronger than during the peak of El Niño because the air-sea interactions induce a positive feedback.

The eleven diabatic heating responses in the idealized cooling experiments are shown in figure 4.44. When the cooling is inserted over the warm SST, most local responses of the diabatic heating are stronger in the SOM experiments than in the non-SOM. In addition, there are significant negative anomalies in the Bay of Benga (100°E, 120°E, and 140°E). For 100°E, 120°E, and 140°E cases the local response is more extended northward in the SOM experiments. In the model, the idealized additional cooling generates downward motion as a branch of the anomalous Walker circulation, consistent with El Niño. Since the cooling at 120°E and 140°E has the extended local response over the Indian region, the downward motion as El Niño influences the Indian monsoon in a wide range of longitudes.

The two anti-cyclonic circulations to the west of the cooling region are stronger in the SOM than in the non-SOM (figure 4.45). The strengthened anti-cyclonic circulation is seen when the idealized cooling is over the warm ocean (60°E ~ 180°): figure 4.45a~g). This is due to the strengthened additional diabatic heating by the positive feedback of the cooling experiments with the air-sea interaction. In contrast, the heating experiments showed the negative feedback of weakened diabatic heating and atmospheric circulation in the SOM experiments. Therefore, the feedbacks operating in the SOM cause the effect of air-sea coupling to be opposite in the case of idealized cooling (stronger response) than in the case of idealized heating (weaker response).
Figure 4.43: The response to the cooling centered at 120°E without SOM (a left panel) and with SOM (a right panel). a), b) divergence fields integrated from the surface to 800hPa ($10^{-3}$kg/(m²s)). c), d) divergent meridional wind (m/s) at 850hPa. e), f) stream-function at 850hPa ($\times 10^6$ s⁻¹). All values of shading and contour are selected at 5% significant level. Black dots show the center of the forcing.
Figure 4.44: The response of diabatic heating (Wm\(^{-2}\)) in the cooling experiments. The left panels are for non-SOM and the right panels are for SOM. Cooling is centered at a) 60°E, b) 80°E, c) 100°E, d) 120°E, e) 140°E, f) 160°E, g) 180°, h) 160°W, i) 140°W, j) 120°W, and k) 100°W. g) to k) (g-1 to k-1) is continued in the next. The values at 5% significance are shaded. Black dots show the center of the forcing.
Figure 4.44: continued
Figure 4.45: The same as figure 4.44 except for streamfunction at 850hPa ($\times 10^6$ s$^{-1}$).
Figure 4.45: continued
This effect is summarized in figure 4.46 for the forcing at 120°E with the non-SOM and SOM experiments. For the heating experiments the response of the diabatic heating is weaker in the SOM (figure 4.46a and a-1) by a negative feedback between SST and wind anomalies inducing latent heat flux. In turn, streamfunction and zonal wind anomalies also weakened in the SOM. On the other hand, the cooling experiments show a stronger response of the diabatic heating (figure 4.46d and d-1) by a positive feedback, with increased wind induced latent heat flux leading anomalously cold SST, and enhanced downward motion. Further the streamfunction and zonal wind anomalies also strengthen.

By applying the slab ocean model in the Indian Ocean and western Pacific, the GCM effect of the idealized heating and cooling has changed. In the heating experiments, the air-sea interaction gives a negative feedback to the responses to the additional diabatic heating, while in the cooling experiments the air-sea interaction gives a positive feedback to the responses. The cooling (heating) experiments generate downward motion as one branch of El Niño (La Niña)-like Walker circulation. In the model, the different feedbacks bring the stronger response during El Niño over a wide range of longitudes than La Niña. This non-linearity might be one of reasons in the complexity of the atmospheric and precipitation response during ENSO, and explains the past difficulties in understanding the relationship between ENSO and Indian monsoon.
Figure 4.46: Response to heating at 120°E: a), a-1) diabatic heating integrated from the surface to 100hPa (Wm\(^{-2}\)). b), b-1) streamfunction at 850hPa (\(\times 10^6\)s\(^{-1}\)). c), c-1) Climatological wind (vectors) and zonal wind anomalies (shading) at 850hPa (m/s). Response to cooling at 120°E: d), d-1) diabatic heating. e), e-1) streamfunction at 850hPa. f), f-1) climatological wind (vectors) and zonal wind anomalies (shading) at 850hPa. The left panels are non-SOM experiments and the right panels are SOM experiments. Values at 5% significance are shaded. Black dots show the center of the forcing.
Chapter 5: Realistic experiments

In the previous section, the sensitivity of the idealized, added heating and cooling is tested according to different longitudinal and latitudinal locations. When the idealized cooling near India is in the western Pacific or eastern Indian Ocean, but does not cover India, the local response of the negative diabatic heating extends toward the Indian region and shows significant responses over India. By this effect the downward branch of anomalous Walker circulation induced by the cooling can influence India. This local response may help to explain the relationship between ENSO and Indian monsoon. However, in real El Niño events, the main characteristics are anomalously warm SST over the tropical central and eastern Pacific in addition to the anomalous decrease of convection over the Maritime continent. The anomalously warm SST forces the atmosphere strongly in the tropical central and eastern Pacific (Lau and Nath, 2000, 2003) and influences diabatic heating vertically. Thus, two questions are raised here: does the anomalous diabatic heating at a distance (over the central/eastern Pacific) influence the Indian monsoon? And do the anomalous diabatic cooling in the Maritime continent and the diabatic heating over the central/eastern Pacific interact each other to impact India?

The additional forcing in this section is still idealized and has the vertical structure discussed in Chapter 2. The vertically integrated spatial pattern of the forcing is shown for each experiment in figure 5.1b–d. In order to consider the influence of the anomalously warm SST over the central/eastern Pacific, idealized heating is added over the central/eastern Pacific, with cooling over the western Pacific in figure 5.1b. The combined forcing is considered as a realistic forcing for El Niño-like forcing in this
section. The composite of diabatic heating during El Niño is shown in figure 5.1a. Five El Niño years are selected to composite (1982, 1987, 1991, 1994, and 1997). In the composite, anomalous cooling appears more in the southern Hemisphere. In nature, each El Niño event has a different spatial pattern of anomalous diabatic cooling over the western Pacific and heating over the central/eastern Pacific. Before considering special and complex El Niño events, here the forcing is kept as symmetric with respect to the equator. The feedback of dynamics on heating in GCM (GCM effect) will be added to the idealized forcing as total diabatic heating in the model so the magnitude of the idealized forcing we use is kept smaller than the composite of El Niño events on purpose. The added heating shown in figure 5.1b is broken into two components, and each component of the realistic forcing is inserted into the GCM in separate experiments. The two components are the anomalous cooling over the western Pacific shown in figure 5.1c (WP case) and anomalous heating over the central and eastern Pacific shown in figure 5.1d (CP case).

