Evaluating and understanding the role of convective processes in general circulation model simulations of the Madden-Julian Oscillation

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Evaluating and understanding the role of convective processes in general circulation model simulations of the Madden-Julian Oscillation

by

Liping Deng

A dissertation submitted to the graduate faculty
in partial fulfillment of the requirements for the degree of

DOCTOR OF PHILOSOPHY

Major: Meteorology

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2010

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ABSTRACT

Weak temporal variability in tropical climate such as the Madden-Julian Oscillation (MJO) is one of the outstanding deficiencies in numerical weather prediction (NWP) models and coupled general circulation models (GCMs), which have beleaguered the model developer and user communities for years. The representation of convection, cloud and radiation processes has long been recognized as one of the major problems responsible for these deficiencies in climate models. Recently, with the improvement made to the convection scheme, the Iowa State University (ISU) GCM, which is based on the NCAR Community Climate Model version 3 (CCM3), is able to simulate many features of the MJO as revealed by the observational studies. It provides a unique opportunity to investigating the mechanisms and physical processes through which convection affects the MJO. In this study, four 10-year (1979-88) simulations by the ISUGCM with observed sea surface temperatures (SSTs) are used for the analysis. The control simulation (CTL) is conducted with the standard NCAR CCM3. The ISUGCM simulation with the revised closure, convection trigger condition and convective momentum transport (CMT) is labeled by ISUCCM3. Two sensitivity simulations, NOCMT and NOTRI, are performed to investigate the impact of convection trigger condition and CMT on the MJO.

The modifications made in convection schemes improve the simulations of MJO in the amplitude, spatial distribution, eastward propagation, and horizontal and vertical structures, especially for the coherent feature of the MJO-related eastward propagating convection and the precursor sign of the convective center. The revised convection closure plays a key role
in the improvement of eastward propagation of MJO. The convection trigger helps produce less frequent but more vigorous moist convection and enhance the amplitude of the MJO signal. The inclusion of CMT results in more coherent structure for the MJO-related deep convective center and its corresponding atmospheric variances.

The kinetic energy budget is conducted to analyze the physical processes responsible for the improved MJO simulated by ISUGCM with the modified convection scheme. The increased equatorial (10°S-10°N) MJO-related perturbation kinetic energy (PKE) is presented in the upper troposphere due to the three modifications, and leads to more robust and coherent eastward propagating MJO signal. In the MJO source region-the Indian Ocean (45°E-120°E), the upper-tropospheric MJO PKE is maintained by the convergence of vertical wave energy flux and the barotropic conversion through the horizontal shear of mean flow. In the convectively active region-the western Pacific (120°E-180°), the upper-tropospheric MJO PKE is supported by the convergence of horizontal and vertical wave energy fluxes. Over the central-eastern Pacific (180°-120°W), where convection is suppressed, the upper-tropospheric MJO PKE is mainly due to the convergence of horizontal wave energy flux. The deep convection trigger condition produces stronger convective heating which enhances the upward wave energy fluxes, and leads to the increased MJO PKE over the Indian Ocean and western Pacific. The revised convection closure affects the response of mean zonal wind shear to the convective heating over the Indian Ocean and leads to the enhanced upper-tropospheric MJO PKE through the barotropic conversion. The stronger eastward wave energy flux due to the increase of convective heating over the Indian Ocean and western Pacific by the revised closure is favorable to the eastward propagation of MJO and the
convergence of horizontal wave energy flux over the central-eastern pacific. The convection-induced momentum tendency tends to decelerate the upper-tropospheric wind which results in a negative work to the PKE budget in the upper troposphere. However, the convection momentum tendency accelerates the westerly wind below 850 hPa over the western Pacific, which is partially responsible for the improved MJO simulation.

With the composite analyses of MJO events, the different phase relationships between MJO 850-hPa zonal wind, precipitation and surface latent heat flux are simulated over the Indian Ocean and western Pacific in ISUGCM, which is greatly influenced by the convection closure, trigger and CMT. With these modifications in the convection scheme, ISUGCM produces better MJO recharge-discharge process of moist static energy than the original GCM. The convection trigger condition for deep convection contributes to the striking difference between ISUCCM3 and CTL and plays the major role for this improved MJO simulation through the horizontal and vertical advections of moist static energy. The inclusion of the revised closure helps build up the precondition of the MJO convection with the redistribution of moisture through the positive contributions of the horizontal and vertical advection of moist static energy before the onset of MJO convection. The impact of CMT through the interaction between horizontal and vertical advection of moist static energy helps the more coherent atmospheric structure over the Indian Ocean and western Pacific. The budget analysis for ISUGCM with the modifications shows the increase of moist static energy is in phase with the horizontal advection of moist static energy over the western Pacific, but in phase with the vertical advection of moist static energy over the Indian Ocean.
CHAPTER 1. GENERAL INTRODUCTION

1.1 Introduction

Several outstanding deficiencies in numerical weather prediction (NWP) models and coupled general circulation models (GCMs) have beleaguered the model developer and user communities for years. Among these deficiencies are weak temporal variability in tropical climate such as the Madden-Julian Oscillation (MJO); double Inter-Tropical Convergence Zone (ITCZ) belts in the central and eastern Pacific, where a spurious precipitation band exists south of the equator, particularly in the Northern Hemisphere summer months; and a too short and too regular El Niño-Southern Oscillation (ENSO) periodicity, every two to three years, compared to the observed periodicity of three to eight years. The representation of convection, cloud and radiation processes has long been recognized as one of the major problems responsible for these deficiencies in climate models (GEWEX Cloud System Science Team 1994). The organization and evolution of tropical convection is a major component of both MJO and ENSO. Convection affects large-scale circulation and wave disturbances through latent heat release; and the redistribution of heat, moisture and momentum. The coupling of convection, cloud and radiation processes with large-scale dynamics is crucial for modeling the global precipitation distribution, ITCZ, MJO and ENSO (e.g., Miller et al. 1992; Slingo et al. 1994; Zhang et al. 1998; Maloney and Hartmann 2001; Wu et al. 2003; Wu and Liang 2005a, b; Zhang and Mu 2005a, b; Zhang and Wang 2006; Wu et al. 2007a, b).
After the discovery of the MJO by Madden and Julian (1971), the characteristics and structure of the MJO have been extensively studied and well documented by many observational papers and reviews (Madden and Julian 1994; Lau and Waliser 2005; Zhang 2005). The MJO is a dominant intraseasonal variability in the tropical atmosphere. The MJO convection develops and propagates eastward along the equator in the Indian Ocean, tends to propagate into the South Pacific Convergence Zone (SPCZ) in the western Pacific, and decays in the central Pacific (Madden and Julian 1972). Convectively coupled MJO signals have been detected in various properties associated with deep convection and atmospheric circulations such as the outgoing longwave radiation (OLR), surface precipitation, 200-hPa velocity potential and zonal wind, 850-hPa zonal wind, surface latent heat flux (e.g., Zangvil 1975, Zangvil and Yanai 1981; Krishnamurti and Subrahmanyam 1982; Weickmann et al. 1985; Chen and Yen 1991; Lin and Johnson 1996; Slingo et al. 1996; Chen and Chen 1997; Maloney and Hartmann 2001; Sperber 2004; Zhang and Dong 2004). The major power of the MJO is concentrated at wavenumber 1 and eastward periods of 30-90 days (Hayashi 1979; Salby and Hendon 1994). The speed of eastward propagating MJO is about 5 m s\(^{-1}\) (Weickmann et al 1985; Knutson et al.1986). Within the MJO, a hierarchical structure of cloud systems with a horizontal scale of several hundred kilometers is identified by Nakazawa (1988), which includes several eastward-moving super cloud clusters (SCCs) near the equator over the western Pacific and several westward-moving cloud clusters within each SCC. The horizontal structure of the MJO shows the coupling of deep convection with the large-scale motion; a pair of upper-level (200 hPa) anticyclonic circulations (easterlies in between) and a pair of low-level (850 hPa) cyclonic circulations (westerlies in between) on both side of equator over the minimum outgoing longwave radiation (OLR) in the Indian
ocean and western Pacific (e.g., Weickmann 1983; Rui and Wang 1990; Hendon and Salby 1994; Yanai et al. 2000; Kiladis et al. 2005). The MJO temperature, specific humidity, wind, divergence, and diabatic heating fields display asymmetry and westward tilt in the vertical (Sperber 2003; Lin et al. 2004; Kiladis et al. 2005). Low-level convergence, upward motion and positive moisture anomaly favorable for the development of new convection and the eastward propagation are present on the east side of the MJO convective center, while low-level divergence, downward motion and negative moisture anomaly exist on the west side.

To explain the origin and observed features of the MJO, two types of theories have been proposed by scientists, i.e., (1) the MJO as an atmospheric response to forcing sources and (2) the atmospheric instability-induced MJO through the interaction of convection with large-scale wave motion. The forcing used in the first type of theory includes the tropical localized thermal forcing and stochastic forcing. The tropical response to the stationary intraseasonal oscillating heat source or the randomly varying heating profiles, such as the convective activity associated with the Asian monsoon and the convective disturbance within the ITCZ, produces some observed features of the MJO (e.g., Yamagata and Hayashi 1984; Anderson and Stevens 1987; Salby and Garcia 1987). But the theory does not provide the mechanism responsible for the origin of low oscillation frequency and eastward movement of the heating source. Hu and Randall (1994, 1995) suggested that the low frequency localized convective heat source is due to the nonlinear interactions among radiation, cumulus convection, and surface moisture fluxes. The radiative cooling and surface evaporation act to destabilize the atmosphere by increasing the lapse rate and the mixed layer relative humidity, respectively, while convection tends to counteracts the radiative and surface processes by reducing the
lapse rate and boundary layer moisture. The shifts between the radiation-convection interaction and the surface moisture flux-convection interaction lead to the low frequency thermal forcing. Blade and Hartmann (1993) introduced a discharge-recharge mechanism to determine the period of the low frequency oscillation of convective heating by the discharge time of convective stabilization together with the recharge time of moist static instability. Grabowski (2003) used the convection-moisture feedback to explain the MJO-like coherent structure in the precipitation, precipitable water and zonal wind. Moncrieff (2004) suggested a nonlinear mechanism of MJO propagation that analytically includes the pivotal role of mesoscale convective organization on the large-scale coherence of tropical convection.

Various mechanisms, including the wave-CISK (Conditional Instability of the Second Kind) and the wind-evaporation feedback, are considered in the second type of theory. The wave-CISK mechanism proposed by Lau and Peng (1987) involves the cooperative feedback between positive-only convective heating and low-level moisture convergence. The moist Kelvin wave is selectively amplified by the condensational latent heat release that feeds on the east-west asymmetry of equatorial waves, and in turn forces the heat source propagating eastward. The unstable modes resemble some observed features of the MJO, such as the eastward propagation and westward tilt in vertical, but have a phase speed of about 19 m s\(^{-1}\) which is larger than the observed speed. Follow-on studies have shown that the vertical distribution of convective heating, the use of convection parameterization, and the friction-induced boundary layer moisture convergence through the impact on the convective heating play an important role in obtaining the slower eastward propagating and planetary-scale unstable modes of moist Kelvin waves (e.g., Takahashi 1987; Wang 1988; Sui and Lau 1989;
Neelin and Yu 1994; Chao 1995; Cho and Pendlebury 1997). The wind-evaporation mechanism proposed by Emanuel (1987) and Neelin et al. (1987) makes use of the instability due to the feedback between zonal wind perturbations and evaporation. It assumes that the mean surface wind is easterly. An enhanced latent heating will force anomalous easterlies to the east of the region and anomalous westerlies to the west at low levels, and result in increased surface zonal winds and evaporation to the east and decreased winds and evaporation to the west. The enhanced (reduced) evaporation in turn will feed back onto the strengthened (weakened) latent heating to the east (west), which produces eastward propagating wave modes. Because of the requirement of the existence of mean surface easterlies, the mechanism is not able to explain the MJO over the Indian Ocean and western Pacific where the mean surface winds are westerlies.

Despite the progress in the understanding of the MJO by the observational and theoretical studies, the simulation of the MJO remains a major challenge for GCMs and NWP models. The unrealistic features in the MJO simulations include weak amplitude, more power at higher frequencies than in observations, temporal and spatial distributions of MJO variances differing from those observed, eastward propagation speed being too fast, and lacking of coherent structure for the eastward propagation from Indian Ocean to the Pacific (e.g., Slingo et al. 1996; Wheeler and Kiladis 1999; Inness and Slingo 2003; Sperber 2004; Sperber et al. 2005; Zhang 2005). While some improvement in simulating MJO variance and coherent eastward propagation has been attributed to model resolutions (e.g., Hayashi and Golder 1986; Kuma 1994; Inness et al. 2001; Sperber et al. 2005), model mean background state (e.g., Slingo et al. 1996; Hendon 2000; Inness et al. 2003; Sperber et al. 2005), and air-sea
interaction (e.g., Waliser et al. 1999; Inness and Slingo 2003; Sperber 2004; Sperber et al. 2005), studies have shown that the model physics, especially the representation of convective processes, may be the key to producing the realistic MJO simulations in GCMs. Tokioka et al. (1988) showed that the addition of a minimum value of cumulus entrainment rate of the environmental air in the Arakawa and Schubert (1974) convection scheme is a key factor for simulating the MJO by a GCM. Slingo et al. (1996) evaluated 15 atmospheric GCMs and found that the convection schemes with the convective available potential energy (CAPE) type closure tend to produce better MJO signals than the moisture convergence type closure. Wang and Schlesinger (1999) showed that the use of large threshold of relative humidity allows the accumulation of moist static energy to a certain amount to trigger the convection in three different convection schemes and improves the simulation of MJO signals. Maloney and Hartmann (2001) improved the MJO simulations by using the microphysics of cloud together with the relaxed Arakawa-Schubert convection scheme (Sud and Walker 1999) and suggested that the simulations are sensitive to the parameterization of convective precipitation evaporation in unsaturated environment air and unsaturated downdrafts. Liu et al. (2005) showed that a GCM with the Tiedtke (1989) convection scheme simulates an improved mean state, intraseasonal variability, space–time power spectra, and coherent eastward propagation of MJO-related precipitation. Lin et al. (2006) evaluated 14 coupled GCM simulations of MJO and found that the intraseasonal variance is too weak in most of the models but two models that have convective closure/triggers tied to the moisture convergence produce better MJO simulations than others.
While the above studies are encouraging, the factors in the convection schemes identified to be crucial for improving MJO simulations are not universal for GCMs. Therefore, understanding the physical mechanism responsible for the MJO simulations remains a major challenge. Physically sound and observationally validated convection schemes are needed for undertaking this task. Recently, Zhang (2002) improved the closure assumption of Zhang and McFarlane (1995) convection scheme by relating convection to the destabilization of the tropospheric layer above the planetary boundary layer by the large-scale processes based on the observations from the Atmospheric Radiation Measurement (ARM) and the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA-COARE). Zhang and Mu (2005) showed that the revised closure assumption helps the MJO simulations by the NCAR GCM. This scheme has been installed in the Iowa State University (ISU) GCM which is based on a version of NCAR Community Climate System Model. Two major improvements have also been implemented in the ISUGCM, i.e., the trigger condition of deep convection and the convective momentum transport parameterization. The trigger condition for deep convection is based on the cloud-resolving simulations, i.e., the convection is activated when the CAPE value increase due to the large-scale forcing exceeds a certain threshold (Wu et al. 2007). The convective momentum transport (CMT) parameterization scheme is validated and simplified using cloud-resolving simulations (Zhang and Cho 1991; Wu et al. 2003; Zhang and Wu 2003). The CMT scheme considers the vertical redistribution of the horizontal momentum by convection, and accounts for the role of perturbation pressure field generated by the interaction of convection with large-scale circulation in vertical momentum transport (Wu and Yanai 1994). The interaction between CMT and the thermodynamic effects of convection plays an important role in shaping
tropical convection and the Hadley circulation. Uncoupled and coupled ISUGCM simulations demonstrate great impacts of improved convection, cloud and radiation schemes on many aspects of climate simulations such as the global distribution, diurnal cycle, frequency, and seasonal variation of precipitation; MJO; ENSO; and the evolution of 1997/98 El Niño-type event (e.g., Wu et al. 2003; Wu and Liang 2005a, b; Wu et al. 2007b). The underlying hypothesis is that moist convection occurs less frequently but better organized with the improved scheme in which convection is controlled by the change of CAPE due to the large-scale advection and is activated when the increase of CAPE reaches a certain threshold. The coupling of CMT with the convective heating and moistening affects the large-scale wind, tropical energy budget, and MJO.

1.2 Research Questions

The objectives of this study are to document the impacts of the modified convection scheme on the MJO simulations, and to investigate the mechanisms and physical processes through which convection affects the MJO. The questions we want to ask are the following:

1) What is the improvement of MJO simulations in ISUGCM with the modified convection scheme?

2) What is the impact of each modification in the convection scheme on the improved MJO simulation?