In the idealized heating experiments of the previous section, the GCM effect is different depending on the longitudinal locations of the additional idealized heating. When the heating is located to the west of 180°, there are significant remote responses near the Indian region. On the other hand, when the heating is to the east of 180°, the remote response does not appear significantly near the Indian region. Therefore, to distinguish which part of the diabatic heating influences India, the heating over the central/eastern Pacific (CP case) is further divided into two components: the eastern and western part of the CP case seen in figure 5.1e (east case) and figure 5.1f (west case). The spatial patterns shown in figure 5.1b∼f with the vertical structure shown in chapter 2 are added in the temperature tendency equation as forcing in the same manner as with the previous experiments.
Figure 5.1: Vertically integrated diabatic heating from the surface to 100hPa (Wm$^{-2}$). 
a) El Niño composite of diagnosed diabatic heating from ERA-40 data. Shading is at 95% confidence. The contours are 200, -100, -50, 50, 100, and 200. b) the cooling over the western Pacific and heating over the central/eastern Pacific (both). c) cooling over the western Pacific (WP). d) heating over the central/eastern Pacific (CP). e) the eastern part of heating over the central/eastern Pacific (east). f) the western part of heating over the central/eastern Pacific (west).
In the previous experiments with prescribed SST, the forcing simulates an exaggerated response over the Indian region. The forced response of wind near the surface boosts the release of latent heat flux from the ocean to the atmosphere. Since the SST is prescribed, the ocean does not cool down by losing heat, so the ocean can keep supplying energy to the atmosphere. In spite of this exaggerated response in the non-SOM experiments, first, the prescribed SST (non-SOM experiments) is used with the realistic forcing and then the ocean is regionally coupled with SOM (SOM experiments) with the same forcing. The same forcing with the non-SOM helps us to study how air-sea interaction modifies the direct response to the forcing by comparison between the non-SOM and SOM experiments.

5.1 non-SOM experiments

5.1.1 GCM effect

The responses of GCM’s internal diabatic heating in each experiment are shown for the non-SOM experiments in figure 5.2. The response of the diabatic heating is considered as the additional heating by GCM (GCM effect). The first goal in this section is to determine if the heating at a long distance from the Indian region (warm SST over the central/eastern Pacific) influences India. When the heating is inserted over the central/eastern Pacific (CP case), significant remote responses over the Indian region are shown in figure 5.2d. These include negative diabatic heating over southern India, western and equatorial Indian Ocean, and positive anomalies over the Bay of Bengal and north western India. These significant diabatic heating anomalies show a possibility of the influence of the CP heating on the Indian region.

To distinguish which part of the CP heating influences India, the eastern and western parts of the CP heating (east and west case) are inserted separately into
Figure 5.2: The response of diabatic heating (Wm$^{-2}$) in each non-SOM experiment is vertically integrated (Wm$^{-2}$). a) the experiment with both cooling over the western Pacific and heating over the eastern Pacific (both). b) linear summation of two experiments from WP and CP (sum). c) the experiment with cooling over WP (WP). d) the experiment with heating over CP (CP). e) the experiment with the eastern part of CP (east). f) the experiment with the western part of CP (west). At 95% confidence level, values are shaded.
the model. In terms of the remote responses, the negative response of the diabatic heating over the Indian Ocean and the positive response in the Bay of Bengal and northwestern India appear for both experiments (figure 5.2e and f). This is similar to the whole CP case responses in figure 5.2d. Although the western part of the CP heating is closer to India than the eastern part, the eastern idealized heating in figure 5.2e shows more significant responses than its western counterpart (figure 5.2f), especially over southern India. So in a real El Niño event, we can expect a different impact on the Indian monsoon depending on details of the spatial pattern of anomalous tropical diabatic heating over the central/eastern Pacific.

To diagnose the downward branch of anomalous Walker circulation during El Niño in the response to the additional cooling over the western Pacific (WP), we consider the model’s anomalous diabatic heating. This experiment is similar to some of the idealized cooling experiment shown in the previous section. The inserted cooling in figure 5.1c does not cover the Indian continent but the response of the GCM’s diabatic heating shows significant negative anomalies in India and the Bay of Bengal in figure 5.2c. As mentioned previously, the diabatic heating consists of moist processes (convection and resolved condensation) and radiative heating, but the moist processes dominate. Anomalously reduced precipitation is expected over the Indian region by the cooling over the western Pacific in this model. The negative anomalies that extend westward toward India are considered as the local response of the cooling over the western Pacific. In addition, positive diabatic heating anomalies over the equatorial Indian Ocean are shown; they are related to compensating vertical motion as discussed previously.

It is of interest to determine if the responses to the cooling over the western Pacific and the heating over the central/eastern Pacific interact with each other in a nonlinear way. The response of the GCM’s diabatic heating to the combined forcing is
shown in figure 5.2a (both), while the sum of the GCM’s diabatic heating for the separate WP and CP experiments is shown in figure 5.2b (sum). One characteristic of the local response induced by the WP cooling is the negative diabatic heating over much of southern India and the surrounding oceans. The response to the added CP heating shown in figure 5.2a shows positive heating anomalies in the Bay of Bengal. Adding both idealized heating fields simultaneously leads to a diabatic heating pattern over the southern Indian Ocean and the Bay of Bengal that more clearly resembles the CP response; the WP response is suppressed in this region in figure 5.2a. This is also the case for the sum of the WP and CP responses shown in figure 5.2b. Comparing magnitudes of heating over the Bay of Bengal, the sum of the separate responses is slightly larger than the response when both idealized heatings are added simultaneously, indicating some non-linear suppression in this region.

On the other hand, the negative anomalies of the diabatic heating over the tropical western Pacific in figure 5.2a (both idealized heatings added) is larger in magnitude than that of either the response to the WP cooling in figure 5.2c (WP) or to the sum of the diabatic heating induced by the WP and CP separately (figure 5.2b: sum). The net effect of both idealized heatings (in panel a or b) is to disconnect the extended local cooling response from that in the Bay of Bengal.

When the heating over the central/eastern Pacific is added to the cooling over the western Pacific, the response of the diabatic heating changes. The responses over the Indian Ocean are mainly influenced and strengthened by the CP heating. The local response to south of the Bay of Bengal weakens by adding the CP heating. However, we know that in the Bay of Bengal and western North Pacific, a non-physical response by the prescribed SST occurs. Hence the current experiments with the local coupling SOM will be applied in the next section.
5.1.2 Atmospheric response

Before the CP heating is added, the local response of the cooling over the western Pacific is shown as the negative diabatic heating anomalies extend toward the Bay of Bengal and India. First, atmospheric responses related to this local response will be considered. In figure 5.3c, the response of the wind at 850hPa also shows significant anomalies near India. There are anomalous easterlies emanating from India towards the western Indian Ocean, where the anomalous winds turn southward. This anomalous wind can be considered as an anti-cyclonic circulation, and can be interpreted as a Rossby response to the anomalous cooling over the western Pacific. The anomalous wind is in opposite direction to the mean monsoon flow shown in figure 5.3a. During the northern summer the climatological surface wind flows toward the Indian subcontinent here, supplying moisture to the monsoon. The opposite wind anomalies shown here weaken the monsoon flow and thus precipitation. This relationship is consistent with the local response, the anomalously negative diabatic heating over India and the Bay of Bengal seen in figure 5.2c.

In addition, the response of vorticity integrated from the surface to 800hPa also shows negative anomalies over the Indian region in figure 5.4c. The anomalous negative seasonal mean vorticity is consistent with fewer monsoon depressions, and relatively small precipitation. So the atmospheric circulation induced by the cooling over the western Pacific weakens precipitation over India.

When the heating over the central/eastern Pacific is inserted simultaneously with the cooling over the western Pacific, the local response of the extended negative diabatic heating is weakened (figure 5.2a: both) and the signal of the wind anomalies related to weak precipitation also changes in figure 5.3b (both). The anomalous wind flowing opposite to the mean monsoon flow in the Arabian Sea becomes weak, and anomalous westerlies over the Indian continent strengthen. These westerly anomalies
are in the same direction with the monsoon flow and strengthen total wind. In figure 5.4a (both), positive anomalies of the vorticity appear in the Bay of Bengal and this strengthened positive vorticity favors increased precipitation.

As before, the anomalous characteristics of the flow near India when both CP and WP idealized heating are used (figure 5.3b, figure 5.4a: both) seem to be more closely related to the response to idealized heating in the CP (figure 5.3d, figure 5.4d: CP) than to cooling in the WP (figure 5.3c, figure 5.4c: WP). This is also shown in the sum of the response in terms of the vorticity (figure 5.4b: sum). Similar to the diabatic heating response, both the eastern and western part of the CP heating show similar wind (vorticity) response with the total CP heating in figure 5.3e (figure 5.4e: east) and figure 5.3f (figure 5.4f: west), although the eastern part of the CP heating has more significant response over the Indian region in spite of the longer distance to India.

Since the experiments carried out with prescribed SST (non-SOM), there are exaggerated responses similar to the previous section. By adding the CP heating, strengthened anomalous westerlies over India, the Bay of Bengal, and the northwestern Pacific appear with the positive anomalous latent heat flux in figure 5.3b (both). As mentioned in the previous section, increased surface wind induces more latent heat flux from the ocean to the atmosphere so in turn the latent heat flux strengthens total diabatic heating in the atmosphere. In the prescribed SST run, the ocean does not cool down by the loss of the energy and so can keep supplying heat to the atmosphere. The increased diabatic heating and circulation is likely to be exaggerated by using the prescribed SST and air-sea coupling will be applied in the next section.

When the cooling over the western Pacific is inserted, the anomalous downward motion over the western Pacific appears as a direct response of the vertical motion. As a compensating motion the upward motion over the equatorial Indian Ocean
Figure 5.3: a) climatology (May~Aug) of wind at 850hPa in non-SOM control. The response of wind at 850hPa (arrows, m/s) and latent heat flux (shading, Wm$^{-2}$) in each experiment. b) both, c) WP, d) CP, e) the eastern part of CP (east), and f) the western part of CP (west). Shading and arrows correspond to 5% significance.
Figure 5.4: The response of vorticity is integrated from the surface to 800 hPa ($10^{-3}$kg/(m²s)). a) both, b) summation, c) WP, d) CP, e) east of CP, f) west of CP. At 95% confidence level, values are selected and plotted.
appears. This anomalous circulation is shown in the vertical structure of equatorial omega and the vertically integrated mid-level omega (figure 5.5c and figure 5.6c: WP). The response of the meridional divergent wind at 850hPa shows divergence with the downward motion over the western Pacific as the direct response and convergence with the ascending motion over the equatorial Indian Ocean as the compensating response (see also figure 5.7b: WP).

When the heating is inserted over the central/eastern Pacific simultaneously with the cooling over the western Pacific, the direct response of the vertical motion appears over the western Pacific and central Pacific as descending and ascending motion, respectively (figure 5.5a and figure 5.6a: both). However, over the equatorial Indian Ocean the compensating upward motion by the WP cooling almost disappears in figure 5.5a and figure 5.6a (both). In the linear summation of the WP and CP forcing (figure 5.5b: sum), a small-scale response is seen (alternating positive and negative anomalies of the omega) over the equatorial Indian Ocean. There are two compensating motions by each forcing: the descending motion by the CP heating (figure 5.5d: CP) and the ascending motion by the WP cooling (figure 5.5c: WP). Over the equatorial Indian Ocean two compensating motions occur with opposite directions so they can cancel each other. In figure 5.6a (both), the vertically integrated mid-level omega also shows the disappearance of the ascending motion and the weakened downward motion over the Indian Ocean.

In addition, by adding the CP heating, the response of the divergent meridional wind shows differences compared to the WP cooling experiment. In figure 5.7a (both), anomalous convergence is seen at the equator over the central Pacific as a direct response to the CP heating. Anomalous divergence is shown as positive and negative anomalies in the v-wind to the north and south (respectively) of the equator over the western Pacific and Indian Ocean. This divergence is the direct response of the
Figure 5.5: The response of vertical omega (Pa/s) in each experiment is averaged from 5°S to 5°N. a) both, b) summation, c) WP, and d) CP. At 95% confidence level, values are selected and plotted.
Figure 5.6: The response of omega (Pa · kg/(m²s)) in each experiment is integrated from 700hPa to 300 hPa. a) both, b) WP, and c) CP. At 95% confidence level, values are selected and plotted.
Figure 5.7: The response of divergence meridional wind at 850hPa (m/s) in each experiment. a) both, b) WP, and c) CP. At 95% confidence level, values are selected and plotted.
WP cooling. In the realistic forcing, only the direct responses of meridional wind are shown. In the WP cooling experiment in figure 5.7b (WP), the anomalous convergence over the western Indian Ocean is seen, but it disappears by adding the CP heating in figure 5.7a (both).

In the non-SOM experiments, by adding the CP heating with the WP cooling as idealized additional heating, the local response of the negative diabatic heating extending westward toward India and the related atmospheric circulation both become weak. The remote response of the CP heating is dominant over the Indian Ocean and Bay of Bengal but in the large scale atmospheric circulation the remote response over the Indian Ocean becomes weak.

5.2 SOM experiments

5.2.1 GCM effect

In this section, the SOM is coupled to the GCM over the Indian Ocean and western Pacific, and the same set of realistic additional idealized forcings used in figure 5.1b~f are inserted into the GCM. In order to compare the response of the GCMs diabatic heating in the SOM experiment, the diabatic heating anomalies in the non-SOM and SOM experiments are shown in the left and right panels of figure 5.8, respectively. Before adding the CP heating the local response of the WP cooling is shown as an extended band of negative anomalies oriented towards southern India and the surrounding oceans in both experiments in figure 5.8c and g (WP). After adding the CP heating, the response in the non-SOM experiments shows the weakened response to the south of the Bay of Bengal (figure 5.8a: both). On the other hand, in the SOM experiments there are still some significant negative anomalies of GCM diabatic heating in this region (figure 5.8e: both).
Figure 5.8: The response of the diabatic heating (W m$^{-2}$) is vertically integrated from the surface to 100hPa for each experiment. The left (right) panel is for the non-SOM (SOM) experiments. a) and e) are both WP cooling and CP heating. b) and f) are summation of two results. c) and g) are WP cooling. d) and h) are CP heating. At 95% confidence level, values are selected and shaded.
Another characteristic of the response is that the added CP heating enhances the negative response of the diabatic heating over the equatorial western Pacific for both experiments. In figure 5.8d and h (CP), the CP heating alone induces the negative anomalies over the western Pacific for both experiments. In addition, the linear summation shows also enhanced the local responses in figure 5.8b and f (sum).