3) What is the maintenance of MJO-related perturbation kinetic energy budget over the Indian Ocean and Pacific? What are the underlying mechanisms and physical processes?
4) What are the impacts of revised convection closure, convection trigger and CMT in the convection scheme on the mechanisms and physical processes responsible for the change of the MJO-related perturbation kinetic energy?

5) What are the mechanisms and physical processes affecting the MJO simulation through the study of moist static energy (MSE) budget over the Indian Ocean and western Pacific?

6) What are the impacts of revised convection closure, convection trigger and CMT on the MSE budgets for the composite MJO?

1.3 Dissertation Organization

Chapter 1 in the thesis includes the general introduction, and Chapter 2 contains the data sources. Chapter 3 is a paper published in the Journal of Climate, and it gives the answer to the research questions 1 and 2 about the improvements of MJO simulations in ISUGCM and the impacts of three modifications. Chapter 4 is the second paper submitted to the Journal of Climate, and it provides answer for the research questions 3-4 through the analysis of tropical MJO kinetic energy budget. The mechanisms and physical processes that contribute to the perturbation kinetic energy associated with the 20-100 days variance over the Indian Ocean and Pacific are illustrated, and the impacts of three modifications on the mechanisms and physical processes are discussed. Chapter 5 gives the answer to the research questions 5 and 6 through the composite analyses of the moist static energy (MSE) budget over the Indian Ocean and western Pacific for MJO simulations in ISUGCM. Chapter 6 gives the general conclusions and future work.
CHAPTER 2. DATA SOURCES

2.1 ISUGCM and Modifications

The ISUGCM is based on the NCAR Community Climate Model version 3 (CCM3), which is a spectral global climate model (Kiehl et al. 1998). It has been used worldwide for climate modeling and climate change studies. The resolution is user-specifiable, with the most common implementation having 18 hybrid vertical levels extending from the surface to 4 mb, and a horizontal resolution of T42 (a roughly 2.8° × 2.8° Gaussian grid). It has a highly sophisticated physical parameterization package for subgrid scale processes such as boundary layer turbulence, radiation, clouds and convection.

In ISUGCM, the shallow convection and deep precipitating convection are treated by two different schemes, i.e., Hack (1994) and modified Zhang and McFarlane (1995, 2002, 2003) schemes. The Hack shallow convection scheme is a modified moist convective adjustment scheme in mass flux form. The scheme checks the local instability of the temperature stratification, as opposed to the deep instability used in the deep convection scheme, and adjusts the three adjacent model layers where local instability is present to a neutral state. The original Zhang and McFarlane deep convection scheme makes usage of the ensemble plume concept to represent convective clouds (similar to the Arakawa and Schubert 1974) and uses the convective available potential energy (CAPE) as closure to decide the amount of convection (Zhang and McFarlane 1995). It assumed that convective-scale updrafts and saturated downdrafts may exist when the atmosphere is locally conditionally unstable in the lower troposphere, and the convection occurs only when there is CAPE and acts to remove
the CPAE at an exponential rate with a specified adjustment time scale. The shortcoming for this scheme is the weak temporal variability due to the fact that convection occurs almost all the time. With the Atmospheric Radiation Measurement (ARM) and the Tropical Ocean Global Atmosphere Coupled Ocean Atmosphere Response Experiment (TOGA-COARE) datasets, Zhang (2002, 2003) found that the convection is more related to the large-scale advective generation of CAPE in the tropospheric layer above the planetary boundary layer instead of to CAPE itself. With these results, Zhang (2002) modified his CAPE-based closure to a new closure following the quasi-equilibrium between convection and environment large-scale processes in the free troposphere, which showed strong impacts on the climate simulations of CCM3 (Zhang and Mu 2005b).

Besides the revised closure assumption of the Zhang and McFarlane scheme, two major improvements have also been implemented in the ISU GCM, i.e., the trigger condition of deep convection and the convective momentum transport parameterization. The trigger condition for deep convection is obtained from the year-2000 Cloud-Resolving Model (CRM) simulation over ARM Southern Great Plains (SGP). The assumption is that the deep convection is only activated when the CAPE value continuously increases due to the large-scale forcing and exceeds a certain threshold (70 J kg$^{-1}$ hr$^{-1}$; Wu et al. 2007b). The convective clouds transport not only heat and moisture, but also momentum. The estimate of the convective momentum transport (CMT) through the momentum budget is a difficult task. Many attempts try to parameterize the CMT and incorporate it in models. With the observational studies demonstrated that the convection-induced pressure perturbation can strongly affect the in-cloud momentum and the CMT (e.g., LeMone 1983; LeMone et al.
1984), Zhang and Cho (1991) develop a scheme for CMT parameterization including the effects of cloud-scale pressure gradient. Wu and Yanai (1994) also represent the convective pressure gradient forces in terms of large-scale vertical wind shear, cloud mass flux, and the scale of convective clouds in their CMT parameterization. The CMT affects the environment flow through the subsidence of environment air that compensates the cloud mass flux, the detrainment of momentum from clouds, and the convection-induced pressure gradient force. In ISUGCM, the CMT scheme considers the vertical redistribution of horizontal momentum by convection, and accounts for the role of the perturbation pressure field generated by the interaction of convection with large-scale circulation in vertical momentum transport (Wu and Yanai 1994; Wu et al. 2003, 2007a; Zhang and Wu 2003).

2.2 Four Model Simulations

Four 10-year (1979-1988) simulations, with observed sea surface temperatures (SSTs), are used for the analysis (Table 2.1). The control simulation (CTL) is conducted with the standard CCM3. The simulation with the revised closure, convection trigger condition and CMT is labeled by ISUCCM3. Two sensitivity simulations, NOCMT and NOTRI, are performed to investigate the impact of convection trigger condition and convective momentum transport on the MJO.

<table>
<thead>
<tr>
<th>Table 2.1 List of all model simulations</th>
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<tr>
<td>CTL</td>
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<tr>
<td>ISUCCM3</td>
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<tr>
<td>NOCMT</td>
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<tr>
<td>NOTRI</td>
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</table>
2.3 Observation Datasets

The pentad (5-day mean) precipitation product of the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) is used to evaluate the climate simulations. Observations from rain gauges and precipitation estimates from several satellite-based algorithms (infrared and microwave) are merged by the technique described in Xie and Arkin (1996, 1997). The Advanced Very High Resolution Radiometer (AVHRR) Outgoing Longwave Radiation (OLR) data is interpolated in time and space from National Oceanic and Atmospheric Administration (NOAA) twice-daily OLR values and averaged to once daily (Liebmann and Smith 1996). The Atmospheric Infrared Sounder (AIRS) Version 5 Level 3 Standard Gridded daily products are used during the period 2002-2008. The AIRS instrument suite is designed to measure the water vapor and temperature profiles on a global scale based on the Aqua mission. The National Centers for Environmental Prediction (NCEP) reanalysis data (Kalnay et al. 1996), including the wind, OLR, precipitation, vertical velocity, latent heat flux, and specific humidity, is used in this study. The European Centre for Medium-Range Weather Forecasts (ECMWF) Interim re-analysis (ERAI) dataset is also used to describe the MJO. The ERAI includes new humidity analysis, better formulation of background error constraint, improved model physics and bias handling compared to the original ERA reanalysis. These datasets are regridded to T42 for comparing with the model output.
CHAPTER 3. EFFECTS OF CONVECTIVE PROCESSES ON GCM SIMULATIONS OF THE MADDEN-JULIAN OSCILLATION

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Abstract

Weak temporal variability in tropical climates such as the Madden-Julian Oscillation (MJO) is one of major deficiencies in general circulation models (GCMs). The uncertainties in the representation of convection and cloud processes are responsible for these deficiencies. With the improvement made to the convection scheme, the Iowa State University (ISU) GCM, which is based on a version of the NCAR Community Climate Model, is able to simulate many features of MJO as revealed by observations. In this study, four 10-year (1979-88) simulations with observed sea surface temperatures by ISU GCM are analyzed and compared to examine effects of the revised convection closure, convection trigger condition and convective momentum transport (CMT) on the MJO simulations. The modifications made in convection schemes improve the simulations of MJO in the amplitude, spatial distribution, eastward propagation, and horizontal and vertical structures, especially for the coherent feature of the MJO-related eastward propagating convection and the precursor sign of the convective center. The revised convection closure plays a key role in the improvement of
eastward propagation of MJO. The convection trigger helps produce less frequent but more vigorous moist convection and enhance the amplitude of the MJO signal. The inclusion of CMT results in more coherent structure for the MJO-related deep convective center and its corresponding atmospheric variances.

1. Introduction

After the discovery of the MJO by Madden and Julian (1971, 1972), its characteristics and structure have been extensively analyzed by many observational studies (e.g., Madden and Julian 1994; Lau and Waliser 2005; Zhang 2005). The MJO is a dominant intraseasonal variability in the tropical atmosphere. The MJO related convection develops and propagates eastward along the equator in the Indian Ocean, tends to propagate into the South Pacific Convergence Zone (SPCZ) in the western Pacific, and decays in the central Pacific (Madden and Julian 1972). Convectively coupled MJO signals have been detected in various properties associated with deep convection and atmospheric circulations such as the outgoing longwave radiation (OLR), surface precipitation, 200-hPa velocity potential and zonal wind, 850-hpa zonal wind, and surface latent heat flux (e.g., Zangvil 1975, Zangvil and Yanai 1981; Krishnamurti and Subrahmanyam 1982; Weickmann et al. 1985; Chen and Yen 1991; Slingo et al. 1996; Chen and Chen 1997; Maloney and Hartmann 2001; Sperber 2004; Zhang and Dong 2004). The major power of the MJO is concentrated at wavenumbers 1-3 and eastward periods of 30-90 days (e.g., Salby and Hendon 1994; Zhang 2005). The speed of eastward propagating MJO is about 5 m s$^{-1}$ (Weickmann et al. 1985; Knutson et al. 1986). Within the MJO, a hierarchical structure of cloud systems is identified by Nakazawa (1988), which includes several eastward-moving super cloud clusters (SCCs) near the equator over
the western Pacific and several westward-moving cloud clusters within each SCC. The horizontal structure of the MJO shows the coupling of deep convection with the large-scale motion; a pair of upper-level (200 hPa) anticyclonic circulations (easterlies in between) and a pair of low-level (850 hPa) cyclonic circulations (westerlies in between) on both side of the equator in the Indian ocean and western Pacific with the minimum OLR (e.g., Weickmann 1983; Rui and Wang 1990; Hendon and Salby 1994; Yanai et al. 2000; Kiladis et al. 2005). The MJO temperature, specific humidity, wind, divergence, and diabatic heating fields display asymmetry and westward tilt in the vertical (Sperber 2003; Lin et al. 2004; Kiladis et al. 2005). Low-level convergence, upward motion and positive moisture anomaly favor the development of new convection and the eastward propagation are present on the east side of the MJO convective center, while low-level divergence, downward motion and negative moisture anomaly exist on the west side.

Despite progress in the understanding of the MJO by the observational study, the MJO simulation remains a major challenge for GCMs and NWP models. The unrealistic features in the MJO simulations include weak amplitude, more power at higher frequencies than in observations, temporal and spatial distributions of MJO variances differing from those observed, eastward propagation speed being too fast, and lacking of coherent structure for the eastward propagation from Indian Ocean to the Pacific (e.g., Slingo et al. 1996; Wheeler and Kiladis 1999; Inness and Slingo 2003; Sperber 2004; Sperber et al. 2005; Zhang 2005). While some improvement in simulating MJO variance and coherent eastward propagation has been attributed to model resolution (e.g., Kuma 1994; Inness et al. 2001; Sperber et al. 2005), model mean background state (e.g., Slingo et al. 1996; Inness et al. 2003; Sperber et
al. 2005), and air-sea interaction (e.g., Waliser et al. 1999; Inness and Slingo 2003; Sperber 2004; Sperber et al. 2005), studies have shown that the model physics especially the representation of convective processes may be the key to producing the realistic MJO simulations in GCMs. Convection affects large-scale circulation and wave disturbances through precipitation, latent heat release, and the redistribution of heat, moisture and momentum. Since the organization and evolution of tropical convection is a major component of the MJO, the coupling of convection scheme with large-scale dynamics is crucial for modeling the MJO.

Tokioka et al. (1988) showed that the addition of a minimum value of cumulus entrainment rate of environmental air in the Arakawa and Schubert (1974) convection scheme is a key factor for simulating the MJO by a GCM. Slingo et al. (1996) evaluated 15 atmospheric GCMs and found that the convection schemes with the convective available potential energy (CAPE) type closure tend to produce better MJO signals than the moisture convergence type closure. Wang and Schlesinger (1999) showed that the use of large threshold of relative humidity allows the accumulation of moist static energy to a certain amount to trigger the convection in three different convection schemes and improves the simulation of MJO signals. Maloney and Hartmann (2001) improved the MJO simulations by using the microphysics of cloud together with the relaxed Arakawa-Schubert convection scheme (Sud and Walker 1999) and suggested that the simulations are sensitive to the parameterization of convective precipitation evaporation in unsaturated environment air and unsaturated downdrafts. Liu et al. (2005) showed that a GCM with the Tiedtke (1989) convection scheme simulates an improved mean state, intraseasonal variability, space–time power spectra, and
coherent eastward propagation of MJO-related precipitation. Lin et al. (2006) evaluated 14 coupled GCM simulations of MJO and found that the intraseasonal variance is too weak in most of the models but two models that have convective closure/triggers tied to the moisture convergence produce better MJO simulations than others.

The above studies indicate that the factors in the convection schemes identified to be crucial for improving MJO simulations are not universal for GCMs. Therefore, understanding the physical mechanism responsible for the MJO simulations remains a major challenge. Physically sound and observationally validated convection schemes are needed for undertaking this task. Recently, Zhang (2002) improved the closure assumption of Zhang and McFarlane (1995) convection scheme by relating convection to the destabilization of the tropospheric layer above the planetary boundary layer by the large-scale processes based on the observations from the ARM and TOGA-COARE. Zhang and Mu (2005b) showed that the revised closure assumption helps the MJO simulations by the NCAR GCM. Uncoupled and coupled simulations using the Iowa State University (ISU) GCM with various modifications to the convection, cloud and radiation schemes demonstrate improvements on many aspects of global climate simulations such as the ENSO, the MJO, and the precipitation and energy budget (e.g., Wu et al. 2003; Wu and Liang 2005a, b; Wu et al. 2007a, b).

The ultimate goal of this project is to understand the mechanisms and physical processes through which convection affects the MJO. In this paper, the objectives are to evaluate the MJO simulated by ISUGCM against observations, and to examine the impacts of revised convection closure, convection trigger and CMT on the simulations. The data and analysis
techniques are described in section 2. The characteristics and structure of the simulated MJO are presented in section 3. The summary is given in section 4.

2. ISUGCM simulations, observational datasets and analysis techniques

ISUGCM is based on the NCAR Community Climate Model version 3 (CCM3), which is a spectral global climate model (Kiehl et al. 1998). It has been used worldwide for climate modeling and climate change studies. The resolution is user-specifiable, with the most common implementation having 18 hybrid vertical levels extending from the surface to 4 hPa, and a horizontal resolution of T42 (a roughly 2.8° × 2.8° Gaussian grid). It has a highly sophisticated physical parameterization package for subgrid scale processes such as boundary layer turbulence, radiation, clouds and convection. Deep precipitating convection and shallow convection are treated by two different schemes, i.e., Zhang and McFarlane (1995) and Hack (1994) schemes, respectively. The Zhang and McFarlane deep convection scheme makes use of the ensemble plume concept to represent convective clouds (Arakawa and Schubert 1974) and simplifies it to a bulk mass flux form. It also includes representation of saturated convective-scale downdrafts as an inverted plume. The Hack shallow convection scheme is a modified moist convective adjustment scheme in the mass flux form. The scheme checks the local instability of the temperature stratification, as opposed to the deep instability used in the deep convection scheme, and adjusts the three adjacent model layers where local instability is present to a neutral state.

Three modifications, i.e., a revised convection closure assumption, convection trigger condition and convective momentum transport (CMT), are made to the deep convection
scheme in ISUGCM. The revised closure assumption is based on ARM and TOGA-COARE observations. It relates convection to the destabilization of the tropospheric layer above the planetary boundary layer by the large-scale processes (Zhang 2002). The trigger condition for deep convection is based on the cloud-resolving simulations, i.e., the convection is activated when the CAPE value increase due to the large-scale forcing exceeds certain threshold (70 J kg$^{-1}$ hr$^{-1}$; Wu et al. 2007b). The CMT parameterization scheme is validated by and simplified based on the cloud-resolving simulations (Zhang and Cho 1991; Wu et al. 2003; Zhang and Wu 2003). The CMT scheme considers the vertical redistribution of the horizontal momentum by convection, and accounts for the role of perturbation pressure field generated by the interaction of convection with large-scale circulation in vertical momentum transport (Wu and Yanai 1994). The CMT-induced convective heating plays an important role in shaping the Hadley circulation (Song et al. 2008a, b).

Four 10-year (1979-1988) simulations with observed sea surface temperatures are analyzed (Table 3.1). The control simulation CTL is conducted using ISUGCM with the original deep convection scheme as in standard CCM3, while the simulation ISUCCM3 is performed with the inclusion of all three modifications in the convection scheme. Two sensitivity simulations, i.e., NOCMT and NOTRI, are performed to investigate the impacts of each of three modifications on the MJO. The simulation NOCMT only includes the revised closure and trigger condition in the convection scheme, and the simulation NOTRI only applies the revised closure in the scheme.
Table 3.1. List of four model simulations

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Description</th>
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<tbody>
<tr>
<td>CTL</td>
<td>Simulation with the original convection scheme as in standard CCM3</td>
</tr>
<tr>
<td>ISUCCM3</td>
<td>Simulation with the revised convection closure, convection trigger and CMT</td>
</tr>
<tr>
<td>NOCMT</td>
<td>As ISUCCM3 but without CMT</td>
</tr>
<tr>
<td>NOTRI</td>
<td>As ISUCCM3 but without CMT and convection trigger</td>
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</table>

The pentad (5-day mean) precipitation product of the Climate Prediction Center (CPC) Merged Analysis of Precipitation (CMAP) is used to evaluate the climate simulations. Observations from rain gauges and precipitation estimates from several satellite-based algorithms (infrared and microwave) are merged by the technique described in Xie and Arkin (1996, 1997). The Advanced Very High Resolution Radiometer (AVHRR) OLR data is interpolated in time and space from National Oceanic and Atmospheric Administration (NOAA) twice-daily OLR values and averaged to once-daily (Liebmann and Smith 1996). National Centers for Environmental Prediction (NCEP) reanalysis datasets (Kalnay et al. 1996), including the wind, OLR, precipitation, vertical velocity, latent heat flux and specific humidity, are used for the comparison. The datasets are regridded to T42 for comparing with the model output.

In this paper, two analysis methods are applied to examine the MJO characteristics and structure. The empirical orthogonal function (EOF) analysis is used to obtain physical and dynamic independent patterns from datasets. Each spatial EOF pattern is associated with a temporal principal component (PC). The first several EOFs can explain the majority of the data variability. Many analyses presented in the following sections are obtained by projecting
simulated fields onto the observed EOF patterns and using the PC time series for the regression. The wavenumber–frequency spectra analysis (Maloney and Hartmann 2001) is also used to examine the MJO power in the study. This method resolves transient wave along latitudinal belt into eastward-moving and westward-moving components. Positive (negative) frequency and period represent the eastward (westward) propagation.

3. Characteristics and structure of the simulated MJO

3.1 Basic features

The October to April rainfall climatology provides the basic state to the discussion of MJO signal. The data during October-April is used for the strong MJO activity during this period. The ten-year (1979-88) October to April averaged precipitation in CTL shows an obviously split ITCZ with less precipitation over the equatorial western Pacific and Indian Ocean compared to the CMAP precipitation (Fig. 3.1). However the split ITCZ problem is improved in NOTRI with the inclusion of new closure in the convection scheme (not shown). And with the convection trigger condition added in the scheme, the amplitude of precipitation along the equator in NOCMT (not shown) becomes close to the CMAP. With all three modifications in the convection scheme, ISUCCM3 precipitation shows a better agreement with the CMAP not only in the amplitude but also in the distribution. In South Pacific Convergence Zone, the precipitation belt is northwestern-southeastern in ISUCCM3 and CMAP, but is west-eastern direction in CTL.

The impacts of modified convection scheme on the MJO characteristics and structure can be visually identified from the hovmöller diagram of 850-hPa zonal wind anomalies averaged
across the equatorial belt (5°N to 5°S) during 1987 (Fig. 3.2). The anomalies are obtained by subtracting the zonal and time means (Yanai et al. 2000). Several eastward-propagating bands of anomalous westerlies can be found in the observations (Fig. 3.2a). These MJO-related disturbances start from the Indian Ocean to the western Pacific, and in some cases it extends to the eastern Pacific. ISUCCM3 shows eastward-propagating feature but with more high frequency variability than the observations (Fig. 3.2b). ISUCCM3 also presents shorter life cycle of MJO signal than the observations. However, the dominant westward-propagating feature is indicated in CTL (Fig. 3.2c). The inclusion of revised convection closure, convection trigger condition and CMT in the convection scheme helps ISUCCM3 simulate the eastward-propagating systems compared to the westward-propagating systems in CTL.

To characterize MJO variability, the variance of daily observed AVHRR OLR for the period of ten-year (1979-88) is presented in Fig. 3.3a. A 20-70-day Lanczos (1956) filter has been used to highlight the MJO-related variability. The observed largest variance of OLR is coincident with the regions of heaviest mean precipitation and the MJO-related strong convective center over the Indian Ocean and western Pacific (not shown). The minimum over the maritime continent may be due to the strong land heating (cooling) during the day (night) which tends to favor the variability of convection on shorter timescales (e.g., diurnal cycle) than the 20-70-day period. The ISUCCM3 catches those features with a ten-year (1979-88) averaged 20-70-day OLR variance (Fig. 3.3b) which is stronger than observations, but CTL does not show a well-defined MJO-related variance over the Indian Ocean and western Pacific, especially near the equator (Fig. 3.3e). The largest OLR variance, which corresponds
to the center of deep convection, shifts from the southern tropics to the Asian monsoon region in both ISUGCM and the observations from winter (December, January, February; DJF) to summer (June, July, August; JJA). The simulations of spatial distribution and amplitude of the MJO-related OLR variance for all four seasons are improved in ISUCCM3 as compared to the CTL. In the NOCMT run which excludes the CMT from the convection scheme, the spatial structure of OLR variances (Fig. 3.3c) is similar to that in ISUCCM3 (Fig. 3.3b), but the amplitude is weaker than that in ISUCCM3. For example, over the western Pacific, the NOCMT OLR variance around 160°E is smaller than ISUCCM3 variance in DJF. In NOTRI run which only includes the revised convection closure, the spatial distribution and amplitude of the OLR variance (Fig. 3.3d) are quite different from those in NOCMT (Fig. 3.3c) but are similar to those in CTL (Fig. 3.3e), which suggests that the trigger condition plays an important role in simulating the intraseasonal variability.

Figure 3.4 shows lag correlations of daily values of OLR at each longitude with a base time series of daily 200-hPa velocity potential at 90°E, both averaged between 10°N and 10°S and filtered to retain the variability at periods of 20-100 days. The observed eastward propagation of the MJO-related convective activity shows clearly with positive correlations extending from the western Indian Ocean at a lag of -15 days to the date line at a lag of around +20 days (a phase speed of ~5 m s⁻¹) in Fig. 3.4a. The ISUCCM3, like observations, shows the MJO-related eastward propagation from the Indian Ocean to the western Pacific with a phase speed of ~5 m s⁻¹ (Fig. 3.4b), but with smaller amplitudes and short periods (the MJO-related signal sustains about 20-day span compared to the 30-day span in NCEP). The main signal of CTL is a westward propagation centered at 90°E, and a weak signal appears to propagate
eastward from 120°E to 180°E (Fig. 3.4e). When the CMT is removed, the eastward propagation of the MJO-related convective activity in NOCMT (Fig. 3.4c) is similar to that in ISUCCM3 and observations, but the MJO phase speed of ~10 m s⁻¹ is greater than that in ISUCCM3. When both the CMT and convection trigger condition are removed and only the revised convection closure is kept in the scheme, NOTRI still simulates an eastward propagation, but the phase speed of ~15 m s⁻¹ is even faster than the speed simulated by NOCMT and the eastward propagation stops around 150°E (Fig. 3.4d). The impacts of both CMT and convection trigger condition decrease the speed of eastward propagation. The inclusion of the revised convection closure in NOTRI results in the eastward propagating signal while the old convection closure produces the westward propagation in CTL.

Averaged wave number-frequency spectra for observed equatorial (10°N-10°S) 200-hPa zonal wind is plotted in Fig. 3.5a. Ten years data is used and a high pass filter is applied to remove the periods of annual cycle and lower before the spectra computation. The observed zonal wind spectrum is dominated by the power at zonal wave number 1 and eastward periods of 30-90 days. The maximum variances are near 60 days and 35 days and wave number 1. Both ISUCCM3 and CTL show a preference for the eastward power, especially in wave number 1 (Figs. 3.5b and 3.5e), which is similar to the observations. But the variance at intraseasonal timescales for ISUCCM3 is larger than that for CTL, and is closer to the observations. With the removal of CMT from the scheme, the maximum power in the NOCMT simulation is near period of 60 days and wave number 1 (Fig. 3.5c), which is similar to the ISUCCM3. However, the peak of power at wave number 1 and eastward periods of 25-33 days disappears. This suggests, with the impact of CMT, ISUCCM3 tends
to have less power at slightly higher frequency (around 30 days). With both CMT and convective trigger excluded from the scheme, NOTRI produces much less variance at intraseasonal timescales than NOCMT for eastward-propagating disturbances, especially near 60 days and wave number 1 (Fig. 3.5d). This indicates that the impact of convection trigger tends to enhance the MJO-related convection activity. The comparison between NOTRI and CTL shows that the revised convection closure plays an important role in simulating the eastward propagating signal at higher frequencies (Figs. 3.5d-e).

3.2 Horizontal structure

In this section, the horizontal structure of simulated MJO in the upper and lower troposphere is compared with observations using the EOF analysis. Following the approach used by Duffy et al. (2003) and Sperber (2004), the simulated fields are projected onto the observed leading patterns to ensure all four simulations are treated identically. The observed leading patterns defined here are based on 10-year (1979-1988) AVHRR OLR. The EOF analysis on band-passed 10-year AVHRR OLR is performed to show the MJO-related convection signature (Fig. 3.6). The total explained variance of first two modes is 23.7%. The EOF-1 suggests a MJO-related convective center over the Indian Ocean around 90°E (Fig. 3.6a). The EOF-2 shows a MJO-related convective center over the western Pacific around 125°E (Fig. 3.6b). The maximum positive correlation coefficient between AVHRR OLR PC-1 and PC-2 time series is about 0.63 around -12 days (not shown). With this lagged correlation between PC-1 and PC-2, the convective anomaly center over the Indian Ocean leads that over the western Pacific. The model and NCEP OLR will be projected to the band-passed 10-year AVHRR OLR patterns in the following analysis.
After projection, the close correspondence between the reduced OLR (negative values) and enhanced precipitation (positive values) due to the MJO-related deep convection over the Indian Ocean and vicinity of the western Pacific is clearly seen in the lag 0 regressions of the PC time series onto the band-passed NCEP reanalysis wind, OLR and CMAP (Figs. 3.7a-b and 3.7g-h). The MJO-related deep convective center with the corresponding enhanced precipitation and reduced OLR (hereafter MDC) is over the region of 70°-100°E and 5°N-15°S in the PC-1 regressions (Figs. 3.7a-b), and moves eastward to 100°-160°E and 5°N-15°S in the PC-2 regressions (Figs. 3.7g-h). In the lower troposphere (850 hPa), the westerly anomalies dominate west of the MDC and reach the east edge of the MDC, and the easterly inflows dominate the east side of the MDC for NCEP (Figs. 3.7b and 3.7h). In the upper troposphere (200 hPa), the dominant outflow from the east edge of the MDC to its whole west side is composed of easterly anomalies, and the westerly anomaly is the dominant wind in the east side of the MDC (Figs. 3.7a and 3.7g).

The ISUCCM3 has an improved MDC simulation compared to the CTL (Figs. 3.8 and 3.9). The MDC is located over 70°-180°E and 5°N-15°S in both NCEP and ISUCCM3, but it shifts slightly north in CTL. The maxima of the MDC shift eastward in ISUCCM3 compared to CTL. The amplitudes of OLR and precipitation anomalies in ISUCCM3 are larger than those in CTL and are close to the observations (Figs. 3.8a-b and 3.9a-b). The coherent relationship between the low-level equatorial convergent westerly inflow and the MDC is better represented in ISUCCM3 than in CTL (Figs. 3.8b, h and 3.9b, h). Both the location and strength of near-equatorial upper-tropospheric divergent flow in ISUCCM3 is closer to the observations than those in CTL (Figs. 3.8a, g and 3.9a, g).
The surface heat flux is an important factor for the evolution of MJO through its interaction with deep convection. Figures 3.7e and 3.7k present the NCEP latent heat flux regressions using PC-1 and PC-2. The latent heat flux has a positive center over the Indian Ocean around 60°-90°E and 0°N corresponding to the MDC (Fig. 3.7e). The center moves eastward to the maritime continent around 120°E in Fig. 3.7k. These features are well represented in ISUCCM3 compared to CTL, especially over the Indian Ocean (Figs. 3.8e and 3.9e). The latent heat flux centers of both observations and ISUCCM3 are around 60°-90°E over the Indian Ocean for regressions using PC-1. However, in CTL, the centers shift westward, around 40°-60°E close to the Africa.

Figures 3.7f and 3.7l show that the observed 500-hPa vertical velocity regressions are consistent with the reduced OLR and enhanced precipitation over the Indian Ocean and western Pacific. The anomalies center of upward motion is around 80°E over the Indian Ocean in regressions obtained using PC-1, and it moves eastward to the western Pacific around 125°E with the use of PC-2. The maximum of ISUCCM3 500-hPa upward motion anomalies is larger than the observations, but it locates in the same area as the NCEP (Figs. 3.8f and 3.8l). The CTL vertical velocity anomalies are comparable to the observations, but the location of upward motion anomaly center is not so well presented as ISUCCM3 (Figs. 3.9f and 3.9l). For example, in regressions using PC-1, corresponding to the latent heat flux center, the upward anomaly center is around 60°-90°E in both observations and ISUCCM3, but is mainly around 60°E in CTL corresponding to its latent heat flux center at 40°-60°E.
Figures 3.7c-d and 3.7i-j show 200-hPa and 850-hPa stream function regressions for NCEP. The low-level leading anticyclone and trailing cyclone responding to the convection are well developed around the equator with larger amplitude in the southern Hemisphere, which is consistent with the maxima of the MDC being displaced south of the equator. The pattern of anticyclone and cyclone in the upper troposphere is just opposite to that in the lower troposphere, which indicates a baroclinic structure in the wind fields. The anticyclone and cyclone move eastward from the Indian Ocean (Figs. 3.7c-d) to the western Pacific (Figs. 3.7i-j). Comparing with the observations, the forced Rossby wave response and the baroclinic wind response to the MDC are represented in ISUCCM3 (Figs. 3.8c-d and 3.8i-j).

Regressions with the 850-hPa and 200-hPa winds in ISUCCM3 and NCEP indicate the near-equatorial convergent flow centered on the convective maxima. While the structure of upper- and lower-tropospheric stream function is simulated by ISUCCM3, the amplitudes of the anticyclone and cyclone in ISUCCM3 are smaller than those in NCEP, but are larger than those in CTL (Figs. 3.9c-d and 3.9i-j).

The impacts of three modifications to the convection scheme on the horizontal structure of simulated MJO are illustrated in the above analysis. To examine the contribution of each modification, three pairs of simulations are compared, i.e., NOCMT vs. ISUCCM3, NOTRI vs. NOCMT, and CTL vs. NOTRI. The analysis of NOCMT in comparison with ISUCCM3 will illustrate which feature of the MJO is influenced by the CMT. The amplitudes of OLR and precipitation anomalies in NOCMT are comparable to those in ISUCCM3, and the location of MDC in NOCMT is also similar to that in ISUCCM3 (Figs. 3.10a-b and 3.10g-h). However, NOCMT produces smaller amplitudes of the low-level leading anticyclone and
trailing cyclone responding to the convection compared to ISUCCM3, especially over the Indian Ocean for the regressions using PC-1 (Fig. 3.10d), and the corresponding equatorial convergent inflow is also weaker in NOCMT than that in ISUCCM3. The coherent relationship between the low-level equatorial convergent flow and the MDC is not represented well in NOCMT as compared to ISUCCM3 (Figs. 3.10b and 3.10h). For example, in the lower troposphere for NOCMT, the equatorial convergent westerly inflow stops west of the MDC and does not reach the east edge of the MDC for the regressions using PC-1, and the equatorial convergent easterly inflow coupled with the MDC is produced from 140°E to 80°E along the equator (Fig. 3.10b). The comparison between ISUCCM3 and NOCMT shows that the inclusion of CMT in ISUCCM3 enhances the amplitudes of lower-tropospheric anticyclone and cyclone, and results in the more coherent lower-tropospheric equatorial convergent inflow and MDC.

Figures 3.11a-b and 3.11g-h show the simulation of MDC in NOTRI. The amplitude of MDC in NOTRI is smaller than that in NOCMT. Corresponding to the weak MDC in NOTRI, the anomalies of 500-hPa vertical velocity, latent heat flux, upper- and lower-tropospheric anticyclone and cyclone, and equatorial divergent and convergent flow are all weaker than those in NOCMT (Figs. 3.11c-f and 3.11i-l). The 500-hPa vertical velocity anomalies in NOTRI are generally weaker than those in NOCMT (Figs. 3.11f and 3.11l). The latent heat flux anomalies in NOTRI are also smaller than those in NOCMT around 90°E (Figs. 3.11e and 3.10e) and 125°E (Figs. 3.11k and 3.10k). The weak responses of anticyclone, cyclone and the corresponding equatorial divergent and convergent flow with respect to the MDC are seen in Figs. 3.11c-d and 3.11i-j, especially for the maritime continent around 120°E in
regressions using PC-2. The comparison between NOTR and NOCMT indicates that the convection trigger enhances the MJO-related signal.