Over other basins the GCM effect has also changed by applying the regional air-sea coupling. The positive anomalies of the diabatic heating in the Bay of Bengal and western north Pacific in the non-SOM experiment (figure 5.8a: both) disappear in the SOM experiment (figure 5.8: both). Previously without air-sea interaction the latent heat flux induced by the forced atmospheric circulation appears with strong positive diabatic heating over the Bay of Bengal and western north Pacific.

Another remote response is the set of negative anomalies over the western equatorial Indian Ocean in figure 5.8a (both); they also disappear when the SOM is applied in figure 5.8e (both). When both the CP heating and WP cooling are inserted separately in the SOM experiments, the two responses over the equatorial Indian Ocean almost disappear. Thus, the air-sea interaction in the SOM experiments removes the diabatic heating response to the forcing (forced dynamics) and the limited statistically significant response over the Indian Ocean implies that internal variability might be more important than forced dynamics.

With the prescribed SST, when the CP heating is added to the WP cooling in figure 5.8a, the responses over the Indian Ocean, Bay of Bengal, and western north Pacific appear strongly and the local response weakens. By applying the regional SOM, on the other hand, the responses over the oceans almost disappear and the local response to the south of the Bay of Bengal still remains. The diabatic heating responses are closely related to the atmospheric response, so we now address how the air-sea coupling interacts with the forced wind anomalies and diabatic heating.
5.2.2 Atmospheric response

The latent heat flux induced by the forced wind in the non-SOM experiment (a left panel of figure 5.9a, b, and c) weakens or disappears to the south of the Bay of Bengal and western north Pacific in the SOM experiment (right panel of figure 5.9e, f, and g). Instead, the loss of latent heat flux in the ocean appears as anomalous negative SST over the northern Bay of Bengal and western North Pacific in figure 5.10a, b, c, and d. This relationship between the atmosphere and ocean is consistent with the previous experiments. The exaggerated latent heat flux is removed and the responses of the total diabatic heating also disappear. The air-sea interaction plays a role to remove the exaggerated GCM effect.

The local response of the diabatic heating in the WP cooling extends toward the Bay of Bengal and India for both non-SOM and SOM experiments (figure 5.8c and g: WP). This local response is related to negative vorticity (figure 5.11b and e: WP) and an opposite direction of surface wind anomalies (figure 5.9b and f: WP) to the monsoon flow (figure 5.9d and h). When the CP heating is added to the WP cooling, in the non-SOM experiment (figure 5.8a) the negative response over the Bay of Bengal is suppressed by the positive diabatic heat flux in the Bay of Bengal in figure 5.8a. The atmospheric circulation related to the positive diabatic heating in the Bay of Bengal also changes as follows: the easterly wind anomalies opposite to the monsoon flow are shifted southward and appear in the western Indian Ocean in figure 5.9a. The negative response of vorticity also does not cover India, and positive anomalies of vorticity are seen in the Bay of Bengal in figure 5.11a. The positive vorticity in the Bay of Bengal and southward shifted easterly response oppose the atmospheric response related to the westward extended diabatic heating anomalies by the WP experiment. In terms of atmospheric circulation, adding the CP heating weakens the influence of the WP cooling on India in the non-SOM.
Figure 5.9: The response of latent heat flux (Wm$^{-2}$) and wind anomalies (arrows) (m/s) at 850hPa. The left (right) panels are for the non-SOM (SOM) experiments. a) and e) are both WP cooling and CP heating. b) and f) are WP cooling. c) and g) are CP heating. d) is climatology wind for the non-SOM control run and f) is climatology wind for the SOM control run and difference (shading in h) of zonal wind between the SOM and non-SOM control run. Except for d) and h) only statistically significant results at 5% are shown.
Figure 5.10: The response of SST (contour, a unit is K.) for the SOM experiments with each forcing. a) both, b) summation, c) WP cooling, and d) CP heating. The shading is significant at 5%.
Figure 5.11: The response of vorticity integrated from the surface to 800hPa \(10^{-3}\text{kg}/(\text{m}^2\text{s})\). The left (right) panels are for the non-SOM (SOM) experiments. a) and d) are both. b) and e) are WP cooling. c) and f) are CP heating. At 95% confidence level, values are selected and shaded.
On the other hand, in the SOM experiments after adding the CP heating, this local response of the anomalously negative diabatic heating extending toward the Bay of Bengal and India still remains significantly. The response of wind at 850hPa is easterly near southern Indian in figure 5.9e (both), which is opposite to the monsoon flow and the negative vorticity covers Indian in spite of its weak magnitudes in figure 5.11d (both). The anomalous easterly wind and negative vorticity in the Indian region favor weak precipitation. Since the air-sea interaction removes non-physical GCM effect, in the vicinity of the Indian region, these negative anomalies of diabatic heating become clear and the relationship between the WP cooling and Indian precipitation remains. Therefore, the response of the both experiment in SOM (figure 5.8e) becomes more like the WP experiment (figure 5.8g) near India than in the case of non-SOM (compare figure 5.8a with figure 5.8c).

The 850hPa streamfunction provides a good summary of the large-scale atmospheric response (figure 5.12). When the realistic forcing is inserted, there are two centers of anti-cyclonic circulation (cyclonic circulation) towards the west of the WP cooling (CP heating) in both non-SOM and SOM experiments (figure 5.12a and b). But in the non-SOM experiments, the anti-cyclonic circulations are weaker and located further south compared to the SOM experiments. This southward shift seems to be related to cyclonic circulation anomalies in the Bay of Bengal. In the non-SOM experiment, anomalously positive diabatic heating appears in the Bay of Bengal and the northwestern Pacific in figure 5.8a and the response to this diabatic heating might help to extend the cyclonic circulation westward in figure 5.12a. However, in the SOM experiment the cyclonic circulation is not as extensive as in non-SOM, and one of the anti-cyclonic circulations covers the entire Indian continent. Since the monsoon flow plays an important role of precipitation over India, this distinct location of the anomalous anti-cyclonic circulation in each experiment may influence Indian precipitation

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differently.

In observed data, the wind anomalies of each El Niño event have different characteristics, but it may be worth distinguishing cases when India had a weak or a strong monsoon during developing El Niño. The rotational wind and streamfunction at 850 from ERA-40 data set is used in figure 5.13. The summer mean wind shows the climatological monsoon flow in figure 5.13a. Four El Niño events are selected as 1982, 1987, 1994, and 1997 and the summer mean of anomalies are shown in figure 5.13b), c), d), and e). By and large all El Niño events have two anti-cyclonic circulation to west of the Maritime continent, which is due to a downward branch of anomalous Walker circulation. However, there are differences of the wind anomalies over the India continent in details.