To examine the contribution of revised convection closure, NOTRI is compared to CTL, and the MDC eastward shift is seen in this comparison. In regressions using PC-1, the MDC appears between 50° and 75°E over the Indian Ocean for CTL (Figs. 3.9a-b), but shifts eastward of 20° in NOTRI (Figs. 3.11a-b). In regressions using PC-2, over the western Pacific, the MDC also shifts eastward of 10° in NOTRI compared to CTL (Figs. 3.11g-h and 3.9g-h). Corresponding to this eastward relocation of the MDC, the 500-hPa upward anomaly center around 65°E over the Indian Ocean in PC-1 regressions of CTL moves 20° eastward in NOTRI (Figs. 3.9f and 3.11f). The positive latent heat flux center also moves eastward about 20° from the west of the Indian Ocean in CTL to the central of the Indian Ocean in NOTRI (Figs. 3.9e and 3.11e). These eastward relocations of MDC and its related atmospheric fields between NOTRI and CTL are mainly due to the use of revised convection closure.

3.3 Vertical structure

Applying the similar projection from the analysis of horizontal structure, the longitude-height cross sections of lag 0 regressions of PC-1 with space (from 5°N to 5°S) averaged band-passed 10-year NCEP data are used to describe the observed vertical structures of MJO. The upper-tropospheric divergence and lower-tropospheric convergence anomalies are the dominant features at 90°E in Fig. 3.12a, which is consistent with convective anomaly center. The westward tilt of divergence anomalies with height around 90°E coupled with the eastward propagation of convective anomaly center suggests the deep convective signal
starting from near surface (e.g., Sperber 2003; Sperber et al. 2005). The observed upward velocity anomalies, corresponding to the anomalies of convergence, are around 700-200 hPa near 90°E (Fig. 3.12b). The low-level easterly anomalies with the positive moisture anomalies on the east side of the convective center (Figs. 3.12c-d) help build up a moisture convergence anomaly. Also, the moisture anomalies show a westward tilt from 60°E to 145°E with the largest enhancement around 600-700 hPa (Fig. 3.12d). The low-level convergence, upward velocity and wet anomalies favor the development of new convection and the eastward propagation is present on the east side of the MJO-related convective center, while the low-level divergence, downward velocity and dry anomalies exist on the west side (Sperber 2003).

In ISUCCM3, dominant upper-tropospheric divergence and lower-tropospheric convergence are present at 90°E (Fig. 3.13a), which is in agreement with observations, but is not represented well in CTL (Fig. 3.13a). The westward tilt of divergence with height is also produced by ISUCCM3, while not show in CTL. The upward velocity anomalies around 700-200 hPa and the low-level wet-than-normal easterly anomalies on the east side of the convective center are better simulated in ISUCCM3 than those in CTL (Figs. 3.13b-d and 3.14b-d). ISUCMM3 shows the westward tilt of the specific humidity anomalies around 90°E (Fig. 3.13d), while CTL does not represent this westward tilt (Fig. 3.14d). With the inclusion of revised convection closure, convection trigger and CMT, ISUCCM3 represents the observed divergence, upward motion and moisture anomalies coupled with the MJO-related convection better than CTL, especially for the westward tilt of moisture anomalies. The westward tilt of divergence and moisture coupled with the low-level easterly anomalies
shows the preconditions of eastward propagation with the low-level moisture convergence on
the east side of the convective center.

To explore the role of each modification in the simulation of MJO, two sensitivity
experiments NOCMT and NOTRI are analyzed and compared with ISUCCM3 and CTL. The
simulation of divergence, upward motion, zonal wind and moisture anomalies coupled with
the convective center in NOCMT (Fig. 3.15) is similar to that in ISUCCM3 (Fig. 3.13),
which indicates the CMT has little influence on the vertical structure of MJO in Indian Ocean
around 90°E. Therefore, the improvement of vertical structure shown in ISUCCM3 is largely
due to the use of revised convection closure and convection trigger. Figure 3.16 shows the
analysis of MJO vertical structure for NOTRI which does not includes the new trigger and
CMT. In the vicinity of convective center, the upper-level divergence and low-level
convergence anomalies coupled with the upward velocity anomalies in NOTRI are much
smaller than those in NOCMT (Figs. 3.16a-b and 3.15a-b). Also, on the east side of the
convective anomaly center, the low-level wet-than-normal easterly anomalies in NOTRI are
weaker than those in NOCMT (Figs. 3.16c-d and 3.15c-d). These results demonstrate the
convection trigger enhances the MJO signal in the vertical. To examine the contribution of
revised convection closure to the improvement, Fig. 3.16 is compared with Fig. 3.14. A
convective anomaly center with the coherent atmospheric variances is present better in
NOTRI than those in CTL. The strongest anomalies of upward velocity between 200-600 hPa
coupled with the divergence and moisture anomalies in NOTRI shift eastward about 15°
compared to those in CTL (Figs. 3.16 and 3.14), and this eastward shift of convective
anomaly center is consistent with the eastward relocation of MDC in section 3.2. The low-
level moisture convergence on the east side of the convective center features the low-level wet-than-normal easterly anomalies in NOTRI in contrast with the dry-than-normal easterly anomalies in CTL (Figs. 3.16c-d and 3.14c-d). The comparison between NOTRI and CTL demonstrates that the revised convection closure is largely responsible for the improved simulations of the MJO-related convective center and the preconditions of eastward propagation in divergence and moisture fields on the east side of the convective center.

To further depict the evolution of physical properties during the life cycle of the MJO with respect to the convective maxima, the lag regressions of the band-passed 10-year divergence, vertical velocity, moisture and wind fields (averaged between 5°N and 5°S) as a function of pressure at 90°E for PC-1 regressions after the projection are analyzed for observations and simulations (e.g., Sperber 2003). In the observations, the near surface convergence anomalies and moistening of the boundary layer appear around -10 day, and the upward motion appears earlier in the low level in advance of the deep convection (Fig. 3.17). These anomalies develop quickly and reach the upper level around lag 0 day with the deep convection building up. During the life cycle of the MJO with respect to the deep convection center, the wind anomaly field changes from easterly through upward to westerly in the low level (Fig. 3.17c). With the modified convection scheme, ISUCCM3 simulates the evolution of atmospheric conditions close to the observations compared to CTL (Figs. 3.18 and 3.19). The convergence and upward motion anomalies initially occur around -15 day in the low level and penetrate to the upper level with the deep convection building up in ISUCCM3 (Figs. 3.18a-b), while in CTL, those anomalies are stationary (Figs. 3.19a-b). The wind anomaly field of ISUCCM3 shows the evolution from easterly through upward to westerly in the low
level within the life cycle of the MJO, which is not present in CTL (Figs. 3.18c and 3.19c). Also, ISUCCM3 shows the evolution of moisture during the life cycle of the MJO closer to the observations compared to CTL (Figs. 3.18d and 3.19d), although the location of initial signal for the wet-than-normal moisture in ISUCCM3 is a little higher than that in the observations. Overall, ISUCCM3 represents the development of atmospheric conditions during the life cycle of the MJO better than CTL, especially for the near surface precursor signals of divergence and upward motion anomalies in advance of the deep convection.

Figure 3.20 shows the evolution of the convective anomaly center coupled with the convergence, upward motion, low-level zonal wind anomalies and moisture anomalies during the MJO life cycle in NOCMT which does not include the CMT in the scheme. Most features in Fig. 3.20 are similar to those in Fig. 3.18 except the near surface moisture precursor signal. In advance of the convective maxima, the positive moisture anomalies begin near surface around -20 day in NOCMT (Fig. 3.20d), but the positive anomalies start around -14 day in ISUCCM3 (Fig. 3.18d). With both CMT and convection trigger excluded from the scheme, the lag regressions of NOTRI shown in Fig. 3.21 is quite different from the NOCMT (Fig. 3.20). First, the precursor signals of anomalies in advance of the convective center initially appear in the upper level during the MJO life cycle in NOTRI (Fig. 3.21), while in NOCMT, the precursor signals begin first at the surface (Fig. 3.20). Second, the amplitude of the MJO-related anomalies in NOTRI is smaller than that in NOCMT. For example, the anomalies of convergence, vertical velocity and moisture are weaker around lag 0 day with the deep convection building up in NOTRI compared to NOCMT (Figs. 3.21 and 3.20). The difference between NOTRI and NOCMT indicates that the impact of the convection trigger
not only improves the evolution of atmospheric conditions during the life cycle of the MJO with respect to the convective center, but also enhances the MJO signal. Finally, the comparison between NOTRI (Fig. 3.21) and CTL (Fig. 3.19) illustrates the impact of revised convection closure on the MJO simulations. With the inclusion of revised convection closure, NOTRI produces the evolution of the convective anomaly center coupled with the convergence, upward motion, zonal wind, and moisture anomalies during the MJO life cycle (Fig. 3.21). However, the atmospheric anomalies associated with the development of MJO-related convective anomaly center are stationary in CTL (Fig. 3.19).

4. Summary

In this study, the MJO simulated by ISUGCM is evaluated against observations. The results demonstrate that the improved MJO simulations are obtained by ISUGCM with the inclusion of the revised convection closure assumption, convection trigger condition and CMT in the convection scheme. The improvements of basic features in the simulations include (1) the improved spatial distribution and amplitude of the MJO-related OLR variance which represents the MJO-related convective center, (2) a more realistic MJO-related eastward propagation from the Indian Ocean to the western Pacific with a phase speed of about 5 m s$^{-1}$ instead of a westward propagation in the control run with the original convection scheme, and (3) the large variance at eastward intraseasonal timescales which suggests an enhanced MJO signal. To study the horizontal and vertical structure of MJO, the ISUGCM simulations are projected to the robust lead-lag relationship of eastward intraseasonal propagating OLR to ensure all simulations are treated identically. The results indicate that the eastward propagating convection and its related atmospheric variances during the MJO life cycle in the
ISUGCM simulation with the modified convection scheme are in agreement with the observations, especially for the preconditions of eastward propagation on the east side of the convective center and the amplitude of the MJO signal.

The MJO is analyzed in three simulations (NOTRI, NOCMT and ISUCCM3) which have the revised closure assumption, convection trigger condition and CMT added one at a time. This enables us to examine the impacts of each modification on the MJO simulation. The use of the revised convection closure assumption results in the eastward propagating signal while the original convection closure produces the strong westward signal, and this improvement is related to the eastward shift of MDC and its consistent atmospheric variances. More realistic eastward location of MJO-related convective center favors the eastward propagating signal from the Indian Ocean to the western Pacific, while the unrealistic location of convective center results in strong westward propagation. With the convection trigger condition added in the scheme, the MJO-related convection activity is enhanced. The explanation for this strong MJO signal is that moist deep convection occurs less frequently but better organized, and the convection is activated only when the increase of CAPE reaches a certain threshold \(70 \text{ J kg}^{-1} \text{ hr}^{-1}\); Wu et al. 2007b). The inclusion of the CMT in the convection scheme leads to less eastward power at slightly higher frequency around 30 days with decreased eastward phase speed and more coherent structure for the MJO-related deep convection and dynamic fields. The decreased eastward power in slightly higher frequency (25-33 days) within the intraseasonal scale is consistent with the reduced kinetic energy corresponding to the decreased phase speed. With the space-time structure analysis, the impact of the CMT is more notable on the horizontal structure than the vertical structure over the Indian Ocean.
The possible reason for this difference is that the variance of CMT-induced convective heating is stronger on the horizontal direction than that on the vertical direction over the Indian Ocean. Further analysis is needed to understand the physical processes responsible for the improvements of the MJO simulations due to the revised convection closure assumption, convection trigger condition and CMT in the convection scheme.
Fig. 3.1. October-April climatology of precipitation rate (mm day$^{-1}$) for (a) CMAP, (b) ISUCCM3, and (c) CTL.
Fig. 3.2. 1987 Hovmöller diagrams of 850 hPa zonal wind anomaly (m s$^{-1}$) averaged between 5°N and 5°S for (a) NCEP, (b) ISUCCM3, and (c) CTL. Areas of negative (easterly anomaly) are shaded. The contour interval is 5 m s$^{-1}$. Straight lines emphasize propagating patterns of zonal wind anomaly.
c) NOCMT OLR ISV 79-88

d) NOTRI OLR ISV 79-88
**Fig. 3.3.** Spatial distribution of variance of 20-70-day band-passed OLR for the four seasons from 1979 to 1988: (a) AVHRR, (b) ISUCCM3, (c) NOCMT, (d) NOTRI, and (e) CTL. In each figure, top left one is for December to February (DJF), top right one is for March to May (MAM), bottom left one is for June to August (JJA), and bottom right one is for September to November (SON). Unit is $W^2 \text{m}^{-4}$. Regions of deep convection are highlighted (shaded $> 240 \ W^2 \text{m}^{-4}$). The contour interval is $80 \ W^2 \text{m}^{-4}$. 
c) NOCMT

\[ \text{lag(days)} \]

\[ \text{lag(days)} \]

\[ -30 \quad -20 \quad -10 \quad 0 \quad 10 \quad 20 \quad 30 \]

\[ 0 \quad 60E \quad 120E \quad 180 \quad 120W \quad 60W \]

\[ -0.5 \quad -0.4 \quad -0.3 \quad -0.2 \quad -0.1 \quad 0 \quad 0.1 \quad 0.2 \quad 0.3 \quad 0.4 \quad 0.5 \]

\[ \text{d) NOTRI} \]

\[ \text{lag(days)} \]

\[ \text{lag(days)} \]

\[ -30 \quad -20 \quad -10 \quad 0 \quad 10 \quad 20 \quad 30 \]

\[ 0 \quad 60E \quad 120E \quad 180 \quad 120W \quad 60W \]

\[ -0.5 \quad -0.4 \quad -0.3 \quad -0.2 \quad -0.1 \quad 0 \quad 0.1 \quad 0.2 \quad 0.3 \quad 0.4 \quad 0.5 \]
Fig. 3.4. Ten-years (1979-88 October-April) lag correlations of 20-100-day band-passed OLR with 200 hPa velocity potential for (a) NCEP, (b) ISUCCM3, (c) NOCMT, (d) NOTRI and (e) CTL. The yellow line represents a phase speed of ~5 m s\(^{-1}\).
Fig. 3.5. Wavenumber–frequency spectra of 200 hPa zonal wind averaged between 10°N and 10°S for (a) NCEP, (b) ISUCCM3, (c) NOCMT, (d) NOTRI, and (e) CTL (years 1979–88). The contour starts from 0.05 m² s⁻² with the interval of 0.05 m² s⁻².
Fig. 3.6. Analysis of AVHRR OLR for 20-100-day band-passed ten years (1979-88 October to April): (a) EOF-1 and (b) EOF-2.
**Fig. 3.7.** Zero lag linear regressions of PC-1 for 20-100-day filtered ten years (1979-88 October to April) (a) NCEP 200 hPa wind (m s$^{-1}$) and OLR (W m$^{-2}$), (b) NCEP 850hPa wind (m s$^{-1}$) and CMAP (mm day$^{-1}$), (c) NCEP 200 hPa wind (m s$^{-1}$) and stream function (m$^2$ s$^{-1}$), (d) NCEP 850 hPa wind (m s$^{-1}$) and stream function (m$^2$ s$^{-1}$), (e) NCEP latent heat flux (W m$^{-2}$), (f) NCEP 500 hPa vertical velocity (Pa s$^{-1}$; negative sign means upward motion), g-l as a-f but for regressions using PC-2. All regressions have been scaled by a one standard deviation of PC-1 to give units. The variables are plotted where the regression is 95% confidence or better.
Fig. 3.8. As Fig. 3.7 but for ISUCCM3. Zero lag linear regressions of PC-1 with 20-100-day filtered (a) 200 hPa wind (m s\(^{-1}\)) and OLR (W m\(^{-2}\)), (b) 850 hPa wind (m s\(^{-1}\)) and precipitation rate (m m day\(^{-1}\)), (c) 200 hPa wind (m s\(^{-1}\)) and stream function (m\(^2\) s\(^{-1}\)), (d) 850 h Pa wind (m s\(^{-1}\)) and stream function (m\(^2\) s\(^{-1}\)), (e) latent heat flux (W m\(^{-2}\)), (f) 500 h Pa vertical velocity (Pa s\(^{-1}\); negative sign means upward motion). g-l as a-f but for regressions using PC-2.
**Fig. 3.9.** As Fig. 3.7 but for CTL. Zero lag linear regressions of PC-1 with 20-100-day filtered (a) 200 hPa wind (m s\(^{-1}\)) and OLR (W m\(^{-2}\)), (b) 850hPa wind (m s\(^{-1}\)) and precipitation rate (mm day\(^{-1}\)), (c) 200hPa wind (m s\(^{-1}\)) and stream function (m\(^2\) s\(^{-1}\)), (d) 850 hPa wind (m s\(^{-1}\)) and stream function (m\(^2\) s\(^{-1}\)), (e) latent heat flux (W m\(^{-2}\)), (f) 500 hPa vertical velocity (Pa s\(^{-1}\); negative sign means upward motion). g-l as a-f but for regressions using PC-2.
Fig. 3.10. As Fig. 3.7 but for NOCMT. Zero lag linear regressions of PC-1 with 20-100-day filtered (a) 200 hPa wind (m s\(^{-1}\)) and OLR (W m\(^{-2}\)), (b) 850hPa wind (m s\(^{-1}\)) and precipitation rate (mm day\(^{-1}\)), (c) 200hPa wind (m s\(^{-1}\)) and stream function (m\(^2\) s\(^{-1}\)), (d) 850 hPa wind (m s\(^{-1}\)) and stream function (m\(^2\) s\(^{-1}\)), (e) latent heat flux (W m\(^{-2}\)), (f) 500 hPa vertical velocity (Pa s\(^{-1}\); negative sign means upward motion). g-l as a-f but for regressions using PC-2.
Fig. 3.11. As Fig. 3.7 but for NOTRI. Zero lag linear regressions of PC-1 with 20-100-day filtered (a) 200 hPa wind (m s$^{-1}$) and OLR (W m$^{-2}$), (b) 850 hPa wind (m s$^{-1}$) and precipitation rate (mm day$^{-1}$), (c) 200 hPa wind (m s$^{-1}$) and stream function (m$^2$ s$^{-1}$), (d) 850 hPa wind (m s$^{-1}$) and stream function (m$^2$ s$^{-1}$), (e) latent heat flux (W m$^{-2}$), (f) 500 hPa vertical velocity (Pa s$^{-1}$; negative sign means upward motion). g-l as a-f but for regressions using PC-2.
Fig. 3.12. NCEP longitude-height cross sections of zero lag linear regressions of PC-1 with 5°N-5°S-averaged 20-100-day filtered (a) divergence (s⁻¹), (b) vertical velocity (Pa s⁻¹), (c) zonal wind (m s⁻¹) and vertical velocity (Pa s⁻¹) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s⁻¹, (d) specific humidity (g kg⁻¹). All regressions have been scaled by a one standard deviation of PC-1 to give units.
Fig. 3.13. As Fig. 3.12 but for ISUCCM3. Longitude-height cross sections of zero lag linear regressions of PC-1 for (a) divergence (s$^{-1}$), (b) vertical velocity (Pa s$^{-1}$), (c) zonal wind (m s$^{-1}$) and vertical velocity (Pa s$^{-1}$) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s$^{-1}$, (d) specific humidity (g kg$^{-1}$).
Fig. 3.14. As Fig. 3.12 but for CTL. Longitude-height cross sections of zero lag linear regressions of PC-1 for (a) divergence (s$^{-1}$), (b) vertical velocity (Pa s$^{-1}$), (c) zonal wind (m s$^{-1}$) and vertical velocity (Pa s$^{-1}$) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s$^{-1}$, (d) specific humidity (g kg$^{-1}$).
Fig. 3.15. As Fig. 3.12 but for NOCMT. Longitude-height cross sections of zero lag linear regressions of PC-1 for (a) divergence (s\(^{-1}\)), (b) vertical velocity (Pa s\(^{-1}\)), (c) zonal wind (m s\(^{-1}\)) and vertical velocity (Pa s\(^{-1}\)) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s\(^{-1}\), (d) specific humidity (g kg\(^{-1}\)).
Fig. 3.16. As Fig. 3.12 but for NOTRI. Longitude-height cross sections of zero lag linear regressions of PC-1 for (a) divergence (s$^{-1}$), (b) vertical velocity (Pa s$^{-1}$), (c) zonal wind (m s$^{-1}$) and vertical velocity (Pa s$^{-1}$) vectors (vertical velocity time -100) and contours of zonal wind with the interval of 0.5 m s$^{-1}$, (d) specific humidity (g kg$^{-1}$).
**Fig. 3.17.** Time lag vs. height plots of linear regressions of PC-1 with 90°E (5°N-5°S averaged) 20-100-day filtered NCEP (a) divergence (s⁻¹), (b) vertical velocity (Pa s⁻¹; negative sign means upward motion), (c) zonal wind (m s⁻¹) and vertical velocity (Pa s⁻¹) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s⁻¹, (d) specific humidity (g kg⁻¹). All regressions have been scaled by a one standard deviation of PC-1 to give units. Time lags run from -25 to 25 days.
Fig. 3.18. As Fig. 3.17 but for ISUCCM3. Time lag vs. height plots of linear regressions at 90°E of PC-1 for (a) divergence (s$^{-1}$), (b) vertical velocity (Pa s$^{-1}$; negative sign means upward motion), (c) zonal wind (m s$^{-1}$) and vertical velocity (Pa s$^{-1}$) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s$^{-1}$, (d) specific humidity (g kg$^{-1}$).
Fig. 3.19. As Fig. 3.17 but for CTL. Time lag vs. height plots of linear regressions at 90°E of PC-1 for (a) divergence (s⁻¹), (b) vertical velocity (Pa s⁻¹; negative sign means upward motion), (c) zonal wind (m s⁻¹) and vertical velocity (Pa s⁻¹) vectors (vertical velocity times 100) and contours of zonal wind with the interval of 0.5 m s⁻¹, (d) specific humidity (g kg⁻¹).
Fig. 3.20. As Fig. 3.17 but for NOCMT. Time lag vs. height plots of linear regressions at 90°E of PC-1 for (a) divergence (s⁻¹), (b) vertical velocity (Pa s⁻¹; negative sign means upward motion), (c) zonal wind (m s⁻¹) and vertical velocity (Pa s⁻¹) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s⁻¹, (d) specific humidity (g kg⁻¹).
Fig. 3.21. As Fig. 3.17 but for NOTRI. Time lag vs. height plots of linear regressions at 90°E of PC-1 for (a) divergence (s\(^{-1}\)), (b) vertical velocity (Pa s\(^{-1}\); negative sign means upward motion), (c) zonal wind (m s\(^{-1}\)) and vertical velocity (Pa s\(^{-1}\)) vectors (vertical velocity times -100) and contours of zonal wind with the interval of 0.5 m s\(^{-1}\), (d) specific humidity (g kg\(^{-1}\)).
CHAPTER 4. PHYSICAL MECHANISMS FOR THE MAINTENANCE OF GCM-SIMULATED MADDEN-JULIAN OSCILLATION OVER THE INDIAN OCEAN AND PACIFIC

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Abstract
The kinetic energy budget is conducted to analyze the physical processes responsible for the improved Madden-Julian Oscillation (MJO) simulated by the Iowa State University general circulation models (ISUGCM). The modified deep convection scheme that includes the revised convection closure, convection trigger condition and convective momentum transport (CMT) enhances the equatorial (10°S-10°N) MJO-related perturbation kinetic energy (PKE) in the upper troposphere and leads to more robust and coherent eastward propagating MJO signal. In the MJO source region—the Indian Ocean (45°E-120°E), the upper-tropospheric MJO PKE is maintained by the convergence of vertical wave energy flux and the barotropic conversion through the horizontal shear of mean flow. In the convectively active region—the western Pacific (120°E-180°), the upper-tropospheric MJO PKE is supported by the convergence of horizontal and vertical wave energy fluxes. Over the central-eastern Pacific (180°-120°W), where convection is suppressed, the upper-tropospheric MJO PKE is mainly
due to the convergence of horizontal wave energy flux. The deep convection trigger condition produces stronger convective heating which enhances the perturbation available potential energy (PAPE) production and the upward wave energy fluxes, and leads to the increased MJO PKE over the Indian Ocean and western Pacific. The trigger condition also enhances the MJO PKE over the central-eastern Pacific through the increased convergence of meridional wave energy flux from the subtropical latitudes of both hemispheres. The revised convection closure affects the response of mean zonal wind shear to the convective heating over the Indian Ocean and leads to the enhanced upper-tropospheric MJO PKE through the barotropic conversion. The stronger eastward wave energy flux due to the increase of convective heating over the Indian Ocean and western Pacific by the revised closure is favorable to the eastward propagation of MJO and the convergence of horizontal wave energy flux over the central-eastern Pacific. The convection-induced momentum tendency tends to decelerate the upper-tropospheric wind which results in a negative work to the PKE budget in the upper troposphere. However, the convection momentum tendency accelerates the westerly wind below 850 hPa over the western Pacific, which is partially responsible for the improved MJO simulation.

1. Introduction

Simulations of the Madden-Julian Oscillation (MJO), a dominant mode of intraseasonal variability, by general circulation models (GCMs) have been gradually improved in the last decade through studies on the representation of cloud systems in terms of grid-scale physical variables (e.g., Wang and Schlesinger 1999; Maloney and Hartmann 2001; Sperber et al. 2005; Zhang and Mu 2005b; Liu et al. 2005; Khairoutdinov et al. 2005; Ziemiański et al. 2005;...
Some of well-observed spatial and temporal features of the MJO (Madden and Julian 1972, 1994) are reproduced by several GCMs with varying degrees of success. Wang and Schlesinger (1999) showed that the enhancement of MJO signal in GCM with the use of large threshold of relative humidity allows the accumulation of moist static energy to a certain amount to trigger the convection in three different convection schemes. Maloney and Hartmann (2001) also improved intraseasonal variability of tropical precipitation and zonal winds by using the microphysics of cloud together with the relaxed Arakawa-Schubert convection scheme (Sud and Walker 1999) in NCAR Community Climate Model, version 3.6 (CCM3). Khairoutdinov et al. (2005) showed that the enhanced intraseasonal variability is simulated by the NCAR Community Atmosphere Model (CAM) with the cloud-resolving model in replacing the convection and cloud parameterization. Analyzing ECHAM4 coupled and uncoupled GCM simulations, Sperber et al. (2005) demonstrated that the eastward propagating MJO zonal wind and latent heat flux are related to the horizontal resolution, a realistic mean state simulation and air-sea interaction. Zhang and Mu (2005b) obtained the enhanced intraseasonal variability in precipitation, zonal wind, and outgoing longwave radiation (OLR), and the eastward propagating MJO from the Indian Ocean to the Pacific by modified the closure assumption in the Zhang–McFarlane convection parameterization scheme of NCAR CCM3. Liu et al. (2005) simulated an improved mean state, intraseasonal variability, space–time power spectra, and coherent eastward propagation of MJO precipitation using a modified Tiedtke (1989) convection scheme in NCAR CAM. Ziemiański et al. (2005) presented a more realistic simulation of MJO-like system including the large-scale organization of the tropical superclusters, eastward propagation, and lower-
tropospheric cyclonic and upper-tropospheric anticyclonic gyres by applying the cloud-resolving convection parameterization over the western Pacific warm pool in NCAR CAM.

The improved simulations of intraseasonal variability allow further diagnostic analysis to investigate the physical processes responsible for the development and maintenance of the MJO. Yanai et al. (2000) examined the structure and evolution of two MJO events during the TOGA COARE Intensive Observing Period (November 1992 to February 1993). They suggested that the interaction between the large-scale motion and convection plays the predominant role in the maintenances of perturbation kinetic energy (PKE) associated with 30-60-day period through the conversion of perturbation available potential energy (PAPE) generated by convection heating over the Indian Ocean-western Pacific. However, over the central-eastern Pacific where the convection is suppressed, the strong equatorward fluxes of wave energy from the extratropic latitudes result in the convergence of horizontal wave energy flux in the equatorial upper troposphere which maintains the MJO PKE. Using 15-year ECMWF Reanalysis datasets, Chen and Yanai (2000) confirmed that the PKE associated with MJO over the warm pool is sustained by the conversion of PAPE generated by the convective heating. Mu and Zhang (2006, 2008) analyzed the energetics of MJO simulated by the NCAR CAM and showed that the observed mechanisms responsible for the MJO PKE production are reproduced by the simulations with the improved convection closure in the Zhang-McFarlane convection scheme (Zhang and McFarlane 1995; Zhang 2002). The maintenance of PKE associated with MJO in the upper troposphere is mainly due to the conversion of PAPE and vertical transport in the convective active region, but the
convergence of horizontal wave energy flux and barotropic conversion in the convectively suppressed region.

With the revised closure, convection trigger condition and the convective momentum transport (CMT) in the ISUGCM convection scheme, Deng and Wu (2010) showed that the MJO simulations are improved in amplitude, spatial distribution, eastward propagation, and horizontal and vertical structures, especially for the coherent feature of eastward propagating convection and the precursor sign of the convective center. The objectives of this paper are to examine the physical processes that contribute to the perturbation kinetic energy of enhanced MJO over the Indian Ocean and Pacific in ISUGCM simulations. The model simulations and observational data are described in section 2. The energetic features associated with MJO are presented in section 3. The perturbation kinetic energy budgets of simulated MJO are conducted in section 4. The summary is given in section 5.

2. Simulations and observational data

Four ten-year (1979-1988) ISUGCM simulations presented in Deng and Wu (2010) are used in this paper. ISUGCM is a global climate model based on the NCAR CCM3 (Kiehl et al. 1998). It has 18 hybrid vertical levels extending from the surface to 4 hPa, and a horizontal resolution of T42 (a roughly 2.8° × 2.8° Gaussian grid). Deep precipitating convection and shallow convection are treated by two different schemes, i.e., Zhang and McFarlane (1995) and Hack (1994) schemes, respectively. Three modifications, i.e., the revised convection closure assumption, convection trigger condition and convective momentum transport (CMT), are made to the deep convection scheme in ISUGCM. The revised closure relates
convection to the destabilization of the tropospheric layer above the planetary boundary layer by the large-scale processes (Zhang 2002). The trigger condition obtained from the cloud-resolving simulations activates deep convection when the CAPE value increase due to the large-scale forcing exceeds a certain threshold (70 J kg\(^{-1}\) hr\(^{-1}\); Wu et al. 2007b). The CMT parameterization validated by the cloud-resolving simulations takes into account the role of perturbation pressure field generated by the interaction of convection with large-scale circulation in the vertical momentum transport (Zhang and Cho 1991; Wu and Yanai 1994; Wu et al. 2003, 2007a; Zhang and Wu 2003).

In Table 4.1, the control simulation CTL is conducted using ISUGCM with the original deep convection scheme as in the standard CCM3, and the simulation ISUCCM3 is performed with the inclusion of all three modifications in the convection scheme. The NOCMT and NOTRI are performed to examine the impacts of each modification on the MJO. The simulation NOCMT only includes the revised closure and trigger condition in the convection scheme, and the simulation NOTRI only applies the revised closure in the scheme. National Centers for Environmental Prediction (NCEP) reanalysis datasets (Kalnay et al. 1996), including the wind, vertical velocity and temperature, are regridded to T42 resolution for matching the model output.

**Table 4.1. List of four ISUGCM simulations**

<table>
<thead>
<tr>
<th>Simulation</th>
<th>Description</th>
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<tbody>
<tr>
<td>CTL</td>
<td>Control simulation with the original convection scheme of CCM3</td>
</tr>
<tr>
<td>ISUCCM3</td>
<td>ISUGCM simulation with the revised convection closure assumption, convection trigger condition and CMT</td>
</tr>
<tr>
<td>NOCMT</td>
<td>ISUGCM simulation with the revised convection closure assumption, convection trigger condition</td>
</tr>
<tr>
<td>NOTRI</td>
<td>ISUGCM simulation with the revised convection closure assumption</td>
</tr>
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3. Energetic characteristics of simulated MJO in comparison with observations

a. Mean vertical circulation along the equator

To illustrate the large-scale vertical circulation, the ten-year (1979-88) October to April mean vertical cross section of zonal and vertical wind averaged across the equatorial belt (10°N to 10°S) is given in Fig. 4.1. In NCEP, the zonal wind from the eastern Indian Ocean (~60°E) to the date line is easterly in the upper troposphere and weak westerly in the lower troposphere, and the zonal wind direction is just opposite over the central to eastern Pacific and Atlantic Ocean (Fig. 4.1a). The strong upper-tropospheric easterly centered at 150 hPa over the Indian Ocean to the date line is coupled with a strong band of upward motion and active convection reflected in the OLR field (not shown). The center of upward velocity is located near 150°E. The strong upper-tropospheric westerly is centered at 150-200 hPa over the central to eastern Pacific where convection is suppressed. The lower-tropospheric easterly and the upper-tropospheric westerly between 180° and 90°W make up the lower and upper branches of Walker circulation.