During the development of the two warm events of 1982 and 1987 India had a weak monsoon. This is consistent with the negative relationship between ENSO and the Indian monsoon. During other events (1994, and 1997), India had a wet and normal monsoon, respectively. The anomalies of wind and streamfunction show corresponding differences. For the dry monsoon in 1982 and 1987 the main anomalies of the wind are easterly to south of India (figure 5.13b and c), which is opposite to the monsoon flow (figure 5.13a). On the other hand, the wet monsoon in 1994 has westerlies across the Indian continent (figure 5.13d), which strengthens the monsoon flow. The easterlies are also shifted further south compared to the dry monsoon. The normal monsoon in 1997 shows a similar wind pattern of anomalies to 1994, despite the weaker magnitude (figure 5.13e). Therefore, the details of anomalous wind during El Niño can be closely related to the strength of the Indian monsoon. If these different atmospheric anomalies are mainly controlled by anomalous tropical diabatic heating over the Maritime continent, it is useful to understand the sensitivity of atmosphere to anomalous tropical diabatic heating according to its location and strength.
Figure 5.12: The response of the streamfunction \((\times 10^6 \text{s}^{-1})\) and rotational wind at 850hPa (m/s) to the realistic forcing (combined both WP cooling and CP heating) a) for the non-SOM experiment. b) for the SOM experiment. At 95% confidence level, values are selected and shaded and plotted. Purple contours in a) and b) are inserted forcing (Wm\(^{-2}\)) and the dotted purple contours are 10 and 30 (WP cooling) and solid contours 10, 30, and 50 (CP heating).
Figure 5.13: Wind (m/s) and streamfunction ($\times 10^6 s^{-1}$) at 850hPa from ERA-40. a) climatology of summer (May ~ Aug). Anomalies of summer are shown during four El Niño events in b) 1982 (dry monsoon) c) 1987 (dry monsoon) d) 1984 (wet monsoon) e) 1997 (normal monsoon).
In the GCM effect, the CP heating induces cooling over equatorial western Pacific in figure 5.14a and b and this anomalous cooling might generate a relatively small anti-cyclonic circulation as a Rossby response over the India in the SOM experiment in figure 5.14d. In the non-SOM, this looks like westward propagated Rossby wave in figure 5.14c. There is another view of this anti-cyclonic circulation near the Indian region. In the previous section, when the heating is over the cold SST (140°W, 120°W, and 100°W), the Kelvin wave response to the east of the heating propagated eastward all the way to the Indian region in figure 5.14g and h. The wind anomalies at 850hPa are shown as significant westerlies over the Indian region in figure 5.14e and f. This can be another hint of the propagated Kelvin wave. So these propagated or extended responses by the CP heating influence and interact with the response of the WP forcing.

So far we have seen a complex response of the diabatic heating and the atmospheric responses in realistic forcing experiments. Therefore, it is of interest to ask if the response to the realistic forcing can be understood as a linear summation (figure 5.15b and f) to the response of the two separate experiments involving added WP cooling (figure 5.15c and g) and CP heating (figure 5.15d and h). The response of the diabatic heating is summed in figure 5.8b and f. Both diabatic heating and stream-function shows weaker magnitudes over India in the full experiment than in the summation, although the main features are similar in both experiments. The non-linear effects induce certain cancellation between the cooling and heating near India, although in the Southern Hemisphere this is not the case.

Previously in the idealized experiments (chapter 4), each idealized heating and cooling was inserted separately into the model. In this section, the idealized heating over the central/eastern Pacific is added simultaneously with the cooling over the
Figure 5.14: The response to the heating over CP from a) to f). The response to heating centered at 100°W from g) to h). The left panels are for non-SOM and the right panels are for SOM experiments. a) and b) is vertically integrated diabatic heating response (W m\(^{-2}\)). c) and d) streamfunction (\(\times 10^6\) s\(^{-1}\)) at 850hPa. e) and f) wind (m/s) at 850hPa. g) and h) streamfunction (\(\times 10^6\) s\(^{-1}\)) at 850hPa. At 5% significant level each value is selected and shaded and plotted.
Figure 5.15: The response of the streamfunction ($\times 10^6$ s$^{-1}$) and rotational wind at 850hPa (m/s). The left (right) panel is for the non-SOM (SOM) experiments. a) and e) are both WP cooling and CP heating. b) and f) are linear summation of two results. c) and g) are WP cooling. d) and h) are CP heating. At 95% confidence level, values are selected and shaded and plotted.
western Pacific to make more a realistic, El Niño-like forcing. In the non-SOM experiments, there are non-physical responses to the south of the Bay of Bengal and western north Pacific so the local response of the western Pacific cooling becomes weak. But by applying the SOM this exaggerated response disappears and the local response related to India becomes clear. In addition, the CP heating plays a role to enhance the downward motion over the Maritime continent. Compensating (downward) vertical motion for anomalous upward motion over the central/eastern Pacific (forced by the CP heating alone) also occurs over the equatorial western Pacific and Indian Ocean in the model.
Chapter 6: Observed forcing experiments

In this section, the observed forcing is inserted into the model. Two warm events, 1997 and 1982 are selected to simulate the influence of El Niño on Indian monsoon. In the previous section, the idealized cooling induced an anomalous atmospheric circulation. That weakened monsoon flow, leading to a weak Indian monsoon. Here we will test whether the observed diabatic heating in the model can simulate reasonable anomalous atmospheric circulation anomalies for the normal monsoon during 1997 event and dry monsoon during 1982 event.

6.1 1997 with prescribed SSTs

Ashok et al. (2004) found that the Indian Ocean warming influences the local Hadley circulation and led to a normal monsoon during developing 1997/98 El Niño in AGCM. Here, we try to use the same argument with the idealized experiments with the climatological SST without SOM before we simulate the observed 1997 forcing experiments with SOM. Since Ashok et al. (2004) used an AGCM with SST forcing, we also compare their findings to our climatological SST experiments with the idealized forcing.

The responses to the idealized cooling at 120°E and the equator are shown in figure 6.1a~c to consider the downward branch of Walker circulation during developing El Niño. In figure 6.1a, the response of vertically integrated mid-level omega shows positive anomalies over the western Pacific and negative anomalies over the equatorial Indian Ocean. The positive anomalies represent one downward branch of
Figure 6.1: The response (a, b, and c) to the idealized cooling centered at 120°E and the equator with non-SOM. a) vertically integrated omega (Pa · kg/(m²s)) from 700hPa to 300hPa. b) divergent meridional wind (m/s) at 850hPa. c) surface pressure (hPa) and divergent wind (m/s). The response (d, e, f) to the idealized cooling centered at 100°E and 7°S. d) vertically integrated omega (Pa · kg/(m²s)) from 700hPa to 300hPa. e) divergent meridional wind (m/s) at 850hPa. f) surface pressure (hPa) and divergent wind (m/s). In c) and f), positive values of the surface pressure are shown in contour (0.5, 1, and 1.5hPa) and negative values are shaded. 5% significant values are only selected and plotted (shading, contours, and vectors) for all panels.
the Walker circulation associated with ENSO, and the negative anomalies imply an additional branch (upward motion) of Walker circulation forced by the western Pacific cooling. This upward motion over the equatorial Indian Ocean is consistent with the convergence at 850hPa (figure 6.1b) as negative and positive meridional wind to the north and south of the equator, respectively. The upward motion over the equatorial Indian Ocean appears with downward motion over India as positive omega, consistent with the pattern of meridional wind at 850hPa over the Indian region and divergence near the surface. This anomalous meridional circulation is referred to a local Hadley circulation. Ashok et al. (2004) suggests that this local Hadley circulation associated with the additional Walker circulation induces a dry monsoon suppressing convection by the anomalous downward motion during developing El Niño. On the other hand, during 1997/98 El Niño the Indian Ocean warming with strong surface convergence does not occur with this local Hadley circulation. In the next section, we will also consider the Indian Ocean warming to simulate a dry monsoon with anomalous 1997 diabatic heating.