The mean vertical circulation simulated by ISUCCM3 is in general agreement with the observations, but the upper branch of Walker circulation is stronger than that in NCEP, as well as the stronger and wider band of upward motion in ISUCCM3 (Fig. 4.1b). The band of upward motion exists from the Indian Ocean to the central Pacific with the maximum upward velocity located over the western Pacific. The impact of modified convection scheme on the mean vertical circulation can be readily identified from the difference between ISUCCM3 and CTL presented in Fig. 4.1c. A notable feature is the enhanced upward vertical velocity over the western Pacific between 150°E and 180°. The lower-tropospheric westerly between
120°E and 150°W is increased from CTL to ISUCCM3 with a peak at 850 hPa and 150°E, and the easterly is increased in the upper troposphere. Outside the western and central Pacific, however, the increased westerly (easterly) is present in the upper (lower) troposphere.

b. Distribution of PKE along equator

In this section, we examine the perturbation kinetic energy associated with the MJO along the equator following the methodology of Yanai et al. (2000). The PKE is defined as

\[ \bar{k} = \frac{\overline{u'^2} + \overline{v'^2}}{2}, \]

where the overbar and prime represent the time mean and the deviation from the time mean, respectively. Figure 4.2a illustrates the longitude-height cross section of the mean NCEP PKE for the period of ten-year (1979-88 October to April) along the equator (10°N-10°S). A 20-100-day filter has been used to highlight the MJO-related variability of PKE. The large PKE is centered at 150 hPa, and locates over the Indian Ocean-western Pacific, eastern Pacific and Atlantic Ocean (e.g., Yanai et al. 2000; Mu and Zhang 2006). The maxima of PKE in the Indian Ocean-western Pacific collocate with the upper-tropospheric easterly (Fig. 4.1a) and is closely related with the strong convective activity in this region (e.g., Chen and Yanai 2000). However, the maxima of PKE in the eastern Pacific, where convection is suppressed, are associated with the upper-tropospheric westerly (Fig. 4.1a) and are due to the accumulation of wave energy through the zonal flow (e.g., Webster and Chang 1988). The PKE distribution of ISUCCM3 (Fig. 4.2b) is in agreement with the observations, but the maximum centers shift slightly westward except the one over the western Pacific (e.g., Mu and Zhang 2006). The maximum of PKE near 120°E in the western Pacific locates higher than the observed peak and also has larger amplitudes, which is
consistent with the higher location of the upper-tropospheric easterly center and the stronger band of upward motion in ISUCCM3 (Fig. 4.1b). Similar distribution of PKE is produced by CTL, but the amplitude of PKE is smaller than that in ISUCCM3 and observations, especially over the Indian Ocean (Fig. 4.2c). The maximum of PKE does not extend over the maritime continent in CTL as in ISUCCM3 and observations.

The enhancement of MJO PKE by the modified convection scheme is more clearly documented in Fig. 4.3a which shows the difference of PKE between ISUCCM3 and CTL. Four major positive centers are over the Indian Ocean around 60°E, western Pacific around 120°E, eastern Pacific around 130°W and west coast of South America around 60°W. To further examine the impacts of each modification on the MJO PKE, the differences of MJO PKE between four ISUGCM simulations are presented in Figs. 4.3b-d. The negative centers of PKE difference between ISUCCM3 and NOCMT locate around 200hPa over the African, western Pacific, eastern Pacific and Atlantic Ocean (Fig. 4.3b). The impact of CMT shows on the reduced MJO PKE, especially from eastern Pacific cross the Atlantic Ocean to the African where the suppressed convection is coupled with the upper-tropospheric westerly (Fig. 4.1b). The enhanced MJO PKE in the upper troposphere around 100 hPa coupled with the strong upward motion and easterly is also present over the maritime continent with active convection. The contribution of the convection trigger condition to the enhanced MJO PKE is demonstrated by the difference between NOCMT and NOTRI in Fig. 4.3c. The magnitudes of MJO PKE in NOCMT in the upper troposphere are obviously larger than that in NOTRI with positive difference centers locate over the African-western Indian Ocean, western Pacific, eastern Pacific, west coast of South America and Atlantic Ocean. The PKE
difference from the west coast of South America to the African is associated with a large upper-tropospheric westerly difference between NOCMT and NOTRI (not shown). The influence of revised closure assumption on the MJO PKE is indicated by the difference between NOTRI and CTL in Fig. 4.3d. The major positive peak of PKE difference over the Indian Ocean is related to the more active convection coupling with the strong easterly and upward motion (not shown) in NOTRI compared to CTL. The positive PKE differences are also present in the upper troposphere around 130°W and 60°W.

4. Kinetic energy budgets of the MJO

The analysis in the last section illustrates the contributions of revised closure, convection trigger condition and CMT to the simulations of MJO PKE in the upper troposphere. To further understand the physical processes responsible for the enhanced MJO PKE, we perform the kinetic energy budget of the MJO in this section. Following Yanai et al. (2000),

$$\frac{\partial k}{\partial t} = -\mathbf{v}' \cdot \nabla \mathbf{v} - \mathbf{v}' \omega' \cdot \frac{\partial \mathbf{v}}{\partial p} - \alpha' \omega' - \nabla \cdot \mathbf{F}_h - \frac{\partial F_p}{\partial p} + \mathbf{v}' \cdot \mathbf{F}' + \mathbf{v}' \cdot \mathbf{f}'. \quad (4.1)$$

In Eq. (4.1), $\mathbf{v}$ is the horizontal velocity, $p$ the pressure, $\omega = dp/dt$ (the vertical $p$ velocity), $\nabla$ the isobaric gradient operator, $\alpha$ the specific volume, and $f$ the frictional force per unit mass. $\mathbf{F} (F_u, F_v)$ is the convection-induced momentum tendency (e.g., Wu and Yanai 1994; Wu et al. 2007b). $\mathbf{F}_h \equiv \phi' \mathbf{v}' + k \mathbf{v} + (u'^2 + v'^2) \mathbf{v}' / 2$ is the horizontal wave energy flux which includes both zonal ($F_u$) and meridional ($F_v$) components, and

$F_p \equiv \phi' \omega' + k \omega + (u'^2 + v'^2) \omega' / 2$ the vertical wave energy flux ($\phi$ is the geopotential). The
equation (1) shows that the local time change of PKE is due to the work done by the horizontal (term A) and vertical (term B) shear generation of the mean flow, the conversion from the perturbation available potential energy (PAPE) with the \((\alpha,-\omega)\) correlation (term C), the convergence of horizontal (term D) and vertical (term E) wave energy fluxes, the work done by the convection-induced momentum tendency (term F), and the work done by the frictional force (term G). Term B is negligibly small for large-scale motions (e.g., Yanai et al. 2000). Term G is the work related to the frictional force, and is negligible in the upper troposphere. Therefore, the maintenance of upper-tropospheric PKE is mainly governed by the barotropic conversion by the horizontal shear of the mean flow, the PAPE conversion, the convergence of horizontal and vertical wave energy fluxes and the work done by the convection-induced momentum tendency. In the following presentation, the cross-spectra analysis is used to evaluate the contribution of two variables’ covariance from the period of 20-100 day with respect to the total covariance. In the calculation of the MJO-related covariance terms in Eq. (4.1), the cospectra are integrated over the period of 20-100 day for October to April of each year from 1979 to 1988, and then are averaged over the ten years to get the climatological mean.

Figure 4.4 presents the MJO PKE production due to the barotropic conversion by the horizontal shear of the mean flow. Large observed PKE productions appear in the upper level (150-200-hPa) over the Indian Ocean from 45°E to 90°E, the eastern Pacific from 135°W to 75°W, and the Atlantic Ocean from 30°W to 15°E (Fig. 4.4a), and correspond to the PKE peaks in these regions (Fig. 4.2a). These large PKE productions normally locate at the downstream of upper-tropospheric zonal wind center (Fig. 4.1a) where the zonal wind shear
helps the accumulation of energy. Negative barotropic conversions are present in the upper troposphere over the western Pacific (125°E-180°) and South America (~60°W) in NCEP. The observed distribution of barotropic conversions is generally simulated by ISUCCM3 and CTL but with the difference in the amplitude (Figs. 4.4b and 4.4c). Comparing with the NCEP, ISUCCM3 has smaller positive barotropic conversions over the west coast of South America and the Atlantic Ocean, but larger positive barotropic conversions over the maritime continent between 90°E and 120°E. This increased upper-tropospheric PKE from CTL to ISUCCM3 over the Indian Ocean (Fig. 4.5a) is mainly due to the impact of revised convection closure and trigger condition as shown in the differences of barotropic conversion (Figs. 4.5d and 4.5c), and the impact of the revised convection closure plays a more important role around 90°E. In Eq. (4.1), the barotropic conversion contains four terms:
\[
-u' \frac{\partial \bar{u}}{\partial x}, \quad -u' \frac{\partial \bar{v}}{\partial y}, \quad -u' \frac{\partial v}{\partial x}, \quad \text{and} \quad -v' \frac{\partial v}{\partial y}.
\]
The zonal wind related term \(-u' \frac{\partial \bar{u}}{\partial x}\) is the dominant term among them. Figure 4.6 examines the difference of \(-u' \frac{\partial \bar{u}}{\partial x}\) between NOTRI and CTL along equator in the 20-100-day period. It explains most features shown in Fig. 4.5d with the positive center in the upper troposphere over the Indian Ocean. With similar positive \(u' u'\) for NOTRI and CTL (not shown), the increases of mean easterly wind shear along the equator (\(\frac{\partial \bar{u}}{\partial x} < 0\)) due to the impact of revised closure is the major reason for the positive difference of barotropic conversion in the upper troposphere over the Indian Ocean. As shown in Fig. 4.7, the mean zonal wind at 200 hPa along the equator (10°N-10°S) illustrates that the mean easterly wind difference is 5 m s\(^{-1}\) in NOTRI between 45°E and 110°E, which is three times larger than the difference of 1.4 m s\(^{-1}\) in CTL. The negative
differences of barotropic conversion between NOTRI and CTL over the eastern Pacific and the Atlantic Ocean can be also explained by the impact of revised closure on the horizontal wind shear. Over these two regions, the zonal wind shear is reduced in NOTRI from CTL (Fig. 4.7). In addition, the large observed upper-tropospheric PKE productions over the Indian Ocean, eastern Pacific, and Atlantic Ocean through the barotropic conversion in Fig. 4.4a also collocate with the increases of mean easterly wind shear along the equator in Fig. 4.7.

Figure 4.8 shows the longitude-height cross section of the partial covariance of specific volume ($\alpha$) and vertical motion ($-\omega$) in the 20-100-day period with the vertical and zonal wave energy flux superimposed. In the analysis of NCEP datasets (Fig. 4.8a), the large band of energy conversion from PAPE to PKE occurs around 200-400 hPa over the Indian Ocean-western Pacific where the active convection and strong upward wave energy flux present. The weak point of this large energy conversion band is around the maritime continent. One possible explanation is that the strong land heating (cooling) during the day (night) tends to favor the variability of convection on shorter timescales (e.g., diurnal cycle) than the 20-100-day period. ISUCCM3 also simulates a band of energy conversion over the Indian Ocean and western Pacific with the weak point over the maritime continent and the strong upward energy flux above the 300 hPa (Fig. 4.8b). However, CTL presents a weak energy conversion band with a gap over the Indian Ocean around 90°E, and the band extends to the eastern Pacific around 135°W where convection is normally suppressed. As shown by previous studies (e.g., Nitta 1970, 1972; Yanai et al. 2000; Mu and Zhang 2006), the conversion from PAPE to PKE through the covariance of ($\alpha$, $-\omega$) is largely supplied by the production of
PAPE through the covariance of \((\alpha, Q_i)\), i.e., 
\[
\frac{R}{c_p} \frac{\alpha}{Sp} \frac{\overline{Q_i'}}{\overline{Q_i}}
\]
(see Eq. 7 in Yanai et al. 2000),

where \(R\) is the specific gas constant, \(c_p\) the isobaric specific heat capacity, \(S = -\alpha \frac{\partial \ln \tilde{\theta}}{\partial p}\) the static stability factor, \(\tilde{\theta}\) the potential temperature, and \(Q_i\) the diabatic heating in which the deep convective heating is the dominant term. Figure 4.9 shows the covariance of specific volume and diabatic heating from observations and model simulations. In NCEP data, large PAPE generation due to the convective heating locates over the Indian Ocean-western Pacific with a small gap over the maritime continent (Fig. 4.9a), which explains most of PAPE conversion to PKE shown in Fig. 4.8a. ISUCCM3 presents a similar PAPE generation band centered over the convection active area around 200-400hPa, but with a larger amplitude than NCEP (Fig. 4.9b). The PAPE generation band around 300 hPa in CTL is weaker than both NCEP and ISUCCM3, but extends to the eastern Pacific (Fig. 4.9c). These results confirm the role of coupling between convection and large-scale circulation in the maintenance of MJO PKE over the Indian Ocean and western Pacific obtained by previous studies. The increase of PAPE production from CTL to ISUCCM3 is mainly over the western Pacific and the Indian Ocean as shown in the difference of covariance \((\alpha, Q_i)\) between ISUCCM3 and CTL (Fig. 4.10a). The CMT, through the influence on the convective heating, strengthens the positive correlation between the specific volume and the convective heating from 140°E to 170°E but weakens the PAPE production near 130°E (Fig. 4.10b). The convection trigger condition contributes to the increase of PAPE production around 300 hPa near 130°E and 90°E (Fig. 4.10c).
Figure 4.11 presents the longitude-height cross section of convergence of horizontal wave energy flux along the equator in the 20-100-day period. NCEP observations indicate two major convergent zones in the upper troposphere over the central-eastern Pacific and South America, and two major divergent zones over the west coast of South America and the Atlantic Ocean (Fig. 4.11a). These features are generally simulated in ISUCCM3, but with stronger convergences over the central-eastern Pacific and the South America and weaker divergence over the west coast of South America (Fig. 4.11b). The convergence over the central-eastern Pacific in ISUCCM3 is much stronger than that in CTL (Fig. 4.11c). Further analysis of 200-hPa horizontal wave energy fluxes in the 20-100-day period from observations and model simulations (Fig. 4.12) reveals that the convergence over the central-eastern Pacific between 150°W-120°W is largely due to the meridional wave energy fluxes from three source regions, i.e., the subtropical latitudes of both hemispheres and the west coast of North America, which confirms the finding of Yanai et al. (2000). The meridional wave energy fluxes from the subtropical latitudes of both hemispheres also contribute to the convergence over the east coast of South America. The enhanced horizontal convergence over the central-eastern Pacific from CTL to ISUCCM3 is largely due to the impact of the convection trigger condition as highlighted in the difference of horizontal wave energy fluxes between NOCMT and NOTRI (Fig. 4.13). The enhanced southward (northward) wave energy fluxes to the equator from the subtropical latitudes of north (south) hemisphere are marked by a circle near 140°W. It is noticed that the westward wave energy flux over the Indian Ocean in CTL (Fig. 4.12c) is reversed to the eastward wave energy flux in ISUCCM3 (Fig. 4.12b) which is also shown in NCEP analysis (Fig. 4.12a). The impact of the revised convection closure plays a role in this reversal as shown in the difference of horizontal wave
energy fluxes between NOTRI and CTL (Fig. 4.14). The enhanced eastward energy flux exists from the Indian Ocean to the central Pacific along equator, which contributes to the convergence of horizontal wave energy flux over the central Pacific in the upper troposphere and supports the eastward MJO signal simulated by NOTRI compared to the westward signal in CTL (Fig. 4 in Deng and Wu 2010).

The convergence of vertical wave energy flux along the equator in the 20-100-day period is given in Fig. 4.15. NCEP analysis shows the convergence above 200 hPa and the divergence below over the Indian Ocean and western Pacific (Fig. 4.15a), which is the result of upward and downward wave energy flux from the peak of PAPE production near 250 hPa (Figs. 4.8a and 4.9a). The convergence of vertical wave energy flux apparently is a major contributor to the maintenance of upper-tropospheric PKE in these two regions (Fig. 4.2a). ISUCCM3 simulates the similar convergence/divergence pattern but with larger amplitude over the western Pacific and Indian Ocean compared to NCEP (Fig. 4.15b). This is due to the larger upward and downward wave energy fluxes from the peak of stronger PAPE production (Figs. 4.8b and 4.9b). In comparison with CTL (Fig. 4.15c), the major impacts of the modified convection scheme are over the western Pacific where the convergence of vertical wave energy fluxes is enhanced above 200 hPa from CTL to ISUCCM3, and over the central-eastern Pacific where the divergence of vertical wave energy fluxes is increased above 300 hPa due to the enhanced downward energy fluxes between 150°W and 120°W (Figs. 4.8b and 4.8c). Further comparisons among the four simulations indicate that the convection trigger condition is the major contributor to the enhanced upper-tropospheric convergence of energy fluxes over the western Pacific and Indian Ocean.
Finally, the ISUCCM3 simulation which includes the parameterization of convective momentum transport allows the analysis of work done by the convection-induced momentum tendency and its influence on the PKE budget. Figure 4.16 shows predominantly negative work above 400 hPa with the peak near 200 hPa corresponding to the large zonal wind velocity. It indicates that the impact of convective momentum transport reduces the mean upper-tropospheric PKE but with relatively small amplitude comparing to other terms in the budget equation. The dominant negative work by the convective momentum tendency is consistent with the estimate of kinetic energy transfer rate due to the momentum budget residual during the TOGA COARE (Tung and Yanai 2002). Further analysis shows that the work done by the momentum tendency mostly comes from the component of $\overline{u'F'_u}$, and the negative $\overline{u'F'_u}$ indicates the tendency to decelerate the zonal wind. It is noted that in the lower troposphere over the western Pacific between 150°W and 180°, a dipole pattern appears with the positive work below 850 hPa and negative above. The positive work indicates the acceleration of low-level westerly wind (Fig. 4.17) by the convective momentum transport which is favorable for the development and eastward propagation of MJO.

5. Summary

The Indian Ocean with the active convection is the source region of MJO. The upper-tropospheric MJO PKE maintenance not only depends on the convergence of the vertical wave energy flux from the convective heating source region where large PAPE production exits, but also the barotropic conversion term through the horizontal shear of mean flow. The
impact of revised closure plays an important role for the enhancement of barotropic conversion through the work done by the mean zonal wind shear, and in turn favors the upper-tropospheric MJO PKE production. The convection trigger condition, through its requirement that convection is only activated when the CAPE continuously increases and exceeds certain threshold, generates stronger convective heating which enhances the upward wave energy fluxes and leads to the increase of MJO PKE over the Indian Ocean.

Over the western Pacific, the upper-tropospheric MJO PKE is supported by the convergence of horizontal wave energy flux as well as the convergence of vertical wave energy flux from the convective heating source region around 200-400 hPa. The work done by the barotropic conversion through the horizontal shear of the mean flow appears to decrease the upper-tropospheric MJO PKE. As over the Indian Ocean, the convection trigger condition helps build up more robust deep convection and stronger upward wave energy flux, and enhances the MJO PKE production in the upper troposphere over the western Pacific. The inclusion of convective momentum transport in the convection scheme reduces the upper-tropospheric MJO PKE through the work done by the convection-induced momentum tendency with the deceleration tendency of the upper-tropospheric wind. But the convective momentum transport tends to accelerate the westerly wind below 850 hPa over the western Pacific and enhances the MJO PKE.

Over the central-eastern Pacific, although the convection is suppressed, the large-scale circulation carries on the MJO signal. The suppressed deep convection coupled with the downward wave energy flow appears to decrease the MJO PKE in the upper troposphere.
The MJO PKE production is mainly due to the work done by the horizontal wave energy flux. The impact of convection trigger condition leads to the enhanced equatorward wave energy fluxes from the subtropical latitudes of both hemispheres, and contributes to the convergence of horizontal wave energy flux. The equatorial eastward wave energy flux from the Indian Ocean to the central Pacific caused by the revised closure also results in the convergence over the central-eastern Pacific, and favors the eastward propagation of MJO.
Fig. 4.1. Ten-year (1979-88 October to April) mean vertical cross section of zonal (m s\(^{-1}\)) and vertical (hPa day\(^{-1}\)) wind averaged between 10°N and 10°S for (a) NCEP, (b) ISUCCM3 and (c) ISUCCM3-CTL. The color contour is for zonal wind.
Fig. 4.2. Longitude-height section of mean PKE in the 20-100-day period along the equator (10°N-10°S) for (a) NCEP, (b) ISUCCM3 and (C) CTL during ten-year (1979-88 October to April). The contour interval 2.5 J kg⁻¹; the areas with values larger than 12.5 J kg⁻¹ are shaded.
Fig. 4.3. As Fig. 4.2 but for (a) ISUCCM3-CTL, (b) ISUCCM3-NOCMT, (c) NOCMT-NOTRI, and (d) NOTRI-CT, and the areas with values larger than 5 J kg$^{-1}$ are shaded.
Fig. 4.4. Longitude-height section of the barotropic conversion term in the 20-100-day period along the equator (10°N-10°S) for (a) NCEP, (b) ISUCCM3 and (c) CTL during ten-year (1979-88 October to April). Negative areas are shaded. The contour interval is 2 J kg⁻¹ day⁻¹.
Fig. 4.5. As Fig. 4.4 but for (a) ISUCCM3-CTL, (b) ISUCCM3-NOCMT, (c) NOCMT-NOTRI, and (d) NOTRI-CTL.
**Fig. 4.6.** Longitude-height section of $-\bar{u} \bar{u} \frac{\partial \bar{u}}{\partial x}$ in the 20-100-day period along the equator (10°N-10°S) for NOTRI-CTL during ten-year (1979-88 October to April). Negative areas are shaded. The contour interval is 2 J kg\(^{-1}\) day\(^{-1}\).

**Fig. 4.7.** Mean zonal wind (m s\(^{-1}\)) along the equator (10°N-10°S) at 200 hPa for NCEP, NOTRI and CTL during ten-year (1979-88 October to April).
Fig. 4.8. Longitude-height section of the wave energy flux \((F_x,F_z)\) in the 20-100-day period along the equator (10°N-10°S) for (a) NCEP, (b) ISUCCM3, and (c) CTL during ten-year (1979-88 October to April). Areas of positive \((\alpha,-\omega)\) covariance are shaded and its contour interval is 2 J kg\(^{-1}\) day\(^{-1}\). \(F_z\) is the hydrostatic translation of \(F_p\). The unit for \(F_x\) is 600 J m s\(^{-1}\) kg\(^{-1}\), and for \(F_z\) is 0.35 J m s\(^{-1}\) kg\(^{-1}\).
Fig. 4.9. Longitude-height section of the ($\alpha, Q_l$) covariance in the 20-100-day period along the equator ($10^\circ$N-$10^\circ$S) for (a) NCEP, (b) ISUCCM3, and (c) CTL during ten-year (1979-88 October to April). Areas of negative ($\alpha, Q_l$) covariance are shaded. The contour interval is 2 J kg$^{-1}$ day$^{-1}$. 
Fig. 4.10. As Fig. 4.9 but for (a) ISUCCM3-CTL, (b) ISUCCM3-NOCMT, and (c) NOCMT-NOTRI.
Fig. 4.11. Longitude-height section of horizontal convergence of $F_h$ in the 20-100-day period along the equator ($10^\circ$N-$10^\circ$S) for (a) NCEP, (b) ISUCCM3, and (c) CTL during ten-year (1979-88 October to April). Negative areas are shaded. The contour interval is $2 \text{ J kg}^{-1} \text{ day}^{-1}$. 
Fig. 4.12. Horizontal map of wave energy flux ($F_h$) in the 20-100-day period range for (a) NCEP, (b) ISUCCM3, and (c) CTL during ten-year (1979-88 October to April) at 200 hPa. Magnitudes of flux (J m$^{-1}$ kg$^{-1}$) are shaded. The circles mark the areas with convergence of horizontal wave energy flux along equator. The arrows present the wave energy flux from both hemispheres. The rectangle shows the Indian Ocean horizontal wave energy flux along equator.
Fig. 4.13. As Fig. 4.12 but for NOCMT-NOTRI, and the circle marks the area with convergence of meridional wave energy flux.

Fig. 4.14. As Fig. 4.12 but for NOTRI-CTL, and the circle marks the area with eastward wave energy flux.
Fig. 4.15. Longitude-height section of vertical convergence of $F_p$ in the 20-100-day period along the equator (10°N-10°S) for (a) NCEP, (b) ISUCCM3, and (c) CTL during ten-year (1979-88 October to April). Negative areas are shaded. The contour interval is 2 J kg$^{-1}$ day$^{-1}$. 
Fig. 4.16. Longitude-height section of the work done by the convection-induced momentum tendency ($\mathbf{v}' \cdot \mathbf{F}'$) for ISUCCM3 during ten-year (1979-88 October to April) in period range of 20-100 days along the equator (10°N-10°S). The contour interval is 0.5 J kg$^{-1}$ day$^{-1}$.

Fig. 4.17. Zonal wind (m s$^{-1}$) in the 20-100-day period range for ISUCCM3 along equator (10°N-10°S) during ten-year (1979-88 October to April). The contour interval is 0.01 m s$^{-1}$. 
CHAPTER 5. ANALYSIS OF MOIST STATIC ENERGY BUDGET FOR THE
MADDEN-JULIAN OSCILLATION OVER THE INDIAN OCEAN AND PACIFIC

5.1 Introduction

MJO is an atmospheric response to forcing sources that has been proposed by many scientists. The forcing source in this type of theory includes the tropical localized thermal forcing and stochastic forcing. The tropical response to the stationary intraseasonal oscillating heat source or the randomly varying heating profiles, such as the convective activity associated with the Asian monsoon and the convective disturbance within the ITCZ, produces some observed features of the MJO (e.g., Yamagata and Hayashi 1984; Anderson and Stevens 1987; Salby and Garcia 1987). However, the theory does not provide the mechanism responsible for the origin of low oscillation frequency and eastward movement of the heating source. Hu and Randall (1994, 1995) suggested that the low frequency and localized convective heat source is due to the nonlinear interactions among radiation, cumulus convection, and surface moisture fluxes. Blade and Hartmann (1993) introduced a discharge-recharge mechanism to determine the period of the low frequency oscillation of convective heating by the discharge time of convective stabilization together with the recharge time of moist static instability. Maloney and Hartmann (1998) also suggested that the drying of the atmosphere occurs rapidly after the passage of convection with the onset of 850-mb westerly perturbations, and the moistening process in front of convection may set the timescale for the reinitiating of convection over the Indian Ocean and western Pacific. The discharge-recharge mechanism was further supported by Kemball-Cook and Weare (2001) with observational study over a radiosonde station. MJO events appear to begin with the
destabilized atmosphere through a combination of low-level moist static energy buildup which is controlled by a corresponding increase in low-level moisture.

Many scientists examined the MJO related moisture and convection process through the composite analysis. Myers and Waliser (2003) presented the composite analysis of the three-dimensional structure and evolution of moisture associated with the MJO events which were selected using the Xie–Arkin bandpassed pentad rainfall data. During the MJO life cycle, the low tropospheric water vapor leads precipitation over the Indian Ocean and western Pacific, and the upper-tropospheric water vapor lags the precipitation peak for its moistening following intense convection. Tian et al. (2006) examined the Atmospheric Infrared Sound (AIRS) data through the composite analysis based on the Tropical Rainfall Measuring Mission (TRMM) data, and showed that the MJO convection is preceded by a low tropospheric moist anomaly and followed by a low tropospheric dry anomaly. With the MJO deep convection events selected from the spectrally filtered TRMM and Global Precipitation Climatology Project (GPCP) data, Benedict and Randall (2007) suggested that the moist precondition of the MJO deep convection includes the vertical transport of moisture related to the shallow convection, while the dry process behind the MJO deep convection maybe related to the mesoscale process. The discharge-recharge mechanism is relevant for explaining many observed MJO feature identified in the composite analysis. Recently, Maloney (2009) studied the moist static energy (MSE) budget through the composite analysis of the MJO events selected using the 850-hPa zonal wind, and suggested that the MSE recharge process occurs in advance of the MJO precipitation with the low tropospheric easterlies and the MSE discharge process is during and after precipitation with the westerly
anomalies. The horizontal advection of MSE and the surface latent heat flux are the leading terms for the MSE budget.

The objective of this chapter is to understand the mechanisms and physical processes affecting the MJO simulation through the study of moist static energy (MSE) budget. The MJO events are selected based on the analysis of 850-hPa zonal wind. The phase relationships between the 850-hPa zonal wind, precipitation and surface latent heat flux will be analyzed for the composite MJOs over the Indian Ocean and western Pacific. The impacts of revised convection closure, convection trigger and CMT on the MSE budgets of composite MJO will be examined in these two regions.

5.2 MJO simulated by ISUGCM

Similar to Slingo et al. (1999), the MJO activity is defined as the variance of the zonal mean 200-hPa daily zonal wind averaged between 5°N and 5°S after applying a 30-90-day band-pass filter. The time series of the variance is plotted in a 100-day moving window. The observed variance reaches 10 m² s⁻² in some years, and shows strong peaks in 1979/1980, 1981, 1985, 1985/1986, and 1987/1988 (Fig. 5.1a). The amplitude of variance in ISUCCM3 is similar to the observations, and the observed strong peak in 1981 is well simulated in ISUCCM3 (Fig. 5.1b). The amplitude of the CTL variance is weak (5 m² s⁻²), and the peaks are not strong (Fig. 5.1c). The weaker MJO-related variance of the 200-hPa wind in CTL suggests that the MJO simulation in ISUCCM3 is in better agreement with the observations.
Figure 5.2 shows the lag correlations of daily values of 850-hPa zonal wind and precipitation onto daily 850-hPa zonal wind time series at 90°E and 155°E, all time series (1979-1988 October-April) averaged between 5°N and 5°S and filtered to retain the variability at periods of 30-90 days. NCEP zonal wind anomaly shows coherent eastward propagation across the Indian Ocean-Pacific, and the speed is about 5 m s⁻¹ over the Indian Ocean, 10 m s⁻¹ over the Pacific (Figs. 5.2a-b). The precipitation anomalous positive peak shows earlier than the wind anomalous positive peak around 90°E in the Indian Ocean (Fig. 5.2a), and is in phase with the wind anomalous positive peak around 155°E in the western Pacific (Fig. 5.2b). ISUCCM3 exhibits similar coherent eastward propagation as the observations, but the amplitude of intraseasonal variability is larger than that in the observations for both zonal wind and precipitation over the western Pacific (Figs. 5.2c-d). In addition, the phase relationships between the precipitation and zonal wind anomalous positive peaks are similar to those in NCEP with respect to the zonal wind. However, CTL zonal wind does not capture the coherent eastward propagation; it shows westward propagation over both the Indian Ocean and western Pacific coupled with the intraseasonal precipitation variability (Figs. 5.2e-f).

5.3 Analyses of the composite MJO

a. Composite MJO

To further illustrate the mechanisms and physical processes controlling the evolution and development of MJO events, the composite MJO is obtained using observations and four ISUGCM simulations. Following the method of Maloney (2009), the MJO events are identified using the 30-90-day band-passed 850-hPa zonal wind during 1979-1988 averaged
from 5°N to 5°S at 90°E (the Indian Ocean) and 155°E (the western Pacific). The selected events have positive maxima larger than one standard deviation of 850-hPa zonal wind time series. An example is shown in Fig. 5.3 for ISUCCM3. There are 24 events over the Indian Ocean and 27 events over the western Pacific in the period of October to April. Using the same method, 25 (30), 27 (23), 19 (25) and 22 (26) events are obtained for NCEP, CTL, NOCMT and NOTRI over the Indian Ocean (the western Pacific), respectively.

Figure 5.4 presents composite intraseasonal 850-hPa zonal wind, precipitation and latent heat flux at 90°E and 155°E as a function of lag in days relative to the zonal wind events. In NCEP, the precipitation peak is about 1 mm day\(^{-1}\) at 90°E, and leads the 850-hPa zonal wind anomalous peak about 7 days (Fig. 5.4a). However, the precipitation peak at 155°E is in phase with the zonal wind maximum (Fig. 5.4b), which is consistent with the phase relationship of precipitation and wind anomalous peak in Fig. 5.2b. These results confirm previous studies that show the MJO convection and precipitation center more often locating between the surface westerlies and easterlies over the Indian Ocean, but collocating with the prevailing westerlies and maximum surface latent and sensible heat fluxes over the western Pacific (e.g., Madden and Julian 1972; Zhang and Anderson 2003).

CTL represents the observed phase relationship over the Indian Ocean, but the amplitudes of the latent heat flux and precipitation are smaller than those of ISUCCM3 and NCEP (Figs. 5.4a, 5.4c and 5.4e). The impact of convection trigger condition plays a major role for this amplitude enhancement (Figs. 5.4g and 5.4i). Over the western Pacific, the observed coherent structure of the latent heat flux, surface westerly, convection and precipitation are
better simulated in ISUCCM3 compared to CTL (Figs. 5.4b, 5.4d and 5.4f), so do the amplitudes of the latent heat flux and precipitation. The enhancement of amplitude is again mainly due to the inclusion of convection trigger condition (Figs. 5.4h and 5.4j). The maximum precipitation anomalous peak shifts from Day -12 to Day -4 (from Day -14 to Day -7) over the western Pacific (the Indian Ocean) in Figs. 5.4h and 5.4d (Figs. 5.4g and 5.4c) due to the impact of CMT. Consequently, the precipitation anomalous peak is more in phase with the 850-hPa zonal wind anomalous peak in ISUCCM3 compared to NOCMT. The inclusion of the revised closure enhances the amplitudes of precipitation and latent heat flux over the Indian Ocean and western Pacific (Figs. 5.4e-f and 5.4i-j).

b. Moist static energy

The moist static energy (MSE) is defined as $m = c_p T + gz + L q$, where $c_p$ is the specific heat at constant pressure, $T$ the temperature, $g$ the gravitational acceleration, $z$ the height, $L$ the latent heat, and $q$ the specific humidity. Since the contribution from the temperature anomalies ($c_p T$) and geopotential height anomalies ($gz$) is small, only the composite intraseasonal moist static energy anomalies and moisture anomalies ($L q$) between 1000 and 300 hPa will be shown in the following analysis. In Fig. 5.5, the MSE and $L q$ are averaged between 5°N and 5°S as a function of lag in days with respect to the zonal wind events. Beside the NCEP reanalysis, the 7-year (2002-2008) ERAI and AIRS datasets are used for comparison with model simulations. 19 (18) MJO events are selected for ERAI during October to April at 90°E (155°E), and 17 (15) MJO events for AIRS at 90°E (155°E). Over the Indian Ocean, the ERAI and AIRS MSE and $L q$ positive anomalies appear near surface around Day -20 and develop upward with the peak near 500 hPa around Day -8 (Figs. 5.5a
and 5.5b), which indicates the moisture transfer from the surface to the upper troposphere.