In addition, Annamali (2010) and Ashok et al. (2004) found that cooling over the western Pacific shifts west-southward near the Sumatra coast for certain El Niño events (e.g. 2006 event in Annamali (2010)). The shift of the cooling over the western Pacific is similar to our idealized experiment when the cooling is centered at 100°E and 7°S. In their discussion, this shifted cooling is near to the Indian region and this cooling can lead to a local Hadley circulation, which has an upward motion near India. The divergent wind is stronger in figure 6.1e and f than figure 6.1b and c. Ashok et al. (2004) argued that this strong divergent wind induces upward motions near India, seen as negative anomalies of the vertically integrated mid-level omega in figure 6.1d. The upward motion can favor monsoon convection over India. In figure 6.1f, negative anomalous surface pressure in eastern north India can be considered
as the strengthened monsoon trough with strong divergent wind flowing toward that region. For the cooling at 120°E and the equator (figure 6.1c) the negative surface pressure is north of India, but in figure 6.1f when the cooling is centered at 100°E and 7°S, the anomalously strengthened monsoon trough is more clear. This strengthened monsoon trough is different from the common El Niño effect on the Indian monsoon (dry monsoon). So this study also shows that the shift of the downward branch of Walker circulation associated with ENSO can change the influence on the monsoon. Annamali (2010) have done similar experiments and found that the divergent flow from the forced cooling toward India induces moisture convergence over India strengthening the monsoon trough by interaction between equatorial waves and moist processes.

### 6.2 1997 with SOM

For the observed 1997 forcing experiment the 1997 summer mean (May∼Aug) diagnosed diabatic heating from the ERA-40 data is shown in figure 6.2a. The main features are negative diabatic heating anomalies over the equatorial western Pacific and positive diabatic heating anomalies over the equatorial central and eastern Pacific. There are comparably large anomalies over the western North Pacific and Indian Ocean. We first consider only anomalous cooling over the western Pacific and heating over the central/eastern Pacific as forcing in the model with SOM (figure 6.2b). The observed SST of 1997 summer (figure 6.2e) shows predominantly positive anomalies over the central/eastern Pacific.

The response of diabatic heating to the 1997 forcing is shown in figure 6.2c. The response is strong over the western Pacific and central/eastern Pacific, which is considered to be the direct response of the forcing. There are also additional responses
Figure 6.2: a) Diabatic heating anomalies diagnosed from ERA40 during 1997 MJJA. b) forcing (Wm$^{-2}$) of 1997 c) the response to forcing b). d) summation of forcing b) and response c). e) Observed SST during 1997 MJJA ($^\circ$C) f) forcing of 1997 with Indian Ocean heating (IO). g) the response to forcing f). h) summation of forcing f) and response g). 5% significant values only are selected and shaded in c) and g). Diabatic heating in all panels is vertically integrated from the surface to 100hPa (Wm$^{-2}$).
such as anomalous cooling in the eastern Pacific, which is also seen in the observed diabatic heating (figure 6.2a). The forcing and response are combined to form the total diabatic heating in the model. Although the response over the eastern Pacific is very weak, the total diabatic heating (figure 6.2d) has comparable magnitude with the observed diabatic heating (figure 6.2a) because the inserted additional heating has almost the same magnitude with the observed heating. In terms of the Indian monsoon, the 1997 forcing simulates negative anomalies of the diabatic heating (figure 6.2c) and negative precipitation (figure 6.3e) over India. This result is the same with the idealized heating experiments with both WP cooling and CP heating.

The anomalous diabatic heating in figure 6.2a is strong and well developed over the central Pacific but this 1997/98 event brought a normal monsoon instead of a very weak monsoon as expected. This can be seen in the observations, which show no diabatic heating anomalies and no precipitation anomalies over India (figure 6.2a and figure 6.3c). In the model, the 1997 forcing in figure 6.2b does not simulate a normal monsoon, so anomalous heating over the Indian Ocean warming is added for the second experiment of the 1997/98 event (figure 6.2f). A similar 1997 experiment with the Indian Ocean warming has also been carried out by Ashok et al. (2004), Su et al. (2001), and Watanabe and Jin (2003), but in an AGCM with SST forcing. The 1997 forcing with the Indian Ocean in our study (figure 6.2f) does not simulate negative anomalies of diabatic heating over India in figure 6.2g and the response of precipitation does not show significant anomalies over the Indian region in figure 6.3f. This is similar to the observed diabatic heating (figure 6.2a) and anomalous precipitation (figure 6.3c), and implies a normal monsoon. The total diabatic heating over the equatorial Indian Ocean, western Pacific, and central/eastern Pacific in the model (figure 6.2h) also have comparable magnitudes with the observed ones. Therefore, the Indian Ocean warming might be one crucial factor contributing to the Indian
Figure 6.3: Precipitation anomalies (mm/day). Summer mean (May-Aug) of CMAP data in a) 1982, b) 1987, and c) 1997. The response of precipitation in experiments: d) 1982 forcing, e) 1997 forcing, and f) 1997 forcing with Indian Ocean heating (IO).
monsoon. Su et al. (2001) found in their simplified AGCM that warm SST over the Indian Ocean contributes to the local enhanced rainfall more than the strong El Niño warming in the eastern equatorial Pacific.

Since the atmospheric response to tropical forcing has been shown to be important to the Indian monsoon in the experiments of the previous sections, we show how the anomalous atmospheric circulation is changed after adding the Indian Ocean warming (figure 6.4). The 1997 forcing without the Indian Ocean warming simulates a large anti-cyclonic circulation extending toward India and the western Indian Ocean in figure 6.4b. The response of rotational wind shows strong easterlies extending from the equatorial western Pacific across the western Indian Ocean, which weakens the climatological monsoon flow in figure 6.4e. On the other hand, when the Indian Ocean warming is added to the 1997 forcing, the easterlies to the south of India do not extend to the western Indian Ocean (figure 6.4c). The weakened and limited easterlies and anti-cyclonic circulation are similar to the observed stream-function and wind anomalies in figure 6.4a.

In order to further understand the role of the Indian Ocean warming, one idealized heating experiment with SOM from the previous section is shown for 80°E case in figure 6.5e and f. The atmospheric response in figure 6.5f can be compared to the response when the Indian Ocean warming is alone inserted into the model. The two cyclonic circulations to the west of the idealized heating are shown in figure 6.5f. Westerlies to the south of India (6.5f) are strong over the equatorial Indian Ocean, as opposed to the easterlies by the 1997 forcing in figure 6.4b. Therefore, the cyclonic circulation by the Indian Ocean warming can weaken the anti-cyclonic circulation induced by the western Pacific cooling. In terms of the atmospheric response, therefore, it is possible that the Indian Ocean warming plays a role in modifying the influence of ENSO on Indian monsoon.
Figure 6.4: Streamfunction ($\times 10^6$ s$^{-1}$) and rotational wind (m/s) at 850hPa. a) summer mean (May∼Aug) of 1997 anomalies for ERA40. b) the response to 1997 forcing. c) the response to 1997 forcing with Indian Ocean heating (IO). d) summer climatology of ERA40. e) summer climatology of the SOM control run.
Figure 6.5: a) The response of diabatic heating (W m$^{-2}$) for the idealized cooling centered at 120°E with SOM. The contours in a) are the idealized cooling (-50 and -20 W m$^{-2}$). b) The response of streamfunction ($\times 10^6$ s$^{-1}$) and rotational wind (m/s) at 850hPa. c) diabatic heating anomalies diagnosed from the ERA-40 data for 1987 summer mean. d) streamfunction and rotational wind at 850hPa of ERA-40 for 1987 summer mean. e) The response of diabatic heating for the idealized heating centered at 80°E with SOM. The contours in e) are the idealized heating (20 and 50 W m$^{-2}$). f) The response of streamfunction and rotational wind at 850hPa. 5% significant values are only selected and plotted in a), b), e), and f) as the response (shading and vectors). Black dots show the center of the forcing.
6.3 1987 warm event

Another El Niño event (1987) is shown in terms of anomalous diabatic heating (figure 6.5c) and atmospheric circulation (figure 6.5d). The anomalous diabatic heating is negative over the western Pacific and positive over the central Pacific. In figure 6.5d, strong anomalous easterlies appear from the equatorial western Pacific to the western Indian Ocean and a strong anomalous anti-cyclonic circulation resides over all of India and the north Indian Ocean. This is similar to the response to the additional cooling centered at 120°E with SOM (figure 6.5b) and the 1997 forcing without the Indian Ocean warming (figure 6.4b). The strong easterly anomalies over the Indian region are opposite to the monsoon flow and bring negative precipitation over India. This seems to be a common relationship between ENSO and the Indian monsoon associated with anomalous atmospheric circulation.

6.4 1982 with SOM

As the last experiment of this study, the observed 1982 forcing is inserted into the model. The anomalies of the diagnosed diabatic heating during developing 1982 El Niño (Chan and Nigam (2009) data) are shown in figure 6.6e. The strength of the cooling over the western Pacific is similar to the 1997 diabatic heating but anomalous heating is not well developed over the central Pacific as in the 1997/98 event (not shown here over the full central/eastern Pacific). There are also positive anomalies over the Indian Ocean but they do not cover the ocean as extensively as in the 1997/98 event. So the western Pacific cooling and central/eastern Pacific heating are inserted as the 1982 forcing shown in figure 6.6a.

The response to the 1982 diabatic heating shows negative heating anomalies over
Figure 6.6: 

a) Vertically integrated forcing of 1982 diabatic heating anomalies (Wm$^{-2}$). 

b) Vertically integrated diabatic heating response (Wm$^{-2}$). 

c) Response of streamfunction ($\times 10^6$ s$^{-1}$) and rotational wind (m/s) at 850hPa. 

d) Model climatology of streamfunction and rotational wind at 850hPa. In b) and c), only 5% significant values are selected and plotted (shading and vectors). 

e) Diagnosed diabatic heating anomalies during 1982 summer from ERA-40. 

f) Streamfunction and rotational wind anomalies at 850hPa from ERA-40. 

g) ERA-40 climatology of streamfunction and rotational wind at 850hPa.
the Indian region in figure 6.6b. The precipitation response in figure 6.3d and observed rainfall in figure 6.3a also show negative anomalies for 1982 summer. It seems that the anomalous tropical forcing can simulate a reasonable Indian monsoon for 1982. In figure 6.6c, the atmospheric response of the model is strong easterlies opposite to the climatological monsoon flow (figure 6.6d), which is related to the dry monsoon. However, the atmospheric circulation response (figure 6.6c) is too strong compared to the observed atmospheric anomalies (figure 6.6f). The observed atmospheric anomalies in figure 6.6f show easterlies to the south of India but they are weak and even similar to the anomalies of 1997/98 event (figure 6.4a). There is one significant difference between the observed wind in the 1997/98 event and that of the 1982 event: anomalous westerlies over the northern India in 1997 (figure 6.4a) strengthen the monsoon flow; in 1982 event, the anomalous westerlies do not appear over the northern India (figure 6.6f).

In the observed forcing experiments, the western Pacific cooling and central/eastern Pacific heating can simulate a dry monsoon and associated anomalous circulation, as in the 1987 warm event. To simulate a normal monsoon for 1997/98 warm event, the Indian Ocean warming is necessary to compensate for the effect of the western Pacific cooling on the Indian monsoon. The anomalous atmospheric circulation on the large scale seems reasonable to explain two El Niño events: 1987 and 1997/98 cases. On the other hand, the 1982 forcing can simulate a dry monsoon but the atmospheric circulation appears stronger than the observed circulation. Here we suggest three possibilities:

1) This might be due to the importance of small-scale of phenomena related to the monsoon. In addition, the intra-seasonal oscillation of the Indian monsoon may contribute to the summer mean during ENSO events (Goswami and Mohan, 2001; Krishnamurthy and Shukla, 2000) as a residual. It is not clear how well
this intraseasonal oscillation is simulated in the AGCM used here.

2) There is some uncertainty in the diagnosed diabatic heating from ERA-40, which is used as forcing in the experiments.

3) This might be a predictability problem. Nature presents only one realization, and in the experiments of this study there are twenty realizations with the same additional (idealized) forcing. When applying the SOM and adding the CP heating in addition to the WP cooling, the significant area over the India region is small compared to a simpler experiment: idealized forcing near the western Pacific with the prescribed SST. Even the complexity of the experiment reduces the predictability of the forced dynamics over the Indian region. Thus, for a certain year the internal dynamics (variability) may play a larger role in controlling the Indian monsoon than the forced dynamics.
Chapter 7: Conclusions

In this study, to understand the relationship between ENSO and the Indian summer monsoon, the atmospheric response to the tropical diabatic heating is studied in an AGCM. Instead of the traditional approaches, such as inserting SST forcing in an AGCM, or specifying diabatic heating in a simple idealized GCM, we modify an existing GCM by adding a relatively small diabatic heating with idealized structure. In this method, the anomalous circulation by the added idealized heating/cooling will influence the model’s circulation, which in turn further changes the moist and radiative heating fields as the coupled dynamical-convective response (GCM effect).

Each component (western, central, and eastern Pacific) of the diabatic heating is studied for its own influence on the Indian region; this helps to understand the impact of the total tropical diabatic heating. Although the general characteristic of ENSO is an anomalous zonal circulation (Walker circulation), each ENSO event has a different detailed spatial pattern. The method used in this study lets us test the atmospheric sensitivity to diabatic heating at different longitudes. This may help to understand how the atmospheric circulation responds differently to each discrete ENSO event. This idea is similar to Slingo and Annamalia (2000)’s hypothesis: subtle shifts in the domain of influence of ENSO can lead to differing response of the monsoon.