The positive MSE anomalous peak (or moisture peak) shows up earlier than the wind anomalous peak and corresponds to the leading observational precipitation peak in Fig. 5.4a. The negative MSE anomalies (or dry processes) display during and after the 850-hPa zonal wind peak and reach the maximum around Day 15 which represents the discharge process of MJO. This MSE recharge-discharge process undergoes gradual moistening and drying for about 50 days over the Indian Ocean (Tian et al. 2006). The MSE recharge-discharge processes over the western pacific are similar to those over the Indian Ocean, but the amplitude is slightly stronger and the maximum positive peak is closer to the 850-hPa zonal wind peak. The MSE and Lq of NCEP also show the moistening and drying processes associated with the evolution of MJO, but are more vertically stacked (Fig. 5.5c). There is no obvious low-level moist preconditioning in NCEP compared to ERAI and AIRS, especially over the Indian Ocean. Since the observational data in this region is sparse, the NCEP reanalysis as well as ERA reanalysis may largely rely on the model physical processes. The uncertainties in the representation of these processes could explain the difference between NCEP and AIRS. Note that ERAI has assimilated more observational data than the original ERA reanalysis.

The observed recharge-discharge process of MSE is simulated by the ISUCCM3 (Fig. 5.5d) with a 50-day period at 90°E and 155°E, but both the positive and negative MSE anomalies start earlier in the lower troposphere which suggests an early precondition of the MJO convection. In addition, the amplitude of MSE maximum is generally weaker than the observations over the western Pacific, which is mainly due to the weak moisture anomalous
center. The observed precondition of MJO convection is not simulated well in CTL (Fig. 5.5e), especially for the western Pacific where the moisture anomaly appears in the upper troposphere not the surface and the MSE positive anomalous peak is later than the wind anomalous peak. The impact of revised closure reshapes the moisture distribution and helps build up the precondition of the MJO convection with the comparisons between NOTRI and CTL (Figs. 5.5g and 5.5e). The initial signal of moisture now begins from the surface presenting the lower troposphere moisture convergence in front of the MJO convection over the western Pacific in NOTRI. The enhancement of MSE recharge-discharge process is mainly contributed by the convection trigger condition over the Indian Ocean and western Pacific through the moisture anomalous process (Figs. 5.5f-g). The possible explanation is that the more robust deep convection due to the convection trigger condition enhances the moisture convergence and transfer from the boundary lay to the upper troposphere. The impact of CMT through the moisture distribution results in generally higher and later MSE anomalous peak and favors the more coherent structure between the surface westerly wind center and the convection and precipitation center over the Indian Ocean and western Pacific in ISUCCM3 compared to NOCMT (Figs. 5.5d and 5.5f). This result is in agreement with the more coherent structure of MJO-related wind and precipitation maxima due to the CMT in Figs. 5.4c-d and 5.4g-h.

c. Moist static energy budget

Following Neelin and Held (1987) and Maloney (2009), the vertically integrated MSE budget after 30-90-day band-pass can be written as:
\[
\frac{\partial \mathbf{m}}{\partial t} = -\left( \omega \frac{\partial \mathbf{m}}{\partial \mathbf{p}} \right) - \langle \mathbf{V} \cdot \nabla \mathbf{m} \rangle + LH + SH + \langle LW \rangle + \langle SW \rangle.
\]  

Term A presents the MSE tendency, term B the vertical advection of MSE, term C the horizontal advection of MSE, term D the latent heat flux, term E the sensible heat flux, term F the longwave heating rate, and term G the shortwave heating rate. Also, \( \langle \cdot \rangle \) represents the mass-weighted vertical integral from 1000 hPa to 100 hPa. Composite intraseasonal vertically-integrated (1000-100 hPa) moist static energy budget anomalies are shown in Fig. 5.6. The sum of latent and sensible heat fluxes is presented as the latent heat flux is dominant. The shortwave heating rate is small and not shown in the plot. The NCEP moisture used in the plot is from 1000 hPa to 300 hPa for the data availability, which will affect the horizontal and vertical advections of MSE, also the budget balance. The charging of MSE in NCEP appears before the latent heat flux (Figs. 5.6a-b), precipitation (Figs. 5.4a-b), and long wave heating rate anomalous peak, and the discharging process is during and after (e.g., Maloney 2009). The MSE tendency is contributed mainly by the horizontal and vertical advection, but horizontal advection is more in phase with the MSE tendency over the Indian Ocean and the vertical advection is more important for the MSE recharge process over the western Pacific.

The recharge process of MSE in ISUCCM3 (Figs. 6c-d) is similar to the observations with the horizontal and vertical advections contributing positively to the MSE, except the horizontal advection over the Indian Ocean. The increase of MSE is in phase with the horizontal advection of MSE over the western Pacific (Fig. 5.6d), but in phase with the
vertical advection of MSE over the Indian Ocean (Fig. 5.6c). CTL shows weak MJO signal (Figs. 5.6e-f), and both the horizontal and vertical advections work negatively during the MSE recharge process, except the horizontal advection over the Indian Ocean. The impact of three modifications enhances the MJO signal and reverses the contributions of the horizontal and vertical advections to the MSE recharge process over the Indian Ocean and western Pacific.

The inclusion of revised closure leads to positive contributions of the horizontal and vertical advections to the MSE tendency (Figs. 5.6i-j and 5.6e-f) before the MJO deep convection (Figs. 5.4i-j), but the amplitude of MJO signal does not change much. In fact, the amplitude enhancement of MJO is due to the use of convection trigger condition which leads to more robust deep convection (Figs. 5.6g-h). The vertical advection plays a more important role in the MSE recharge process than the horizontal advection over the Indian Ocean when the trigger condition is added in the scheme (Figs. 5.6g and 5.6i). The impact of CMT over the western Pacific (Figs. 5.6d and 5.6h) during the MSE recharge process leads to the enhanced vertical advection positive peak (from 6 W m\(^{-2}\) to 21 W m\(^{-2}\)) and later horizontal advection positive peak (from Day -18 to Day -11), which interact with each other illustrating a later recharge process (from around Day -30 to around Day -18). Corresponding to the shift of the precipitation and MSE anomalous maximum from around Day -12 to around Day -4 (Figs. 5.4h, 5.4d, 5.5f and 5.5d), the later MSE recharge process favors the more coherent structure of the MJO convective center and the prevailing westerly anomalies over the western Pacific in ISUCCM3. Over the Indian Ocean, a later MSE recharge process presents due to the CMT impact (Figs. 5.6c and 5.6g), which is also helpful to build up a more coherent structure.
5.4 Summary

The MJO variability and coherent eastward propagation across the Indian Ocean-Pacific in ISUGCM is in better agreement with the observation due to the modified convection scheme. The amplitude of the MJO-related variance for the 200-hPa wind in ISUGCM with three modifications (revised closure, trigger and CMT) to the deep convection scheme is stronger than that in control run with the original convection scheme and is comparable to the observations. Also, the observed coherent eastward propagating structure that the convection and precipitation center leads the surface westerlies anomalous peak over the Indian Ocean and collocates with the prevailing westerlies over the western Pacific, are simulated better with the modified scheme than the original scheme.

With the composite analyses, the lag (in phase) relationships between MJO 850-hPa zonal wind, precipitation and surface latent heat flux are simulated over the Indian Ocean (western Pacific) in ISUGCM, which are greatly influenced by the convection closure, trigger and CMT. The moist static energy builds up from the lower troposphere 15 to 20 days before the peak of MJO precipitation, and reaches the maximum in the middle troposphere near the peak of MJO precipitation. The gradual lower-tropospheric heating and moistening and the upward moist static energy transport are important aspects of MJO events which are documented in observational studies but poorly simulated in most GCMs. With the modifications in the convection scheme, ISUGCM produces better MJO recharge-discharge process of moist static energy than the original GCM. The trigger condition for deep convection contributes to the striking difference between ISUGCM and the control run with the original convection scheme plays a major role for this improved MJO simulation through
the horizontal and vertical advections of moist static energy. The inclusion of the revised closure helps precondition the MJO convection with the redistribution of moisture through the positive contribution of the horizontal and vertical advection of moist static energy before the onset of MJO convection. The impact of CMT through the interaction between horizontal and vertical advection of moist static energy helps the more coherent atmospheric structure over the Indian Ocean, and favors the collocation of the MJO convective center and the prevailing westerly anomalies over the western Pacific. In addition, the budget analysis for ISUGCM with the modifications shows the increase of moist static energy is in phase with the horizontal advection of moist static energy over the western Pacific, but in phase with the vertical advection of moist static energy over the Indian Ocean.
Fig. 5.1. Interannual variability of MJO depicted by 1979-88 variance time series of 30-90-day band-pass 200-hPa zonal wind which averaged between 5°N and 5°S for (a) NCEP, (b) ISUCCM3, and (c) CTL. A 100-day running mean was used on the variance time series. Unit is m² s⁻².
Fig. 5.2. Ten-years (1979-88 October-April) lag correlations of 30-90-day band-passed daily 850-hPa zonal wind (contours) and precipitation (colors) onto the daily 850-hPa zonal wind time series at 90°E for (a) NCEP, (c) ISUCCM3, and (e) CTL, at 155°E for (b) NCEP, (d) ISUCCM3, and (f) CTL. Data are averaged between 5°N and 5°S. Correlation coefficients are scaled by 1 σ value of the reference time series. Contours are plotted every 0.3 m s⁻¹, starting at 0.15 m s⁻¹. Negative contours are dashed. The zero contour is not shown. Precipitation anomaly units are mm day⁻¹. Yellow solid line represents a phase speed 5 m s⁻¹.
Fig. 5.3. The variability of 850-hPa zonal wind for ISUCCM3 at 90°E a) and 155°E b) during 1979-88 after 30-90-day band-pass filtering averaged between 5°N and 5°S. Unit is m s⁻¹. One reference line represents zero, the other reference line represents positive one standard deviation of the 850-hPa zonal wind time series after 30-90-day band-pass filtering.
Fig. 5.4. Composite intraseasonal 850-hPa zonal wind (solid; m s$^{-1}$), precipitation (dot; mm day$^{-1}$) and latent heat flux (dashed; 10 W m$^{-2}$) for NCEP at a) 90°E and b) 155°E averaged between 5°N and 5°S as a function of lag in days relative to the zonal wind events. c-d, e-f, g-h and i-j as a-b but for ISUCCM3, CTL, NOCMT and NOTRI, respectively.
Fig. 5.5. Composite intraseasonal moist static energy (top) and moisture (bottom) anomalies for a) ERAI at 90°E (left) and 155°E (right) averaged between 5°N and 5°S as a function of lag in days relative to the zonal wind events. b, c, d, e, f and g as a but for AIRS, NCEP, ISUCCM3, CTL, NOCMT and NOTRI, respectively. The contour level is 100 J kg$^{-1}$, starting at 50 J kg$^{-1}$. Areas greater (less) than 50 (-50) J kg$^{-1}$ are dark (light) shaded.
Fig. 5.6. Composite intraseasonal vertically-integrated (1000-100 hPa) moist static energy budget anomalies for NCEP at a) 90°E and b) 155°E averaged between 5°N and 5°S as a function of lag in days relative to the zonal wind events. c-d, e-f, g-h and i-j as a-b but for ISUCCM3, CTL, NOCMT and NOTRI, respectively. The unit is W m⁻². The gray, green, red, blue and orange lines are for long wave heating, latent and sensible heat flux, horizontal advection, vertical advection and moist static energy tendency, respectively.
CHAPTER 6. GENERAL CONCLUSIONS AND FUTURE WORK

6.1 General Conclusions

The MJO signal simulated by ISUGCM is improved with the use of revised closure assumption, convection trigger condition, and CMT in the convection scheme. The improvements include the amplitude, spatial distribution, eastward propagation, and horizontal and vertical structure of MJO, especially for the coherent structure and precondition of the eastward-propagating MJO convection. The inclusion of revised closure in the convection scheme favors the eastward propagation of MJO. The convection trigger condition enhances the MJO convection activity and results in a strong MJO signal. The impact of CMT leads to more coherent structure of the MJO convection and its corresponding variance.

To illustrate the physical processes responsible for the improved MJO simulations, the tropical kinetic energy budget is analyzed. The equatorial (10°S-10°N) MJO-related perturbation kinetic energy in the upper troposphere is enhanced due to the modified deep convection scheme, and this enhancement results in more robust and coherent eastward propagating MJO signal. In the MJO source region—the Indian Ocean, the upper-tropospheric MJO PKE is maintained by the convergence of vertical wave energy flux and the barotropic conversion through the horizontal shear of mean flow. Over the western Pacific where convection is active, the upper-tropospheric MJO PKE is supported by the convergence of horizontal and vertical wave energy fluxes. Over the central-eastern Pacific where the
convection is suppressed, the convergence of horizontal wave energy flux is the main source for the upper-tropospheric MJO PKE.

The inclusion of the revised convection closure affects the response of mean zonal wind shear to the convective heating over the Indian Ocean, and results in the increased upper-tropospheric MJO PKE through the barotropic conversion. The stronger eastward wave energy flux due to the increase of convective heating over the Indian Ocean and western Pacific by the revised closure helps the eastward propagation of MJO and favors the convergence of horizontal wave energy flux over the central-eastern pacific. The stronger convective heating due to the deep convection trigger condition enhances the perturbation available potential energy (PAPE) production and the upward wave energy fluxes, and contributes to the enhanced MJO PKE over the Indian Ocean and western Pacific. The trigger condition also helps the MJO PKE production over the central-eastern Pacific through the increased convergence of meridional wave energy flux from the subtropical latitudes of both hemispheres. The convection-induced momentum tendency works negatively to the PKE budget in the upper troposphere through the deceleration of the upper-tropospheric wind. However, the convection momentum tendency accelerates the westerly wind below 850 hPa over the western Pacific, which is partially responsible for the improved MJO simulation.

With the composite analyses, the different phase relationships between MJO 850-hPa zonal wind, precipitation and surface latent heat flux are simulated over the Indian Ocean and western Pacific in ISUGCM, which are greatly influenced by the convection closure, trigger
and CMT. The moist static energy builds up from the lower troposphere around 15 to 20 days before the peak of MJO precipitation, and reaches the maximum in the middle troposphere near the peak of MJO precipitation. The gradual lower-tropospheric heating and moistening and the upward energy transport are important aspects of MJO events which are documented in observational studies but poorly simulated in most GCMs. With the modifications in the convection scheme, ISUGCM produces better MJO recharge-discharge process of moist static energy than the original GCM. The trigger condition for deep convection contributes to the striking difference between ISUCCM3 and CTL (control run with the original convection scheme) and plays the major role for this improved MJO simulation through the horizontal and vertical advections of moist static energy. The inclusion of the revised closure helps build up the precondition of the MJO convection with the redistribution of moisture through the positive contributions of the horizontal and vertical advection of moist static energy prior to the intense MJO convection. The impact of CMT helps the coherent atmospheric structure over the Indian Ocean, and favors the collocation of the MJO convective center and the prevailing westerly anomalies over the western Pacific. With the modifications in the convection scheme, the budget analysis for ISUGCM shows the increase of moist static energy is in phase with the horizontal advection of moist static energy over the western Pacific, but in phase with the vertical advection of moist static energy over the Indian Ocean.

6.2 Future Work

More work is recommended to further examine the impacts of the revised closure, convection trigger condition, and CMT in the convection scheme for ISUGCM. Above analysis is mainly based on the 10-year mean, so the evolution of MJO life cycle and its related super
cloud clusters and other associated perturbations would be useful to further investigate the MJO simulations in ISUGCM. Future analysis is recommended to illustrate the relationship between the wave energy flux and the convective heating through geopotential, especially for the zonal wave energy flux over the Indian Ocean and the meridional energy flux from extratropic to tropic over the central-eastern Pacific. Studying the low tropospheric process of convergence, moistening, and upward transport prior to the intense eastward propagating MJO convection from the Indian Ocean to the Pacific would be an interesting future work.

Deep convection is the major considered factor in this analysis for MJO, but the shallow convection in front of the MJO deep convection and its role in the precondition of the MJO eastward propagation still lacks comprehensive understanding. More diagnostic work using better collected observational data on the shallow cumulus will favor the mechanism study of MJO. The model simulation of the shallow convection and its interaction with the deep convection associated with the MJO life cycle is valuable to examine those theories proposed for MJO, and favors the MJO prediction. The MJO simulations discussed above are mainly over the Indian Ocean and Pacific, the maritime continent due to its land heating has strong influence on the eastward propagation of MJO. Further discussion about the influence of extratropical forcing to the MJO onset and eastward propagation through the horizontal wave energy flux is also useful to explore the MJO revolution.
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