In terms of the atmospheric response, the influence of the tropical diabatic heating on the Indian region becomes different depending on its location. When the idealized heating/cooling is over the warm SST, there is an anchored response over the Bay of Bengal even as the idealized heating moves eastward. The anchored response extends to the Indian region and can interact with the monsoon flow. This anchored response
disappears when the forcing is over cold SST (east of the date line). On the other hand, when the idealized forcing is over the eastern Pacific, there is a Kelvin wave-like response over the Indian region, which propagates eastward. Moreover when the equatorial forcing is moved further south to 7°S but still near the western Pacific, the response of the atmospheric circulation and diabatic heating becomes weak.

7.1 Role of heating/cooling near the Maritime continent

Before considering special and complex El Niño events, the first question is how the descending motion as one branch of the anomalous Walker circulation influences the Indian monsoon for warm events. When the idealized and symmetric cooling with the respect to the equator is inserted near the Maritime continent in the model, anomalous downward motion over the western Pacific is induced as well as Rossby wave response (anticyclonic circulation) to the west of the forcing. The Rossby wave response is associated with winds that oppose the mean monsoon flow and tend to decrease vorticity, which in turn weakens precipitation. This anomalous atmospheric circulation is stronger to the north of the equator (the cooling center) than to the south of the equator. Therefore, although the symmetric idealized cooling does not cover the Indian region, by the interaction between anomalous convection and atmospheric response, the asymmetric response shows that the downward motion branch of the anomalous Walker circulation can influence India. This response seems to explain the relationship between ENSO and the Indian monsoon. This asymmetric response to the tropical heating is also found in Wang et al. (2003) and Xie and Wang (1996): they found that during the northern summer, easterly vertical shear near the Indian region induces stronger atmospheric response to the tropical heating.
The interaction between cumulus convection (diabatic heating in the model) and the large-scale dynamics is also shown in Watanabe and Jin (2003).

Another interaction between convection and equatorial waves is shown with convergence near the surface when the idealized heating is inserted over the warm SST. With the development of the Gill type circulation the low in surface pressure spreads to the north and south of the forcing region and the surface convergence consistent with the surface pressure can induce moisture convergence. The response of the diabatic heating is also meridionally extended and centered at the north of the forcing region so the additional heating response can be explained by the surface moisture convergence. In Wu et al. (2006)'s symmetric (to the equator) forcing experiment in AGCM, the northward shift precipitation is induced with northward propagation of surface convergence. Therefore, the northward shift in the local response of the diabatic heating might be related to the boundary layer and/or lower tropospheric physical process.

7.2 Role of ocean-atmosphere coupling

A negative feedback between atmosphere and ocean is suggested in the regions of large mean rainfall (Wu and Kirtman, 2005). Coupling the atmosphere and ocean in the GCM can generate more realistic precipitation in the Indian region (Fu and Wang, 2002; Wu and Kirtman, 2005). In the SOM control of this study, the Indian summer rainfall is reduced compared with the non-SOM control run (prescribed SST over the entire ocean). When the forcing is near the Indian region, the idealized heating experiments with SOM coupling over the Indian Ocean and western Pacific also show the negative feedback compared to the heating experiments with only prescribed SST.

The anomalous wind response to the heating in the Arabian Sea induces latent
heat flux from the ocean to the air. While the release of latent heat flux leads to a cooling of the ocean (and hence less subsequent heat flux) in the SOM experiments, the ocean with prescribed SST continues to produce a latent heat flux and in turn the diabatic heating. So the anomalously cool SST in the heating experiments with SOM weakens the positive response of the diabatic heating. On the other hand, the idealized cooling experiments show a positive feedback of the diabatic heating response. The latent heat flux related to the anomalous wind over the western Pacific also releases energy from the ocean to the atmosphere, which decreases the SST. This anomalously cold SST enhances the negative response of the diabatic heating and in this region this constitutes a positive feedback. Thus, in the SOM experiments, the response to the additional idealized cooling is stronger than the response to the additional heating due to different feedbacks. The positive feedback is also emphasized for the reason of the maintaining anomalous circulation during the retreating phase of El Niño (Wang et al., 2000, 2003).

7.3 Role of heating over central/eastern Pacific

In a real El Niño event, the main characteristics are anomalously warm SST over the tropical central/eastern Pacific in addition to the anomalous decrease of convection over the Maritime continent. So we ask the following questions: does the anomalous diabatic heating at a distance (over the central/eastern Pacific) influence the Indian monsoon? Do the anomalous diabatic cooling in the Maritime continent and the diabatic heating over the central/eastern Pacific interact with each other to impact India?

The central/eastern Pacific heating itself (still idealized but with realistic longitudinal extent) induces anomalously negative diabatic heating over the equatorial Indian
ocean and western pacific by compensating vertical motion (downward motion), thus
enhancing the western pacific cooling. Idealized heating over the central/eastern pa-
cific also induces a westward propagating rossby wave, while at the same time the
kelvin response to the east of the heating can propagate eastward all the way to the
Indian region. The atmospheric response near the Indian region can interact with
the monsoon low influencing the Indian rainfall. In the experiment using the western
pacific cooling and the central/eastern pacific heating, the prescribed SST exagger-
ates the diabatic heating response to the south of Bay of Bengal and western north
pacific due to lack of air-sea feedback, so that the relative influence of the western
pacific cooling on India becomes weak. But by applying the SOM this exaggerated
response disappears and the local response related to India becomes clear.

7.4 Simulated observed warm ENSO events

Based on the all idealized heating experiments, the observed diabatic heating is
mimicked for the 1997/98 and 1982 warm ENSO events in the model. To simulate
the observed normal monsoon for the 1997/98 warm ENSO event, the Indian Ocean
is necessary to compensate for the effect of the western pacific cooling on the Indian
monsoon. The anomalous atmospheric circulation in the large scale seems reasonable
to explain the relationship between El Niño and the Indian monsoon. Therefore, in
this study, the large atmospheric circulation can be considered as one of main factors
to control the Indian monsoon. However, for the 1982 warm event the simulated
atmospheric circulation is too large compared with the observed circulation. Three
possibilities are suggested: 1) difficulties in the reasonable simulation of the intra-
seasonal oscillation in the Indian region, 2) uncertainty in diagnosed diabatic heating,
which used as forcing in the experiments, 3) large influence of internal dynamics
compared to the forced dynamics.

In summary, the additional diabatic heating inserted in the AGCM generates anomalous atmospheric circulations (mainly equatorial waves), and leads to further interactions of the circulation with convection. In this way, anomalous downward motion of the Walker circulation over the Maritime continent can induce a weak Indian monsoon. Adding the central/eastern Pacific in addition to the western Pacific cooling, a diabatic heating response is generated which is similar to the diagnosed heating from the observation data. In addition, the response of the diabatic heating and the atmospheric circulation influencing India is sensitive to the longitudinal and latitudinal locations of the forcing. Therefore, the spatial pattern and locations of the tropical diabatic heating is important to understand the large-scale atmospheric circulation associated with the Indian monsoon. However, since 1982 warm event cannot be clearly explained by this large-scale atmospheric circulation, other factors should be further studied.
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Curriculum Vitae

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